Glacial North Atlantic Millennial Variability Over the Last 300,000 Years
by
Stephen Phillip Obrochta

Division of Earth and Ocean Sciences
Duke University

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Gary S. Dwyer

Dissertation submitted in partial fulfillment of
the requirements for the degree of Doctor
of Philosophy in the Division of
Earth and Ocean Science in the Graduate School
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2008
ABSTRACT

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Abstract

The hematite-stained grain (HSG) proxy method, commonly employed by the late G.C. Bond to detect the “1500-year cycle” in North Atlantic climate, is reproduced and verified for the first time. The exact method is compiled from various sources and presented in Chapter 1. In Chapter 2, an HSG record from classic North Atlantic DSDP Site 609 is reconsidered. While the Site 609 HSG record was initially interpreted to exhibit 1500-year variability, it did not actually contain spectral power at the 1500-year band. The chronology for Site 609 is based on radicarbon dates to 26 ka, beyond which the sea surface temperature record is matched to the record of air temperature variations over Greenland from the GISP2 ice core. However, it is now evident that the lack of spectral power at the primary period of the observed fluctuation was likely due to the GISP2 chronology, which has been subsequently shown to become progressively deficient over the course of the last glaciation. Updating the Site 609 chronology to the latest chronology for the virtually complete NGRIP Greenland ice core, which is based on layer counting to 60 ka, results in 99% significant spectral power at a 1/1415 year frequency.

In Chapter 3, the classic Site 609 lithic records are extended to the previous two glaciations, glacial Stages 6 and 8, at IODP Site U1308 (reoccupied Site 609). The “1500-year cycle” is not detected within Stage 6, perhaps indicating that D-O Events were not manifest in a similar fashion, if at all. Heinrich Event are also not detected, indicating relative stability of the North American Laurentide Ice Sheet during Stage 6. As a result, individual North Atlantic sites recorded lower-
amplitude, asynchronous hydrographic changes. The SST proxy record at Site U1308 during Stage 6 primarily records intermediate temperatures. The subtle SST changes detected likely indicate local as opposed to basin-scale changes related to the migration of oceanic frontal boundaries. During Stage 6, benthic δ\(^{13}\)C changes are of lower amplitude than Stages 2 - 4 and correspond more strongly to variations in SST than to ice rafting, indicating that ice-rafting events did not as strongly influence NADW formation.

During Stage 8, however, well-structured cycles in HSG with a mean event spacing of ~ 1500 ± 500 years are detected, potentially indicating a greater likelihood of D-O Events during Stage 8. In addition, three Heinrich Events, defined by a large abundance of DC, occurred during MIS 8, indicating surging of the Laurentide Ice Sheet. Stage 8 is therefore more analogous to that of the last glaciation than Stage 6.

Chapter 4 explores the link between HSG and cosmogenic nuclide production, which are highly coherent at a frequency of 1/950 years. A 950-year period is present in the HSG records of the last three glaciations. While a 950-year oscillation may be the product of solar forcing, due to uncertainty in paleomagnetic reconstructions and in the Site U1308 chronology, the null hypothesis that the HSG proxy does not reflect variable solar irradiance cannot be unequivocally refuted. Solar forcing does however provide an explanation for climate variability in the 950-year band during the last three glaciations.
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Chapter 1. First-Time Reproduction of the Hematite-Stained Grain Proxy for North Atlantic Climate Variability

1.1 Introduction

Marine sediments in the North Atlantic Ocean record relatively well-preserved information regarding the oceanographic and climatic conditions at the time of deposition. The lithology varies from biogenic calcareous oozes to silty clays due to fluctuating input of terrigenous and biogenic components. This is most notable on a glacial-interglacial scale, with deposition of terrigenous, ice-rafted sediment during times of cold climate. Rapid melting occurs between 45° and 50° N as icebergs encounter relatively warm water, resulting in the high accumulation of this ice-rafted debris (IRD) (Ruddiman, 1977). The concentration of IRD in this latitudinal belt varies by several orders of magnitude on suborbital- to millennial-scales. Bond and Lotti (1995) determined that these increases were often coincident with Greenland stadial events, and they attributed higher deposition to increased iceberg calving, as opposed to increased iceberg survivability, based on the discordant SST and IRD records from several marine sediment cores.

The relative concentration of several grain types is also variable on millennial time scales. High concentrations of Canadian Shield Limestone indicate that the North American Laurentide Ice Sheet was responsible for the largest ice-rafting events, known as Heinrich Events (Heinrich, 1988; Bond et al., 1992). Increases in the amount of Icelandic volcanic glass (IG) and hematite-
stained quartz and feldspar grains (HSG), which are of differing provenance, preceded Heinrich Events and indicate synchronous discharge from multiple glaciers (Bond and Lotti, 1995).

HSG in particular received considerable attention during the past decade. The late G.C. Bond identified a pervasive, 1500-year cycle in HSG over the last glacial cycle, including the last interglaciation, Marine Isotope Stage (MIS) 5e (Bond et al., 1997). During the Holocene, this cycle coincides with well-known climatic events, including the Little Ice Age, and appears to covary with cosmogenic nuclide production, indicating a link to solar variability (Bond et al., 2001). Despite the large discussion of this work (the Bond et al. 2001 Research Article in Science has been cited over 450 times, and a 1997 paper over 700 times), no researchers have reproduced these results, casting considerable doubt on these results.

1.2 Objectives

The objective of this study is to clarify the HSG methodology and demonstrate that it is reproducible. First, information regarding the exact methodology used is compiled from various sources. Next, the methodology is employed in a reproducibility study.

1.3 Background

1.3.1 Hematite-Stained Grain Sources

HSG are derived from Proterozoic and Phanerozoic red beds surrounding the North Atlantic basin (Bond and Lotti, 1995) (Figure 1). Proterozoic red beds began to form as free oxygen became abundant (Dott and Batten, 1981).
Phanerozoic red beds range in age from Cambrian to Triassic, and reflect in situ diagenesis of ferrous minerals in hot, arid climates when these continental masses were at much lower paleolatitudes (Habicht, 1979).

**Figure 1 | North Atlantic Red Beds**

Location of red beds surrounding the North Atlantic Ocean basin and cores located to monitor inferred iceberg trajectories during the last glaciation (Bond and Lotti, 1995). VM = V (R/V Vema). Arrows indicate major currents.

Given the widespread nature of red beds surrounding the North Atlantic, it is difficult to determine the exact provenance of HSG deposited on the seafloor. Sediment cores taken along major iceberg trajectories, as proposed for the last glaciation, indicate that iceberg discharge through the Gulf of St. Lawrence was a primary source for HSG, though this does not rule out increases from other locations as well (Bond and Lotti, 1995).
During the Holocene, however, core top analyses indicate that red beds from East Greenland and Svalbard, and perhaps around the Arctic Ocean, contributed significant amounts of HSG and that changes in ocean circulation as opposed to increased calving are responsible for variations in the amount of HSG deposited in the subpolar North Atlantic (Bond et al., 1997). Thus it would appear that HSG deposition from melting drift ice was controlled by differing mechanisms during the Holocene (ocean circulation changes) and during glacial times (increased discharge) (Bond et al., 1997).

### 1.3.2 HSG as a Climate Proxy

HSG is not in and of itself a proxy for climate change, rather it is a sensitive recorder of ice sheet calving during glacial times and of circulation changes during interglacial and extended interstadial periods. The best evidence for the utility of HSG as a climate proxy is circumstantial. HSG cycles exhibit relatively consistent pacing at 1500-years, which is similar to the pacing of the stadial-interstadial, Dansgaard-Oeschger (D-O) cycle (Bond et al., 1997, 1999) (See Chapter 2.7, Page 29). D-O Events, however, did not occur with such regular spacing, but rather at intervals that are approximately equal to multiples of 1500 years, i.e., a warming event may occur after a waiting time of 1500, 3000, or 4500 years. The number of events that occurred during the last glaciation decayed exponentially with increased wait time. Stochastic resonance, which requires a weak periodic signal superimposed upon white noise, potentially explains these observations (Alley et al., 2001). This would imply that neither the signal nor the noise is strong enough alone to cause a mode switch, which only
occurs with constructive interference of the two weak signals. This would produce a preferred but not perfect 1500-year pacing, as observed in Greenland air temperature. HSG is the only paleoclimate proxy to contain evidence of the periodic signal.

1.4 HSG Methodology

A detailed description of the methodology for identifying HSG is lacking from the published literature. As such, there is much confusion in the community regarding exactly how these data were generated and what was reported. In addition, prior to this study, no researcher has succeeded in reproducing the method, which appears relatively simple in that the red color of the hematite staining is the defining characteristic of HSG. The lack of reproduction has cast considerable doubt on the validity of HSG results as it is essentially an unverified proxy. Herein, all the information describing the HSG proxy are compiled into one location and data are presented that verify the reproducibility of the method.

1.4.1 Size Fraction

While bulk lithic grain counts are typically performed on the > 150 µm size fraction, all HSG counts by Bond were performed on the 63 - 150 µm size fraction using grain-mount slides analyzed with a petrographic microscope (Bond and Lotti, 1995; Bond et al., 1997, 1999, 2001). Counts of detrital carbonate grains and Icelandic glass were also performed in the same fashion. This size fraction was chosen to provide “greater accuracy in petrologic identification” (Bond and Lotti, 1995), presumably indicating that the order of birefringence is
less affected than with thicker grains, though this will be a problem with any relatively “thick” section. It would also be impractical, if not impossible to make grain-mount slides of the > 150 µm size fraction. Grain-mount slides allow multiple grain faces to be viewed at once, increasing identification efficiency (See “Petrographic Microscope”, Page 6).

Most importantly, standardized measurements from both interglacial and glacial times are facilitated by the use of the 63-150 µm fraction because its relative abundance remains high during interglacials, which is not the case for the > 150 µm fraction (Bond et al., 1999). The use of a relatively small and narrow size range also reduces variability resulting from a relationship between grain size and composition (Bond et al., 1999) without introducing a compositional bias (Bond and Lotti, 1995).

1.4.2 HSG Identification

1.4.2.1 Petrographic Microscope

Grain-mount slides and a “specially prepared microscope” are used for identification of HSG (Bond et al., 1997) (Figure 2). This is a petrographic scope that has a white reflector placed on top of the condenser, below the stage. An external, fiber-optic light source is used to transmit light through the slide and grains from above. The height of the condenser relative to the stage controls the amount of light reflected back through the slide from below. As such, “a position can be found that creates a strong impression of relief and brings into striking view details of surface textures and coatings on grains” (Bond et al., 1997).
This particular microscope-light-reflector combination is essential for accurate and efficient identification of HSG in this size fraction. Identification of HSG in loose sediment on a picking tray with a reflected light microscope would require rotating each grain under the highest magnification to expose and examine all sides. Also of great importance is a white light source with a color temperature of approximately 3000 K or higher. Lower color temperature light sources do not reveal the true, red color of the hematite staining.

**Figure 2** | Gerard Bond’s Microscope

A) Bond’s petrographic microscope using an external, fiber optic light source and substage reflector above the condenser (not visible). B) Petrographic scope used in this verification study with condenser flipped out exposing substage reflector (red box).

1.4.2.2 Binary Classification and Reporting

Hematite-stained grains are classified in a binary system, i.e., visibly stained or visibly not stained; the amount of staining is not considered (Bond et al., 1997). The minimum amount of hematite staining required for a grain to be classified as HSG is approximately the size of a pin head as visible through a 10x
objective (Figure 3); Bond used only a 10x objective for determining whether a grain was or was not stained with hematite (personal communication, Almasi, 2006). Rather than a point counting method, lines across the slide are used for counting HSG. Bond reported an error of ± 2 percentage points, the 2σ range of replicate analyses of one sample (Bond et al., 1997).

Results are reported as a percentage of the lithic fraction, as opposed to percent of total sediment or concentration (i.e., grains / gram of bulk sediment). The concentration of lithic grains may represent iceberg survivability as a function of sea surface temperature, while percentages may provide evidence of changes in circulation or iceberg discharge (Bond et al., 1999). Additionally, during interstadial and interglacial conditions, HSG concentration invariably decreases dramatically while percentages remain a robust recorder of ice sheet instability.

**Figure 3 | Minimum Hematitic Stain Area**

The minimum hematitic stain area required for classifying a grain as HSG (h) is defined by the size of the dot (placed directly in the ocular) when viewed through a 10x objective. Scale bar is 120 µm. After Bond et al (1997).

**1.4.2.3 Sample and Slide Preparation**

Grain-mounts of the 63 - 150 µm size fraction are used for petrographic identification of HSG. Standard slides and cover slips are prepared with a split
of sediment in a mounting medium. Bond used glycerin as a mounting medium
due to its clarity and lack of selective wavelength attenuation. However, this
resulted in slides that must remain horizontal and are therefore not
transportable. For this study, an optical adhesive, such as NOA 61 (Norland
Optical Adhesive 61) was used. This is the same mounting medium used for
making smear slides on IODP and ODP expeditions, and is a “clear, colorless,
liquid photopolymer that cures under ultraviolet light” (Norland Products Inc.,
2008), resulting in permanently-mounted, transportable slides of similar quality
to those of Bond (Figure 4).
Figure 4 | EW93-03 40GGC Hematite-Stained Grain

Photomicrograph of ~ 120 µm (long axis) HSG in a permanently-mounted slide using NOA 61 as a mounting medium. The top or bottom of the grain may be clearly viewed by adjusting the focus.

Bulk sand samples are split into archive and working halves with a microsplitter. The method is described as follows. 1) The working half was sieved to separate the 63 - 150 µm fraction. 2) Bond used hand-splitting for the 63 - 150 µm size fraction (personal communication, Almasi, 2006). 3) For a given sample, this entire size fraction is placed on weighing paper. The weighing paper is picked up such that the worker is holding opposite sides, with the paper making a U-shape. 4) the paper is agitated in an “up and down motion” such
that sediment collects in a linear fashion at the lowest point. 5) The paper is then rotated 90° and the process repeated several times. 6) Finally, the sample, which has now collected in an elongated oval shape, is split with a razor blade. Half of the sediment is retained and the process repeated until there is an appropriate amount of sediment remaining, such that grain overlapping is minimized and yet still enough to count ~ 200 HSG grains.

For preparation of Holocene samples, calcium carbonate is removed from the 63 - 150 µm size fraction prior to mounting due to the lower percentage of ice-rafted, lithic grains (personal communication, Almasi, 2006). This method was also employed for the MIS-6 interval of Site U1308 (reoccupied Site 609) (See Chapter 3.5.2 “MIS 6 Sampling”, Page 48).

An additional washing treatment is required for preparing sediment from certain locations for petrographic analysis. At Site 609 in particular, fine-sediment aggregates commonly remained in the sand-sized fraction when washing methods appropriate to the preservation of foraminifera were employed (Bond and Lotti, 1995). This was also a common occurrence in both the MIS-6 and MIS-8 intervals of Site U1308, with grains often being coated in fine sediment, obscuring surface textures. Thus, the 63-150 µm fraction was also briefly sonicated and re-sieved prior to slide preparation to remove residual fine sediment.

**1.4.3 HSG Methodology Reproduction**

A key element of this study was verifying the HSG identification methodology by demonstrating that the results of Bond are indeed reproducible
by another researcher. This task was complicated by the untimely passing of Gerard Bond on June 29, 2005 (LDEO, 2005; Pearce, 2005), approximately one month prior to a scheduled visit to LDEO for training.

This visit was rescheduled for May 2006. During this visit, access to the equipment and laboratory of the late Dr. Bond was provided by Ms. Rusty Lotti-Bond. Peter Almasi, a graduate student working under Bond at the time of his passing, provided detailed information regarding Bond’s techniques. Using this information and original slides prepared by Bond, work began attempting to duplicate all analytical steps followed by Bond.

Bond indicated that identification of HSG was non trivial (personal communication, Bond, 2005). While heavily coated grains are readily identifiable, neither the full amplitude nor the structure of HSG variability is captured by simply counting these easily recognized grains, which may indicate a different provenance between HSG grains deposited during intervals of peak abundance versus during intervals exhibiting only background levels (personal communication, Almasi, 2006) (See “Hematite-Stained Grain Sources”, Page 2).

An unforeseen complication was that, before his passing, Bond requested that his raw data not be made available. However, without raw data generated by Bond, this verification study would not be possible. Therefore, limited data were provided from random down-core intervals of EW93-03-40GRC that constitute a representative sampling, in terms of minimum, maximum, average, and standard deviation, of the typical HSG population in a Bond record (unpublished data, Lotti-Bond, 2006). This compromise allowed for the honoring
of Bond’s wishes with respect to the dissemination of his data while making this study possible.

1.4.4 Reverse Engineering the HSG Counting Technique

Two months were invested in “calibration of the observer” in order to produce data consistent with those of Bond, during which, over 10,000 lithic grains were categorized in this step. Given the extended time required, this work was not performed at LDEO. Thus, new slides were prepared because the original slides made by Bond are not transportable (See “Sample and Slide Preparation”, Page 8). In May 2006, fourteen subsamples were obtained from core EW93-03-40GGC (41°39.042’N, 51°0.972’W; 3722 mbsl) archived at the LDEO Deep Sea Sample Repository. This core had been previously sampled. These samples had also been processing to separate the calcium carbonate-free, 63 - 150 µm size fraction from bulk sediment. The exact same methods as those used by Bond, aside from slide mounting medium, were used in this study (See “HSG Methodology”, Page 5).

Slide preparation and petrographic analysis took place at the University of Tokyo, Ocean Research Institute during the summer of 2006. At least two grain-mount slides were created per sample, using an optical adhesive that allows for slide transportation. Petrographic analysis for determination of HSG percent was conducted on each sample while ignorant of the result from Bond. Each sample was counted approximately six times prior to checking whether the result was consistent with that of Bond.
At first, results were typically not consistent with those of Bond (Table 1). However, using an iterative process, a means for categorizing the lithic grains was developed such that the results, over time, became consistent with Bond’s results. These results also demonstrate that the methods for slide preparation did not bias the population structure of the sediment grains, as results.

**Table 1 | HSG Method Initial Calibration**

<table>
<thead>
<tr>
<th>Slide #</th>
<th>Count #</th>
<th># of HSG</th>
<th>Total Lithics</th>
<th>HSG (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>20</td>
<td>182</td>
<td>11</td>
</tr>
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<td>3</td>
<td>11</td>
<td>144</td>
<td>8</td>
</tr>
</tbody>
</table>

Example of calibration for EW93-03-40GGA 54 cm, resulting in an average of 9% with a standard deviation of 3 percentage points. These counts were performed without knowledge of Bond’s result, which is 2% HSG.

1.5 Results

One slide from each sample was randomly selected, sample number obscured, and petrographically analyzed for HSG percent. Raw data from these samples generated by Bond were arranged into an artificial HSG cycle ranging from 2% to 35% IRD for comparison with the results of this study (Table 2).
<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Depth (cm)</th>
<th>HSG (%) Bond</th>
<th>HSG (%) This Study</th>
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<td>1</td>
<td>54</td>
<td>2</td>
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<td>12</td>
<td>2</td>
<td>3</td>
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</table>

Sample numbers and corresponding depths for EW93-03-40GGC that create an artificial cycle in HSG. HSG data are from Bond’s original counts (unpublished data, Lotti-Bond, 2006) and the method verification study.

The results of this study are consistent with those of Bond. Nine of fourteen samples fall within the reported error of Bond, ± 2 percentage points (Bond et al., 1997) (Figure 5). All but two of the remaining five samples exhibit overlapping error, assuming the same 2σ range. Results for two samples, samples 1 (54 cm) and 8 (100 cm) were significantly different from those of Bond. During subsequent recounts, including those performed with duplicate slides, reproducing Bond’s results in these two intervals remained problematic.
Hematite-Stained Grain (% of lithic fraction)

Sample Number

1 2 3 4 5 6 7 8 9 10 11 12 13 14

0 5 10 15 20 25 30 35

Bond
This study

r = .92

1.6 Discussion

This study reproduces, for the first time, the HSG proxy for North Atlantic climate variability. This verification study required over two months to complete, during which time over 10,000 siliciclastic grains were categorized. This chapter stops short of resampling the last glacial interval and wholly reproducing the data of Bond. Rather, these energies are spent reevaluating...
published results (See Chapter 2, “DSDP Site 609 Chronology: Implications for HSG Cyclicity”, Page 18) and applying this proxy to the two glaciations (MIS 6 and 8) directly preceding the well-studied last glacial stage (See Chapter 4, “Significance of Reproducible Millennial Ice-Rafting Signals During the Last Three Glaciations”, Page 86).

1.7 Conclusions

The HSG method (Bond and Lotti, 1995; Bond et al., 1997, 1999, 2001) is reproducible. The minimum stain area used to define HSG is too small for accurate categorization of loose sediment with a reflected-light microscope as commonly used for identification of foraminifers. Grain-mount slides analyzed with a properly configured petrographic microscope with a white substage reflector and a bright, white fiber optic light source with a color temperature of at least 3000 K are required. A significant investment of time for calibration of the observer is also necessary.
Chapter 2. DSDP Site 609 Chronology: Implications for HSG Cyclicity

2.1 Introduction

DSDP Site 609, along with V23-81, was instrumental in linking the marine record to the climate variations recorded in Greenland ice cores. The correspondence between SST and Greenland air temperature proxies allowed the recognition that successively cooler D-O Events were packaged between extremely cold (Heinrich) ice rafting events (Bond et al., 1993). These sites are located in the North Atlantic IRD Belt and record a high abundance of both total IRD (Ruddiman, 1977) and HSG (Figure 6).

HSG exhibits high-amplitude fluctuations at Site 609 (Bond et al., 1999).

HSG are eroded by glaciers overlying red beds primarily located around the Gulf of St. Lawrence (Bond and Lotti, 1995) (See Chapter 1.3.1 “Hematite-Stained Grain Sources”, Page 2). Radiocarbon-dating of DSDP Site 609 and
V23-81 revealed that during MIS 2, deposition of HSG increased dramatically on average every 1470 years (Bond et al., 1997). The chronology for the older interval of these sediment cores was derived by correlating to the Greenland GISP2 ice core, which resulted in a similar pacing of HSG events during Stages 3 and 4 (Bond et al., 1999).

Therefore, HSG was often described as exhibiting a “1500-year cycle” (Bond et al., 1997, 2001). A geologic definition of cycle was adopted that, unlike the statistical definition, did not imply periodicity but, rather, repetition (Bond et al., 2001). The term “quasi-periodic” was often employed, indicating that increases in HSG did not occur at a precise frequency. The longest portion of the Site 609 record was tuned to the GISP2 ice core. A more recent ice core (North GRIP) has a better constrained chronology (Svensson et al., 2006), which may have implications for the pacing of HSG events.

2.2 Objectives

The objective of this study is to update the chronology of classic DSDP Site 609 beyond the relatively short radiocarbon-dated interval by correlation to the North GRIP (NGRIP) ice core (Andersen et al., 2004). Next, given a new chronology, the robustness of conclusions regarding both the pacing and spectral nature of the Site 609 HSG record are considered.

2.3 Study Area: DSDP Site 609

North Atlantic Site 609 (49° 52.7’N, 24° 14.3’W; 3884 mbsl) was drilled within the “IRD belt” of Ruddiman (1977) in 1983, during Leg 94 of the Deep Sea Drilling Project, and is located on the upper-middle eastern flank of the Mid
Atlantic Ridge, approximately 240 km south of the Charlie Gibbs Fracture Zone at the modern boundary between the subpolar and subtropical gyres (For location map, see Chapter 3.4 “Study Area: IODP Site U1308/DSDP Site 609”, Page 43) (Ruddiman et al., 1987). Site 609, as opposed to V23-81, is focused on here because it was subsequently reoccupied in 2004 on Integrated Ocean Drilling Program Expedition 303 (Site U1308) (Expedition 303/306 Scientists, 2006). The use of modern coring and splicing techniques allowed for the recovery of a demonstrably complete sedimentary section from this location to a composite depth of 355 meters below the sea floor, thus extending the continuous Site 609 record into the Pliocene (Expedition 303 Scientists, 2006). Site U1308 is the subject of subsequent chapters.

2.4 Background

The radiocarbon chronology of DSDP 609 resulted in good agreement with GISP2 (Bond et al., 1997). Therefore, from 26 ka, after which no radiocarbon dates exist for Site 609, the *N. pachyderma* (s) SST proxy record was correlated to GISP2 to provide suborbital age control for Site 609 (Bond et al., 1999) (Figure 7). The current chronology for the GISP2 ice core is the Meese/Sowers 1994 (M/S94) chronology, which is based on layer-counting to ~ 50 ka and correlation of $\delta^{18}O$ of oxygen bubbles to the Antarctic Vostok ice core (Bender et al., 1994; Meese et al., 1997). However, very high-density sampling of the more recent, virtually complete North GRIP (NGRIP) ice core (Andersen et al., 2004) indicates that GISP2 suffers from excessive layer thinning (Svensson et al., 2006). Layer
counting of NGRIP to 60 ka (GICC05) (Svensson et al., 2008) resulted in a significantly older ages for MIS 4 (up to 4000 years; Figure 8).

Therefore, it is conceivable that an updated chronology may require reevaluation of conclusions based on previous work at Site 609 with respect to the pacing of HSG. Interval counts between HSG cycle midpoints indicated that the elapsed time between HSG events was 1476 ± 584 years (1σ) (Bond et al., 1999). HSG records do, however, exhibit periodicity, but the most significant frequency is 1/1250 years for Site 609 (Figure 9).
Beyond the reliable radiocarbon calibration to calendar years, the Site 609 \textit{N. pachyderma (s)} SST proxy record was correlated to the GISP2 ice core record on the M/S94 chronology ($r = 0.82$) (Bond et al., 1993). Because the \textit{N. pachyderma (s)} record is unavailable as raw data in calendar years, figure was reproduced by digitizing the Site 609 record in Bond et al. (1999) and therefore is not an exact representation of the data.
NGRIP (black, GICC05 ss09sea) and GISP2 (blue, M/S94) on their respective chronologies. GISP2 become progressively younger than NGRIP.
The published Chronology for DSDP Site 609 for MIS 2 - 4 is based primarily on correlation of *N. pachyderma* (s) SST proxy record to GISP2 and results in primary spectral power at ~ 1/1250 years. Significance lines are 99% and 95%.

### 2.5 Methods

The unavailability of the Bond data complicates this process (See Chapter 1.4.3 “HSG Methodology Reproduction”, Page 11). Only the > 150 μm data for Site 609 are currently available on the NCDC paleoceanography data repository (Bond et al., 1992; Bond, 1996), though HSG data from Bond et al. (2001) was added in February, 2008, the first time Bond HSG data have been made publicly available on the repository (Bond et al., 2008). Not available for Site 609 are
the 63 - 150 µm data (HSG, detrital carbonate, and volcanic glass) and the tie points to GISP2.

The 63 - 150 µm data sets were received (personal communication, Voelker, 2007) but only in the calendar-year domain based on radiocarbon dates and the M/S94 chronology. No information regarding 1) sample depths and IDs, 2) radiocarbon ages, or 3) age-depth tie points are available. Thus the HSG (as well as detrital carbonate and volcanic glass) data are unlinkable to the *N. pachyderma* (s) record. Therefore, updates to the 609 chronology attained by simply retuning this SST proxy record to NGRIP cannot be applied to the 609 HSG record.

Therefore, the Site 609 HSG calendar-year chronology was updated indirectly by correlating GISP2 to NGRIP. This method relies on the original tie points of Bond and Lotti (1995) and Bond et al. (1999) (Figure 7). First, the GRIP ss09sea flow model was applied to the non-layer counted interval of NGRIP beyond 60 ka (e.g., Andersen et al., 2004). The ss09sea time scale was shifted by 705 years toward younger ages to (personal communication, Svensson, 2008). Next, the GISP2 \( \delta^{18}O \) record was resampled to the time scale of Site 609 HSG, resulting in both data sets on a common scale (Figure 10). Finally, the resampled \( \delta^{18}O \) record of GISP2 was aligned to that of NGRIP, exporting this new chronology to the Site 609 HSG record (Figure 10). Detrital carbonate and volcanic glass were updated as well, and these data can be used as a cross check on the validity of the updated chronology. The packaging of progressively colder D-O Events between Heinrich Events (“Bond Cycle”), though first noted for the GRIP ice core (Bond et al., 1993), was also evident between Site 609 and
GISP2. After updating the Site 609 63 - 150 µm petrologic tracer records to the GICC05 chronology, this relationship remains robust between Site 609 detrital carbonate and NGRIP $\delta^{18}O$ (Figure 11).

**Figure 10** | DSDP Site 609 HSG Updated to GICC05 Chronology

Update of Site 609 HSG data from M/S94 chronology (blue) to GICC05/ss09sea chronology (black). A) NGRIP on GICC05 and ss09sea. B) GISP2 on M/S94 and resampled to to Site 609 time scale (red). C) Original Site 609 HSG chronology (blue) compared to updated chronology (black).
Verification of updated Site 609 chronology using detrital carbonate to confirm that successively cooler D-O Events remains packaged between Heinrich Events after updating from the A) M/S94 GISP2 chronology to the B) GICC05/ss09sea NGRIP chronology.

2.6 Results

The M/S94 chronology resulted in an average elapsed time between HSG midpoints of $1476 \pm 584$ years at Site 609. After updating the Site 609 record to
the GICC05/ss09sea chronology of the NGRIP ice core, the elapsed time between midpoints of HSG events is 1573 ± 620 years (Figure 12). Spectral analysis is performed using the SSA-MTM Toolkit (Allen et al., 2008), a standard tool in the field of paleoceanography and paleoclimatology. Relative to an AR(1) (red noise) background, the Site 609 HSG record on the new GICC05 chronology exhibits 99% confident peaks of similar spectral amplitude at periods of 1800, 1415, and 1200 years (Figure 13). By chance, a 1/1415-year spectral peak is present. While the spectrum is not dominated by variance near this band, this does indicate the record contains statistically significant periodicity near the Bond frequency of 1/1500 years.

**Figure 12** | Mean Pacing of Site 609 HSG Events (GICC05/ss09sea chronology)

The updated age model for Site 609, based on the NGRIP GICC05 layer-counted chronology (to 60 ka) and the ss09sea flow model, results in an average cycle length of 1573 ± 620 years.
Figure 13 | Site 609 HSG Power Spectrum (GICC05/ss09sea Chronology)

MTM power spectrum for Site 609 HSG after updating the chronology to NGRIP GICC05/ss09sea (3 tapers, frequency from 0 to 1/500 years). Significance lines are 99% and 95%.

2.7 Discussion

The M/S94 chronology of the GISP2 ice core resulted in a mean HSG event spacing of $1476 \pm 585$ years at Site 609 (Bond et al., 1999), which is very similar to the primary spectral frequency of $1/1470$ year detected in the GISP2 $\delta^{18}O$ record (Grootes and Stuiver, 1997) (Figure 14). However, the spectral amplitude at this frequency in GISP2 has been subsequently shown to be non-stationary, reaching maximum amplitude between 31 and 36 ka during the
interval containing the extremely regularly spaced D-O Events 5, 6 and 7 (Schulz, 2002) (Figure 15). Schulz (2002) also demonstrated that these three D-O events were solely responsible for the statistical significance of the 1/1470 year spectral peak.

**Figure 14 | GISP2 MIS 2-4 Power Spectrum**

GISP2 δ¹⁸O (M/S94 chronology) MTM power spectrum (3 tapers, frequency from 0 to 1/500 years). Significance lines are 95% and 99%. Time series between 14 and 71 ka (MIS 2 and 4 boundaries in LR04 benthic isotope stack) was interpolated to a 100 year sampling. Sharp spectral peak is centered at 1470 years.
Interstadial warmings (D-O Events) numbered in the GISP2 δ¹⁸O ice core record (Johnsen et al., 1992; Dansgaard et al., 1993). The regular spacing of D-O Events 5, 6, and 7 are the reason for statistically significant 1/1470-year spectral density.

Application of the NGRIP GICC05/ss09sea chronology results in a mean HSG pacing of 1573 ± 620 years, which is not significantly different from the spacing of 1476 ± 585 years produced by the M/S94 chronology (Bond et al., 1999). This slight difference may result from differences in midpoint determination, as the exact location of the midpoints determined by Bond et al. (1999) is unknown. Alternatively, this could reflect a subtle difference in the NGRIP δ¹⁸O power spectrum, which unlike GISP2, contains two 99% confident, statistically significant spectral peaks at 2775 and 1540 years (Figure 16). The mean 1573-year pacing (GICC05/ss09sea) of HSG events is very close to the millennial-band 1/1540-year frequency observed in NGRIP. This same relationship existed between the GISP2 primary spectral peak (1/1470 years) and the original mean spacing of HSG events (1476 years). Nonetheless, given the
existence of three millennial-band spectral peaks significant at a 99% confidence level in the updated HSG time series, more advanced spectral analysis (e.g., Monte Carlo singular spectrum analysis) is warranted prior to publication of these results in order to better constrain the degree of statistical significance of these frequencies.

**Figure 16 | NGRIP MIS 2-4 Power Spectrum**

NGRIP δ¹⁸O (GICC05/ss09sea chronology) MTM power spectrum (3 tapers, frequency from 0 to 1/500 years). Significance lines are 95% and 99%. Time series between 14 and 71 ka (MIS 2 and 4 boundaries in LR04 benthic isotope stack). Sampling interval is 50 years. 99% confident peaks are at 1540 and 2775 years.
There are other options for updating the Site 609 HSG chronology. The NGRIP GICC05/ss09sea chronology, as opposed to the SFCP04 chronology (Shackleton et al., 2004) is chosen for several reasons. First, the chronology for Site 609 HSG was originally based on ice core correlation, and, in the absence of a common depth scale for HSG and the other Site 609 proxy data, the best course of action is to remain on a Greenland-based chronology. Second, significant deviations occur between GICC05 and SFCP04, that, if the latter is the correct chronology, would require too few annual layers from ~ 15 – 28 ka and too many annual layers from ~ 28 – 38 ka (Svensson et al., 2006). It is more likely that the radiocarbon-based SFCP04 may suffer from $^{14}$C calibration and/or reservoir age issues; a - 1ka adjustment to SFCP04 at the onset of Greenland Interstadial 3 reconciles the records to within the reported age uncertainty.

### 2.8 Conclusions

The classic DSDP Site 609 63 - 150 µm petrologic tracer records of hematite-stained grains, Icelandic volcanic glass, and Canadian Shield detrital carbonate are placed in the NGRIP GICC05 layer-counted chronology to 60 ka. The ss09sea flow model is used beyond 60 ka. The mean HSG event spacing is slightly longer at 1573 ± 620 years, which potentially corresponds to the primary millennial frequency of 1/1540 years in the NGRIP ice core. Three 99% confident, millennial-band spectral peaks are identified in the updated HSG time series at 1/1800, 1/1415, and 1/1200 years. Additional time series analysis is required to determine the degree of statistical significance of these frequencies.
Chapter 3. Paleoceanographic Differences Between the Last Three Glaciations

3.1 Introduction

Flow through the Straits of Florida constitutes the largest component of the North Atlantic western boundary current, the Gulf Stream. The Gulf Stream diverges from the North American shelf break at Cape Hatteras, forming the northeastward North Atlantic Drift, which transports warm, saline water into the Norwegian-Greenland Sea where it is cooled and subducted. The Norwegian-Greenland Sea basin is separated from the North Atlantic by the shallow Greenland-Scotland Ridge. Deep water overflows this ridge, primarily through the Denmark Strait between Greenland and Iceland but also through the Iceland and Faeroe-Shetland Channels, forming the northernmost limb of the meridional overturning circulation and, at an estimated total of 20 Sv, the greatest volumetric component of North Atlantic Deep Water (NADW) (Dickson et al., 1990). However, during the last glacial maximum (LGM) the polar front was located farther south at ~ 45°N, resulting in a steep north-south thermal gradient and diminished oceanic heat transport to high northern latitudes (Climap Members, 1976). Correspondingly, the geostrophic flow through the Straits of Florida was reduced indicting decreased formation of NADW (Lynch-Stieglitz et al., 1999).

Broecker (1998) proposed that changes in thermohaline circulation were intimately linked to the Greenland D-O cycle during the last glaciation. The decreased North Atlantic sea-surface temperature (SST) and ice-rafting events
concurrent with Greenland D-O cycles, first noted at DSDP Site 609 (Bond et al., 1993), have since been detected in a number of North Atlantic marine records (e.g., Voelker, 2002). These basin-wide temperature fluctuations, observed as far south as the northern Sargasso Sea at ~ 30˚N, are most likely indicative of a regional pattern as opposed to a local response to changing oceanic front positions (Sachs and Lehman, 1999) and provide evidence of decreased northward heat transport during Greenland stadial events (Keigwin and Boyle, 1999). This is consistent with observed warming of Antarctic air temperature synchronously with Greenland stadials and North Atlantic Heinrich Events (EPICA Community Members, 2006) (Figure 17) and can be explained by the cooling effect of NADW formation on the southern hemisphere (Crowley, 1992).

The D-O pattern of variability was not limited to the Atlantic Basin. It is globally observed in proxy records from locations including the Santa Barbara Basin (Hendy et al., 2002), the Arabian Sea (Schulz et al., 1998), and China (Wang et al., 2001, 2008). A “1500-year cycle” in the percentage of ice-rafted hematite-stained grains (HSG) and IG during the last glaciation is postulated to be pace the D-O cycles. Thus, the ice-rafting events associated with Heinrich Events and Greenland stadials are a potential mechanism for decreasing NADW formation through buoyancy forcing and, therefore, potentially explain much of the observed global-scale millennial climate variability of the last glaciation.
A study of millennial climate during previous glaciations indicates that variability is related to continental glacier size, with the greatest stability observed during times of maximum and minimum ice volume (McManus et al., 1999). Ice volume during the last (MIS 2 - 4) and penultimate glaciations (MIS 6), as inferred from a globally-distributed, stacked benthic foraminiferal $\delta^{18}O$ record (Lisiecki and Raymo, 2005), appears to have been strikingly similar during all
glacial phases, from ice growth to decay. Both glaciations exhibit not only a nearly identical duration, but also a comparable amplitude of variation (Figure 18). Glacial Stage 8, on the other hand, exhibits generally lighter values, with a much less pronounced glacial maximum and a significantly lighter interstadial event (MIS 8.5) than both of the subsequent glaciations (Figure 18). Thus, it would be reasonable to expect that the millennial-scale variability of MIS 6 was similar to that of MIS 2 - 4.
The last three glaciations overlain such that the LR04 glacial stage boundaries (14, 130, and 243 ka) are approximately aligned. The statistical MIS 1/2 boundary occurs 3 ky earlier relative to the previous two boundaries.

3.2 Objectives

The objective of this study is to extend the MIS 2 - 4 HSG, Icelandic volcanic glass (IG), and detrital carbonate (DC) records of Site 609 (Bond et al., 1999) to the preceding two glacial intervals, MIS 6 (130 - 191 ka) and 8 (243 - 300 ka).
ka) (Figure 19), at new IODP Site U1308 (reoccupied Site 609) in order to test the hypothesis that a “1500-year cycle” and Heinrich Events were manifest prior to the last glacial cycle. Alternatively, in the absence of such ice-rafting events, what was the nature of North Atlantic hydrographic changes. Such a test is important because statistical considerations suggest that stochastic events can by chance occur at a statistically significant period for a particular glacial or interglacial stage (Hyde and Crowley, 2002), but it is much less likely that essentially identical, periodic oscillations could originate by chance alone if they occur in other glacial “realizations”.

**Figure 19** | Marine Stable Isotope Record of the Last Three Glaciations

The LR04 (Lisiecki and Raymo, 2005) stacked benthic foraminifer $\delta^{18}$O record with dates of stage boundaries illustrates the temporal intervals for 1) the original Site 609 HSG record (Bond et al., 1999) from the last glaciation and 2) the reproducibility study of the preceding two glaciations at the the reoccupied Site 609 (U1308).
3.3 Background

The petrologic and planktic foraminifer records of DSDP Site 609 were instrumental in linking variations in the North Atlantic marine records to Greenland air temperature changes, with the SST proxy exhibiting a D-O pattern and IRD deposition increasing during stadials (Bond et al., 1993). The fluctuations in HSG and IG were shown to covary during the last glaciation, indicating synchronous iceberg discharge from (at least) the Gulf of St. Lawrence and Iceland (Bond and Lotti, 1995) (See Chapter 1.3.1 “Hematite-Stained Grain Sources”, Page 2) (Figure 20). This provides evidence for climatic influences on unstable glaciers during glacial times (Bond and Lotti, 1995; Bond et al., 1999) and for southward advection of cool, ice-laden water during long interstadials (Bond et al., 1997, 1999) (See Chapter 1.3.2 “HSG as a Climate Proxy”, Page 4).
As noted earlier, SST and IRD records throughout the North Atlantic exhibit similar fluctuations linked to the Greenland D-O cycles (e.g., Voelker, 2002) (Figure 21), though it cannot be ruled out that this wide-scale correspondence stems from the common practice of tuning the SST records to the
Greenland air temperature record. However, an independently dated SST record from MD95-2040 off the Iberian margin shows similar fluctuations in a radiocarbon chronology calibrated by paired $^{14}$C and $^{230}$Th measurements on pristine corals (Shackleton et al., 2004) and was used to revise the GRIP ice core model-based chronology (ss09sea flow model) of (Johnson et al., 2001). Thus the link between North Atlantic SST and Greenland air temperature is reproducible on an absolute time scale within the range and uncertainty of radiocarbon dating.
Comparison of ice rafting and SST for Site 609 (top; GICC05/ss09sea), M23414 (middle), and MD95-2040 (chronology of de Abreau et al., 2003; bottom), which is proximal to the stacked MD01-2443,4 record for which there are no published IRD data.

### 3.4 Study Area: IODP Site U1308/DSDP Site 609

IODP Site U1308 (59°52.7’N, 24°14.3’W; 3871 mbsl) (Figures 22 and 23), a reoccupation of DSDP Site 609 (See Chapter 2.3 “Study Area: DSDP Site 609”,

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**Figure 21 | Site 609 IRD Record Compared to Other Locations**

- N. pachyderma (s) %
- % lithics
- SST °C
Page 19), was drilled in 2004 during IODP Exp. 303 (Expedition 303 Scientists, 2006). Most of the work on Site 609 has been conducted on the interval younger than Stage 6, partly due to the lack of a pristine composite record prior to the last glacial cycle. This, along with the current condition of the heavily-sampled Site 609 cores, which have suffered from desiccation/contraction and mold-grown during storage, precludes extending the classic records of Site 609 without redrilling (Expedition 303/306 Scientists, 2006).

Therefore, to address the primary climate-related Expedition 303 objectives concerning whether or not the “1500-year cycle” is evident in previous glacial intervals, six holes were drilled at Site U1308 ensuring recovery of a complete, undisturbed composite section. To minimize coring disturbance, the advanced hydraulic piston corer (APC) was used exclusively. This large number of holes was required due to the initially poor sea-state conditions experienced on station that caused excessive ship heave. Ultimately, a demonstrably complete composite section was spliced together using five holes between 0 and 247.14 meters composite depth (mcd) (2006).
The IODP North Atlantic Climate Expedition was broken into two legs, Exp. 303 (9/15/2004 - 11/17/2004) and Exp. 306 (3/2/2005 - 4/26/2005). Site 609/U1308 (red text) is located at 49˚53'N, 24˚14'W. Other core locations, from north to south: ENAM 97-09(54˚32'N, 16˚22'W), M23414 (53˚32'N, 20˚17'W), Site 609/U1308 (49˚53'N, 24˚14'W), TP88-9P (48˚23'N, 25˚5.'W), SU90-08 (43˚30'N, 30˚24'W), MD95-2040 (40˚35'N, 9˚52'W), MD01-2443 (37˚53'N, 10˚11'W), MD01-2444 (37˚34'N, 10˚9'W), MD95-2037 (37˚5’N, 32˚2’W), and MD95-2036 (33˚41’N, 57˚35’W). Base map after Expedition 303/306 Scientists (2006)
Site U1308 is located approximately 240 km south of the Charlie Gibbs Fracture Zone in 3871 m of water at 29°52.7′N latitude and 24°14.3′W longitude.

### 3.5 Methods

Analyses for this study were performed on IODP Site U1308. Methods are identical to those of the Site 609 studies (Bond et al., 1992, 1993; Bond and Lotti, 1995; Bond et al., 1997, 1999). Petrologic analyses of the lithic fraction was
conducted at a 1-cm interval to determine percentages of HSG, IG, and DC. Site U1308 was sampled at the University of Florida. Samples were washed according to standard practices for preserving foraminifera at the University of Florida and picked for planktic and benthic stable isotope analysis of *N. pachydera* (s) and *C. wuellerstorfi* (Hodell et al., submitted).

### 3.5.1 Chronology

The chronology for Site U1308 is based on correlation of benthic foraminifer $\delta^{18}O$ (Hodell et al., submitted) to the LR04 $\delta^{18}O$ stack (Lisiecki and Raymo, 2005) (Figure 24). As the current Greenland ice core record does not extend beyond the last interglacial (MIS 5), creation of a millennial-resolution chronology is not possible for MIS 6 and 8, unlike at Site 609, which was tied to Greenland $\delta^{18}O$ during the last glaciation. See “Chronological Uncertainty” (Page 63) expanded discussion.
The chronology for Site U1308 is based on *C. wuellerstorfi* δ^{18}O (Hodell et al., submitted) tied to the LR04 stack (Lisiecki and Raymo, 2005) at 8.75 mcd (127 ka), 9.17 mcd (136 ka), 11.54 mcd (191 ka), 15.32 mcd (243 ka), and 21.42 mcd (329 ka).

### 3.5.2 MIS 6 Sampling

Petrologic and stable isotope analyses were performed on the same samples. Site U1308 was sampled along the shipboard spliced section at a 1-cm interval within the MIS-6 section. Outside of this section, Site U1308 was sampled at a 2-cm interval. Subsequently, an error in the spliced section was detected and new composite splice for Site U1308 was constructed (Table 3) (Hodell et al., submitted) that differs from that published in the Expedition Proceedings (Expedition 303 Scientists, 2006). In the corrected composite section, the MIS-6 interval is from 8.84 to 11.56 meters composite depth (mcd). There is
one splice point at 10.42 mcd at a light-colored, high calcium carbonate layer (Figure 25). This resulted in a short interval of 2-cm resolution samples being incorporated in the spliced MIS-6 section. For this study, a slight deviation from the corrected composite splice compensated for all but a 4-cm continuous interval. This very short interval was not analyzed, resulting in continuous 1-cm sampling from 8.84 to 10.32 mcd and from 10.36 to 11.56 mcd for a total of 267 samples (Table 4).

Table 3 | Site U1308 Corrected Composite Splice Table For MIS-6 Section

<table>
<thead>
<tr>
<th>Hole, core, section, interval (cm)</th>
<th>Depth</th>
<th>Hole, core section, interval (cm)</th>
<th>Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mbsf</td>
<td>mcd</td>
<td>mbsf</td>
</tr>
<tr>
<td>1308E, 1H, 7, 10</td>
<td>9.1</td>
<td>10.42</td>
<td></td>
</tr>
<tr>
<td>1308C, 2H, 5, 106</td>
<td>11.46</td>
<td>14.42</td>
<td></td>
</tr>
<tr>
<td>tie to</td>
<td>1308C, 2H, 3, 6</td>
<td>7.46</td>
<td>10.42</td>
</tr>
<tr>
<td>tie to</td>
<td>1308E, 2H, 2, 12</td>
<td>11.12</td>
<td>14.42</td>
</tr>
</tbody>
</table>

The MIS-6 Section of Site U1308 has one splice point splice point at 10.42 mcd. Mbsf is meters below sea floor. After Hodell et al (submitted).
Shipboard sediment color data illustrates the location of the single splice point within the MIS-6 section of Site U1308 at 10.42 mcd.

### Table 4 | Site U1308 MIS 6 Sampling

<table>
<thead>
<tr>
<th>Interval top</th>
<th>Depth (mcd)</th>
<th>Interval bottom</th>
<th>Depth (mcd)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1308E-1H-6A 2-3 cm</td>
<td>8.84</td>
<td>1308E-1H-6A 149-150 cm</td>
<td>10.31</td>
</tr>
<tr>
<td>1308C-2H-3A 0-1 cm</td>
<td>10.36</td>
<td>1308C-2H-3A 119-120 cm</td>
<td>11.55</td>
</tr>
</tbody>
</table>

The MIS-6 section is sampled at a 1-cm interval from 8.84 to 11.56 mcd, with a 4-cm gap between 10.32 and 10.36 mcd. Table depths are for the top of each sample; sample identifications refer to Hole, Core, Archive Section, Interval (cm).

### 3.5.3 MIS 8 Sampling

The MIS-8 section of the Site U1308 composite section is contained entirely within Core U1308E-2H and therefore contains no splice points. This core was also processed at the University of Florida in the same way as the Stage-6 section. However, the isotopic sampling at a 2-cm interval was too coarse for this study, requiring sampling of an alternate Site 1308 hole. Core 1308A-2H, which had not been previously sampled and exhibited no signs of coring disturbance within the
desired interval, was sampled between 12.88 (1308A-2H-3A 128 cm) and 17.59 (1308A-2H-6A 150 cm) meters below sea floor (mbsf) (Figure 26). resulting in 468 samples.

Core 1308A-2H shows no signs of coring disturbance within the MIS-8 interval from Section 3 128 cm to Section 6 150 cm.

As a consequence, stable isotope analyses were performed on Core 1308E-2H while petrologic analyses were conducted on samples from Core 1308A-2H. These two cores experienced differential compaction during coring,
requiring adjustment of the depth scale of Core 1308A-2H to match that of 1308E-2H (Figure 27). This was accomplished by tying the Core 1308A-2H sediment lightness (L*) record to Core 1308E-2H at the peak whiteness values of MIS 7.5 after glacial termination III to create a preliminary composite depth scale for Core 1308A-2H (Figure 27). Moving away from the tie point at MIS 7.5, the progressive offset becomes increasingly apparent. Therefore, the Core 1308A-2H depth was adjusted to correct for this by correlating the L* records for each core (Figure 28a) (Table 5). Finally, petrologic tracer data from Core 1308A-2H were compared to scanning XRF data from from Core 1308E-2H to verify that the depth scales matched (Figure 28b).
Creating a composite depth scale for Core 1308A-2H requires correlation to Core 1308E-2H, which comprises the entire MIS-8 interval of the composite splice and is offset horizontally by ~ 95 m. A) Shipboard core-scanner images showing progressively increasing vertical offset between these two adjacent cores with examples of tie points. B) Shipboard sediment lightness ($L^*$) data showing tie point at 12.50 mbsf in Hole A to 14.58 mcd (MIS 7.5) in Hole E. Stage naming refers to SPECMAP conventions (Imbrie et al., 1984). In the composite splice, Core 1308C-5H is spliced into 1308E-2H at 14.42 mcd.
A) Depth-corrected L* record for Core 1308A-2H. B) Detrital carbonate (percent of 63 - 150 µm lithic fraction from Core 1308A-2H compared to scanning XRF data from Core 1308E-2H (Ca/Sr) which is a proxy for concentration of Paleozoic Calcium Carbonate (Hodell et al., submitted).

**Table 5 | Site U1308 MIS 8 Sampling for Petrologic Analysis**

<table>
<thead>
<tr>
<th>Interval top</th>
<th>Depth</th>
<th>Interval bottom</th>
<th>Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mbsf</td>
<td>mcd</td>
<td>mbsf</td>
</tr>
</tbody>
</table>

The MIS-8 section is sampled at a 1-cm interval from 12.88 to 17.59 mbsf in Core 1308A-2H. Table depths are for the top of each sample.
3.5.4 Analyses

3.5.4.1 Analysis of the 63 - 150 µm Size Fraction

Petrologic analysis of the 63 - 150 µm size fraction was performed at a 1-cm interval at Akita University, Japan on 267 (250 year resolution) samples for MIS 6 and 468 samples for MIS 8. For the MIS-6 interval calcium carbonate was digested from the 63 - 150 µm fraction. Therefore, for MIS 6, only HSG and IG were analyzed in this size fraction, while HSG, IG, and DC were analyzed in the MIS-8 interval. Grain-mount slides were prepared and analyzed as described in Chapter 1.4 (Page 5) using a petrographic microscope with substage reflector and an external light source (See Figure 2, Page 7).

3.5.4.2 Analysis of the > 150 Size Fraction

Analysis of the > 150 µm fraction was conducted at a 2-cm interval (500 year resolution) on 142 samples for MIS 6 only. A representative subsample of the > 150 µm fraction was obtained using a microsplitter with the goal of retaining at least 200 specimens of the left-coiling planktic foraminifer *N. pachyderma*. The entire subsample was placed on a picking tray and counted with a reflected-light microscope for 1) *N. pachyderma* (s), total planktic foraminifers, and total lithic grains, as well as for IG, DC, HSG, and other trace components such as mica and chalk. DC data are only from the > 150 µm fraction.

As with Site 609 (Bond and Lotti, 1995), aggregates of fine fraction sediment were common throughout the MIS-6 interval. When required, these mud clasts were disaggregated to verify that they were not lithic grains. Dilute
HCl was used to determine whether questionable lithic grains were comprised of calcium carbonate. Chalk grains are visually distinguishable from DC under low magnification based on the bright white color of the former. Selected DC grains were analyzed by DCP-OES.

3.6 Results

3.6.1 MIS 6

Selected data for Site U1308 from MIS 6 that will be discussed in subsequent sections are shown in Figure 30. Results for > 150 µm fraction IG were broadly consistent with results from grain mount slides (63 - 150 µm; not shown), but due to the low number of grains per sample typically counted, these data are not reported here. HSG results for the > 150 µm and 63 - 150 µm fractions were less consistent (not shown), likely due to the employed counting method and the low number of grains within each subsample. (See Chapter 1.4.1 “Size Fraction”, Page 5).

The percentage of HSG and IG in the 63 - 150 µm fraction do not covary. DC (> 150 µm) is virtually absent from the entire interval including Heinrich Event 11 (H11) at Termination II (~ 130 ka). Of the 77,325 total > 150 µm-sized grains counted in the MIS-6 interval, only 112 were classified as DC. The N. pachyderma (s) SST proxy data are decoupled from both total lithic content relative to the mass of bulk sediment (grains/g) and the percentage of lithic grains. Additionally, lithic percent and lithics/gram are highly correlated (r = 0.92). IG exhibits several intervals of highly reduced deposition. Finally, the maximum accumulation of lithic grains is ~ 400 grains/g, and there are no
pronounced, rapidly decreasing spikes in the planktic $\delta^{18}O$ record, though there is variability related to the amount/percentage of lithic grains.

**Figure 29** | Site U1308 Petrologic and Foraminifer Records for MIS 6

Site U1308 petrologic, planktic foraminifer, and benthic foraminifer data for MIS 6. All analyses were performed on the same samples. Nps = *N. pachyderma* (s).
3.6.2 MIS 8

All petrologic analyses were performed on the 63 - 150 μm fraction. Three rapid increases in the percentage of DC, exceeding 20% of the lithic fraction occur beginning at 263 ka just after benthic δ¹⁸O exhibited the highest values (~ 4.2‰) recorded during MIS 8 (Figure 30). These DC events are accompanied by rapidly decreasing planktic δ¹⁸O. Preliminary analysis of the > 150 μm size fraction indicates that these intervals are virtually devoid of planktic foraminifers. Two DC grains from these layers were analyzed by DCP-OES. Results indicated these grains have an approximate molar ratio of Mg/Ca 1, indicating a dolomitic mineralogy.

Deposition of IG is relatively low and does not covary with HSG at a millennial-scale. After the MIS 8.5 interstadial, as benthic δ¹⁸O gradually increases, the percentage of IG also increases dramatically. From 280 to 275 ka, glass is virtually absent, rarely exceeding trace percentages. From 275 ka, IG increases to maximum values of > 20% of the lithic fraction just prior to the onset of the DC events at 263 ka. HSG also increases over this interval.
Site U1308 petrologic, planktic foraminifer, and benthic foraminifer data for MIS 8. Stable isotope analyses were performed on Core 1308E-2H at a 2-cm resolution (~ 300 years) while petrologic analyses were performed on Core 1308A-2H at a 1-cm resolution. The cores were aligned with shipboard physical property data.

### 3.6.3 Original Glacial-Stage HSG Records

Here, two original, new HSG records are presented from Site U1308 spanning glacial Stages 6 and 8 in their entirety. The Stage 6 record has 267
points and spans the interval from 127 to 191 ka with a range of 1 to 15% HSG and a mean and standard deviation of 6.3 and 2.5, respectively (Figure 31). The Stage 8 record (238 - 309 ka; N = 467) is statistically similar to the MIS 6 record, with a range of 1 to 17% HSG and a mean and standard deviation of 6.2 and 2.8, respectively (Figure 32).

While the Stage 6 HSG record does not exhibit well-structured repetitive increases, it exhibits one 99% confident spectral peak at a 1/950-year frequency. The Stage 8 HSG record exhibits several 99% confident millennial-band spectral peaks. There is a discrete peak at 1/1450 years and elevated spectral density in a broad band between 1/1180 and 1/990 years. Between ~ 300 ka and the onset of DC events at ~ 263 ka, the Stage 8 HSG record exhibits well-structured cycles with a mean pacing of 1421 ± 461 years, which is indistinguishable from the 1/1450-year spectral peak (Figure 33).

Error for the Site U1308 records is 3 percentage points, the 2σ-range of all duplicate analyses. This differs from the error reporting method for the Site 609 HSG record of MIS 2-4 (Bond et al., 1999), which was calculated by 20 replicate counts on the same sample and resulted in a 2σ-range of ± 2 percentage points (Bond et al., 1997). Reporting a lower error than the original Bond studies (Bond et al., 1997, 1999, 2001) results from the reduced variance of the population of HSG grains deposited at Site 609/U1308 during Stages 6 and 8 relative to that of the last glaciation; the MIS 2 - 4 HSG record has a standard deviation of 5 (Figure 36), twice that of the records reported here.
Figure 31 | MIS 6 HSG Record

A) HSG record for MIS 6. B) Distribution of HSG data. C) Spectral analysis reveals a spectral trough at 1470 years. Statistical significance is based on multi-taper methods (black line), with upper and lower dashed lines representing the 99% and 95% confidence intervals resulting from fitting to a red noise background. The record has one 99%-confident peak at 950 years. 95%-confident peaks are at approximately 1900, 1615 (not labeled in figure), 750 and 600 - 550 years. The spectrum calculated from the more traditional Blackman-Tukey (BT) method is also shown for comparison (gray line). Units of spectral amplitude for the BT spectrum are scaled to those of the MTM spectrum. The BT method typically results in approximately an order of magnitude less power than MTM.
A) HSG record for MIS 8. B) Distribution of HSG data. C) Spectral analysis reveals a spectral peak near 1470 years. Statistical significance is based on MTM (black line), with upper and lower dashed lines representing the 99% and 95% confidence intervals relative to a red noise background. The record has several 99%-confident peaks at approximately 1450 (not labeled), 1180 - 990, 900 - 800, and 625 years. 95%-confident peaks are at approximately 1850 and 725 (not labeled) years. Blackman-Tukey spectrum (gray line) is scaled to MTM units.
The elapsed time between HSG cycle midpoints during between 300 and 263 ka is $1421 \pm 461$ years.

### 3.7 Discussion

#### 3.7.1 Chronological Uncertainty

The resulting sedimentation rates are relatively stable throughout both the Stage 6 and 8 glaciations (Figure 34). The average sedimentation rate for Stage 6 ($\sim 4.5$ cm/k.y.) is lower than that of Stage 8 ($\sim 7$ cm/k.y.). This is consistent with that reported for Site 609 during Stages 2 - 4 ($\sim 5.5$ cm/ky). During Stage 6, there is reason to expect a relatively stable sedimentation rate due to the lack of large fluxes of DC, which are indicative of Heinrich Events and during which sedimentation rates would have increased dramatically (McManus et al., 1998), The need for two tie points at Termination II and resulting increased sedimentation rate is consistent with increased IRD deposition during H11.

The higher sedimentation rate during Stage 8 is also consistent with the identification of three high detrital carbonate (Figure 28b), foraminifer-barren
layers. These are the same characteristics of classic Heinrich Events 1, 2, 4, and 5 of the last glaciation (e.g., Hemming, 2004). However, due to the use of only two tie points for Stage 8, the sedimentation rate remains unchanged throughout the duration of the glaciation. This is inconsistent with the known increased sediment flux during Heinrich Events.

Given that precessional changes in solar insolation were maximized during Stage 8 (due to the high eccentricity associated with the ~ 412 k.y. cycle) alternate stratigraphies may be created by further tying the Site U1308 benthic δ¹⁸O to the LR04 stack at and/or around the MIS 8.5 interstadial. These age models, however, result in a non-intuitive relationship. Sedimentation rates are reduced during the interval containing all Heinrich Events, yet enhanced during the ice-growth phase from MIS 9.1 to MIS 8.6. Therefore, this study opts to use a more parsimonious age model that produces good agreement to LR04 with the minimum number of tie points.

While conducting a millennial-scale paleoceanographic study with an orbital-scale chronology will result in inherent uncertainty, there is simply no other means available at present for creating a suborbital chronology. The well-dated, non-marine records of millennial-resolution than span all or part of glacial Stages 6 and 8 do not provide sub-orbital chronological constraint for Site U1308 (Figure 35).
Site U1308 Linear sedimentation rate (LSR) and age-depth relationship for Stages 6 to 9. Vertical lines indicate LR04 boundaries for MIS 5/6, 6/7, 7/8, and 8/9.
Site U1308 benthic $\delta^{18}$O (Hodell et al., submitted) and the LR04 stack cannot be tied to either the Antarctic Dome Fuji ice core (Kawamura et al., 2007) or Sanbao Cave speleothem (Wang et al., 2008) $\delta^{18}$O records, providing no suborbital chronological constraint.

### 3.7.2 HSG Record of the Last Three Glaciations

Including the original Bond et al. (1999) study, the HSG record at Site 609/U1308 spans three discrete glacial intervals, MIS 2-4, MIS 6, and MIS 8. Rather than creating a new HSG record from Site U1308 sediments, the HSG record of Bond (1999), with an updated chronology (See Chapter 2 “DSDP Site 609 Chronology: Implications for HSG Cyclicity”, Page 18), is considered.

The MIS 2 - 4 HSG record ($N = 302$) (Bond et al., 1999) ranges between 0 and 30% HSG with a mean and standard deviation of 13 and 2, respectively (Figure 36), which is considerably more variable than the Stage 6 and 8 record due primarily to increased HSG amplitude (Figure 37). For the HSG...
methodology reproduction study, only two (out of fourteen) samples contained HSG in excess of 20% (See Chapter 1.5, Page 14). One of these samples was reproduced within 1 percentage point of the Bond result (33%). Multiple attempts to reproduce the result of the other sample (30%) failed. Therefore, it is unclear whether “operator bias” or truly lower-amplitude variability resulted in the lower standard deviation of HSG during the preceding two glaciations.

From 263 ka, the reduced amplitude in HSG variability is likely due to dilution by DC. Prior to this during Stage 8, the pacing of HSG cycles (1421 ± 461 years), though of lower amplitude, resemble that of Stages 2 - 4, which exhibits 1573 ± 620 year cyclicity on the GICC05/ss09sea chronology (See Chapter 2.6, Figure 12, Page 28). The Stages 2 - 4 and 8 HSG records also share spectral similarities. Each record exhibits statistically significant spectral density in similar millennial bands: 1850 - 1800, 1475 - 1415, ~ 1200, and 1000 - 950 years⁻¹ (Figures 32 and 36). During Stage 6, however, deposition of HSG occurred in much less well-defined cycles. The Stage 6 HSG spectra contains one 99% confident spectral peak at a 1/950-year frequency, which is similar to a Holocene HSG record (See Chapter 4.2.3, Page 87). Therefore, given the caveat of increased chronological uncertainty, there appear to be similarities between the pacing of HSG events between MIS 2- 4 and 8.
A) HSG record for MIS 2-4 (Bond et al., 1999) on an updated Greenland ice core chronology based on radiocarbon to 26 ka, GICC05 layer counting to 60 ka, and ss09sea flow model from 60 ka (Chapter 2). B) Distribution of HSG data. C) Spectral analysis reveals a spectral peak near 1470 years (1415 years). MTM reveals other 99%-confident peaks (upper dashed line) are at approximately 1800, 1200, 1000, and 825 years. One 95%-confident peaks (lower dashed line) is at 825 years. Blackman-Tukey spectrum (gray line) is scaled to MTM units.
Benthic *C. wuellerstorfi* $\delta^{18}O$ (Hodell et al., submitted) and HSG for Stages 2 - 4 (Bond et al., 1999) (black; Site 609; GICC05/ss09sea chronology) and Stages 6 (red) and 8 (blue) from Site U1308. All records plotted on same vertical scale, illustrating the reduced amplitude of HSG variability during Stages 6 and 8. Stage 8 HSG cycles occur in bundles: 308 - 295 ka, 295 - 280 ka, 280 - 273 ka, and 273 - 260 ka, after which Stage 8 HSG is further reduced coincident with dramatically increased deposition of detrital carbonate (not shown).
3.7.3 Heinrich Events Prior to the Last Glaciation

Heinrich Events (Heinrich, 1988; Bond et al., 1992) are interpreted to be the result of surging of the Laurentide Ice Sheet (LIS), though the exact mechanism remains unclear (Hemming, 2004). While internal instabilities related to pressure-induced basal melting could cause a large ice sheet such as the LIS to surge once basal frictional forces are overcome (MacAyeal, 1993), this does not explain the evidence suggesting synchronous discharge from multiple ice sheets just preceding H1 through H6 (Bond and Lotti, 1995). It has also been suggested that sea-level rise due to these “precursor events” could have destabilized grounded continental shelf Laurentide ice (Scourse et al., 2000).

The nature and occurrence of Heinrich Events prior to the last glaciation is also not well understood, with only a few studies having been conducted (Hemming, 2004). Split-core scanning XRF analyses at Site U1308 indicate that Heinrich Events began during MIS 16, at the end of the Mid Pleistocene Climate Transition, perhaps as the LIS ice volume and duration surpassed a threshold (Hodell et al., submitted).

3.7.3.1 MIS 6 Heinrich Events

During Termination II at approximately 130 ka, sediment cores throughout the North Atlantic record an episode of ice rafting commonly referred to as Heinrich Event 11 (H11). The petrologic record from Site 609 exhibits marked increases in DC for H1 though H6, including H3 (e.g., Figure 17). However, Site U1308 is devoid of DC during the H11 event, which is
comprised primarily of siliciclastic ice-rafted grains. In addition, while *N. pachyderma* (s) comprises nearly 100% of the planktic foraminifers during H11, planktic foraminifers still accounted for 20% of the total sediment (Figure 38). In contrast planktic foraminifers are virtually devoid from Site 609 sediments during the Heinrich Events of the last glaciation. In the absence of DC, the maximum concentration or IRD (~ 400 grains/g) is an order of magnitude less during MIS 6 relative to MIS 2 - 4.

The petrologic record of IRD deposition during MIS 6 is corroborated by scanning XRF results from the same core (Figure 38). The ratios of Ca to Sr (Ca/Sr) and of Si to Sr (Si/Sr) are proxies for DC and siliciclastic grains, respectively, at Site U1308 (Hodell et al., submitted). The Paleozoic limestone and dolostone of the Canadian Shield are depleted in Sr relative to biogenic carbonate, while the lithic (Si-rich, Sr-poor) to foraminifer ratio (relatively Sr-rich) is reflected in Si/Sr.
Scanning XRF-derived Si/Sr correlates highly with lithic data over the interval of overlapping data \((r = 0.87)\). During the Heinrich Events of the last glaciation, Ca/Sr exceeded 300 and total lithic grain deposition exceeded 4000 grains/g, an order of magnitude higher than during H11 (\(\sim 130\) ka). XRF data from Hodel et al. (submitted).

Given the virtual absence of DC at Site U1308 and no compelling evidence for significant deposition of DC elsewhere in the North Atlantic during H11, the interpretation of H11 as a Heinrich Event is questionable. Only two other studies have published a record of DC for MIS 6. Nearly adjacent to Site U1308, core TP88-9P (48°23.03’N, 25°5.10’W; 3193 mbsl; Figure 22) shows deposition of DC but at low percentage (van Kreveld et al., 1996; Hemming, 2004). Another core from the Feni Drift (ENAM 97-09; 54°32.4’N, 16°21.6’W; 2452 mbsl; Figure 22) contains a high percentage of detrital carbonate within a condensed, winnowed,
likely discontinuous section with an average sedimentation rate of < 1 cm/ky (Richter et al., 2001) from which it is difficult to interpret the significance of the DC content. No DC was identified in MD95-2040 off the Iberian Margin within the MIS-6 interval (de Abreu et al., 2003).

3.7.3.2 MIS 8 Heinrich Events

Three prominent DC events are identified in the 63 - 150 µm fraction during the latter portion of MIS 8 at ~ 263, 249, and 243 ka (Figure 39). The maximum percentage of DC is ~ 20%, which is less than during the last glaciation (e.g., Figure 17). Preliminary examination of the > 150 µm fraction indicates that these intervals have greatly reduced numbers of planktic foraminifers, as during the last glaciation. Benthic foraminifer δ¹³C and planktic foraminifer δ¹⁸O (Hodell et al., submitted) both decrease during these intervals, which is interpreted to indicate decreased NADW production due to increased meltwater input. The number of benthic foraminifers are also reduced such that the benthic δ¹³C has an approximate 3 k.y. gap at ~ 260 ka. This is likely due to an inferred increased sedimentation rates consistent with the last glaciation (McManus et al., 1998). These events are also detected in Ca/Sr (Hodell et al., submitted) (Figure 39). Therefore, these DC events are interpreted to be true Heinrich Events caused by, for at least the oldest two events, surging of the LIS.
Increases in %DC and Ca/Sr (top) corresponds to decreasing planktic and benthic $\delta^{18}O$. XRF data from Hodel et al. (submitted).

These two older Heinrich Events occurred subsequent to an inferred increase in ice volume based on benthic $\delta^{18}O$ (Figure 30), indicating the events began once the LIS surpassed a critical size threshold. The youngest event appears to have occurred quite late during Termination III, and is preceded by an increase in siliclastic IRD as indicated by Si/Sr, perhaps indicating that other ice sheets had begun to collapse prior to the LIS.

Benthic $\delta^{13}C$ corresponds to Si/Sr (Hodell et al., submitted) in MIS 8 (Figure 39) though less so for MIS 6 (e.g., Figure 45, Page 81) This relationship during MIS 8 can be used to infer $\delta^{13}C$ values during the intervals in which specimens of *C. wuellerstorfi* are absent and indicates a large $\delta^{13}C$ decrease during the first Heinrich event. While the onset of these events is abrupt and
synchronous in both the petrologic and XRF data, the petrologic data indicate a much longer persistence of DC deposition, especially for the oldest event at ~ 263 ka. The XRF data reflects bulk sediment composition, while the DC data reflect only a narrow size fraction (63 - 150 µm). Heinrich layers typically exhibit sharp bases and bioturbated tops, so it is expected that the DC signal will be smoothed into the sediments subsequently deposited but unlikely that this only occurred with the 63 - 150 µm fraction (Figure 40). Benthic δ¹³C returns to pre-event values prior to the abrupt decrease in DC. The highest Ca/Sr corresponds to the darkest area of the event.

**Figure 40** | Sediment Core Image of MIS 8 Heinrich Event

Core 1308A-2H digital image showing Heinrich Event beginning at ~ 262 ka with A) %DC and B) Ca/Sr.
3.7.4 Differences Between the Last and Penultimate Glaciations

The Greenland ice core record does not extend beyond the last interglaciation, MIS 5. Differences between the last and penultimate glaciations at Site 609/U1308 allow for inference regarding the nature of the D-O stadial-interstadial cycle. The absence of Heinrich Events during MIS 6 indicates that packaging of progressively cooler interstadials between surges of the Laurentide Ice Sheet (Bond Cycle) did not occur. In fact, the lack of a “1500-year cycle” may indicate that D-O Events would have likely been manifest differently, if at all, during MIS 6. During MIS 2 - 4 the distribution of \( \%N. pachyderma \) (s) data at Site 609 is strongly bimodal, reflecting a predominance of either relatively warm or relatively cold conditions (Figure 41a). The bimodal distribution of \( \%N. pachyderma \) (s) data remains for the last glacial interval after clipping stratigraphic layers corresponding to Heinrich Events 1 to 6 (Figure 42).

However, the distribution of \( \%N. pachyderma \) (s) is normal during the penultimate glaciation, with a single mode between 50% and 70% (Figure 41b). The prevalence of intermediate numbers of \( N. pachyderma \) (s) may reflect the absence of D-O Events and associated large changes in SST. A similar difference existed off the Iberian Margin at MD95-2040 (de Abreu et al., 2003) and MD01-2443,4 (Martrat et al., 2007). These SST records exhibited less severe temperature changes during MIS 6 relative to MIS 2 - 4.

Alternatively, this may indicate an expanded subpolar water mass in general during MIS 6. Based on planktic foraminifer faunal assemblages, Crowley (1981) noted that the North Atlantic Drift was apparently located
farther south during the Stage 6 glacial maximum than during the LGM (Climap Members, 1976) but the polar front was located farther north, resulting in a latitudinally expanded subpolar water mass as indicated by the high abundance of *G. quinqueloba*. This is supported by a reduced SST gradient between cores SU90-08 and MD95-2037, which are separated by 6° of latitude across the inferred location of the last glacial polar front (Calvo et al., 2001). Perhaps this configuration was the case for much of MIS 6.

**Figure 41** | Comparison of MIS 2-4 Proxy SST Variability to MIS 6

<table>
<thead>
<tr>
<th>Site</th>
<th>N. pachderma (s) percent</th>
</tr>
</thead>
</table>
| Site U1308 | N = 162
Min = 15%
Max = 98%
Mean = 58%
Std. Dev. = 21% |
| Site 609   | N = 159
Min = 9%
Max = 99%
Mean = 57%
Std. Dev. = 28% |

A) Histogram of MIS 2 - 4 (Site 609) %*N. pachyderma* (s) data down sampled to a 2-cm interval to correspond to the Site U1308 sampling interval. B) Histogram of MIS 6 (Site U1308) %*N. pachyderma* (s) data.
A) Histogram of MIS 2 - 4 (Site 609) %N. pachyderma (s) data as in Figure 41. B) Site 609 data with stratigraphic interval corresponding to Heinrich Events 1 - 6 removed also exhibits a bimodal structure.

Variability in the concentration of lithic grains per gram of bulk sediment at Site U1308 is greatly reduced during Stage 6 relative to Site 609 during Stages 2 - 4, with an order of magnitude lower concentration during peak ice-rafting events (Figure 43). This is likely related to the lack of Heinrich Events during Stage 6. While the overall concentration of lithic grains is markedly decreased at Site U1308 during Stage 6, the structure of the of the lithic grain population is quite similar during both glaciations, with a single, low-amplitude mode corresponding to minimal concentrations and a long tail comprising ice-rafting events. This is also evident in the distribution of the percentage of lithic grains (Figure 44).
Figure 43 | Comparison of IRD Concentration of MIS 2 - 4 to MIS 6

A) Site 609 lithic grain concentration and B) Site U1308 concentration with axes scaled the same.  C) Inset depicts the Site U1308 data scaled to maximum values.

Figure 44 | Comparison of IRD Variability Between MIS 2 - 4 and 6

Histogram of the percentage of lithic grains for A) Site 609 and B) Site U1308.

Though this site does not directly monitor NADW strength, during the last glaciation, proxy data for lithic grain deposition at Site U1308 was tightly coupled to benthic foraminifer $\delta^{13}C$, indicating shoaling and/or reduced NADW
formation and infiltration of relatively $^{12}\text{C}$-enriched bottom waters into the North Atlantic basin (Hodell et al., submitted). However, this relationship is less evident during MIS 6.

Both total lithic concentration and the percentage of lithic grains track changes in planktic $\delta^{18}\text{O}$ ($r = -0.55$), likely reflecting freshening surface waters due to the melting of $^{16}\text{O}$-enriched icebergs (Figure 45). Benthic $\delta^{13}\text{C}$, however, is more highly correlated with changes in $\% N. \text{pachyderma} (s)$ ($r = -0.3$) than with lithic grain variability ($\% \text{lithic} r = -0.22$, lithics/g $r = -0.17$). This relationship indicates more subtle changes in circulation than during the last glaciation in either the position of the polar front and/or the North Atlantic Drift, which in turn, may reflect changes in the overturning circulation.
Top) very little DC was deposited at Site U1308 during MIS 6. Middle) planktic $\delta^{18}O$ is coupled to variations in lithic grains, while (bottom) the %$N. pachyderma$ (s) SST proxy tracks changes in benthic $\delta^{13}C$. Isotope data from Hodell et al. (submitted).

Increased stability of the Laurentide Ice Sheet (LIS) is inferred for the penultimate glaciation. While the near absence of LIS-derived DC in the distal sedimentary record of Stage 6 could have been the result of decreased iceberg
survivability, this is unlikely based on the basin-wide decoupling of SST records. During MIS 6 North Atlantic SST records do not exhibit synchronous changes as during the last glaciation, and global records exhibit fewer and lower amplitude variations that are limited to the early glaciation (Figure 46).

**Figure 46** Site U1308 SST Compared to Other Locations

Comparison of Site U1308 proxy SST (% *N. pachyderma* (s)) to other records from Antarctica (Dome Fuji air temperature; blue), China (Sanabo Cave summer/winter precipitation ratio; black and green to illustrate overlapping stalagmites), the North Atlantic (M23414 (% *N. pachyderma* (s): brown), and the Iberian Margin (Uk37 SST; green)

A transect of cores across the glacial polar front from 43°N to 53°N, which recorded synchronous SST changes during the last glaciation, exhibited only
subtle, asynchronous changes during MIS 6 (Figure 47). Had the hydrography of the North Atlantic been repeatedly hammered by “armadas” of melting icebergs, these SST records would likely record larger-amplitude, synchronous changes.

**Figure 47** MIS 6 SST Transect Across the Glacial Polar Front

A transect of cores across the glacial polar front includes M23414 (53°N, top) (Kandiano et al., 2004), Site U1308 (50°N, middle) and SU90-08 (43°N, bottom) (Villanueva et al., 1998).

In addition, there is absolutely no evidence that the LIS contributed to the ice-rafting event, typically referred to as H11, at Termination II. Not only has there not been any significant deposition of DC detected and reported in the literature (Hemming, 2004), none was deposited at Site 609/U1308. As a result, the concentration of lithic grains was an order of magnitude lower than for the classic Heinrich Events of the last glaciation.

This transect of cores clarifies frontal boundary movements within Stage 6 (Figure 47). During the glacial maximum, previous work suggests that the polar
front was located north of SU90-08 (Calvo et al., 2001). The dominance of *N. pachyderma* (s) just prior to Termination II and the high lithic content suggest the front was south of Site U1308, placing between 43°N and 50°N.

Just prior to the glacial maximum, intermediate temperatures at 53°N and only ~ 50% *N. pachyderma* (s) at 50°N likely reflects the expanded subpolar water mass noted by Crowley (1981). The lowest SSTs are recorded from 144 to 156 ka at 53°N, during which time the polar front is indicated to have been between 53°N and 48°N (Kandiano et al., 2004). Given these constraints, the front was likely directly over Site U1308 based on the high percent of *N. pachyderma* (s). From ~ 145 ka, the front likely moved farther north, just south of 53°N. There are however, several intervals that are difficult to interpret. At ~ 140, 160, 174 ka, Site U1308 recorded increasing numbers of *N. pachyderma* (s) while M23414 to north recorded relatively warm temperatures.

### 3.8 Conclusions

Marked difference exist between the proxy records of ice-rafting at Site 609/U1308 during the last and penultimate glaciations. During Stage 6, there was no 1500-year cycle in HSG, perhaps indicating that D-O Events were not manifest in a similar fashion, if at all. In addition, the near absence of DC indicates there were no Heinrich Events. The ice-rafting event at Termination II, commonly referred to as H11 also contains no DC. To date, no other studies report significant DC deposition within Stage 6. Therefore, no evidence exists for surging of the Laurentide Ice Sheet during the penultimate glaciation.
During Stages 2 - 4, the effects of D-O and Heinrich Events were apparent on a basin-wide scale, causing synchronous changes in SST. However, in the absence of Heinrich Events, and perhaps D-O Events, the pronounced swings in SST associated with these events during the last glaciation are not detected during the penultimate glaciation. The SST proxy record at Site U1308 during Stage 6 primarily records intermediate temperatures. The subtle SST changes detected likely indicate local as opposed to basin-scale changes related to the migration of oceanic frontal boundaries. During Stage 6, benthic $\delta^{13}$C changes are of lower amplitude than Stages 2 - 4 and correspond more strongly to variations in SST than to ice rafting, indicating that ice-rafting events did not as strongly influence NADW formation.

The Stages 2 - 4 and 8 HSG records share several similarities. Each record contains well-structured cycles in HSG with a mean event spacing of $\sim 1500 \pm 500$ years. These records also exhibit similar spectral power at $\sim 1/1500$-years, though the millennial-band spectral variance is not dominated by this frequency as other spectral peaks are present. The existence of a 1500-year cycle in HSG indicates a greater likelihood of D-O Events during Stage 8. In addition, three Heinrich Events, defined by a large abundance of LIS-derived DC, occurred during MIS 8, indicating surging of the LIS Ice Sheet. Improved methods for correlation of marine sediment and Antarctic sediment cores are required to understand the exact relationship of these ice-rafting events on the northward transport of heat and global climate.
Chapter 4. Significance of Reproducible Millennial Ice-Rafting Signals During the Last Three Glaciations

4.1 Introduction

Bond et al (2001) attributed the “pervasive” 1500-year cycle to solar forcing based on a close match in the Holocene to cosmogenic nuclide production, thought to be the indicate variable solar irradiance. However, the existence of a 1500-year climate cycle in the Holocene is questionable. Supplemental online material and endnotes indicated that the Holocene HSG record exhibits ~ 950-year spectral power and is highly coherent with $^{14}$C production at this frequency (Bond et al., 2001). There is no evidence for solar variability at 1500-years in any of the cosmogenic nuclide records, prompting speculation that a 1500-year period could be produced through combination tones of higher-frequency centennial cycles (Braun et al., 2005; Clemens, 2005).

In addition to cooling events such as that at 8.2 ka (Alley et al., 1997) and the “Little Ice Age” (Matthes, 1939), periodic variations are evident in a large number of climate records during the Holocene from across the globe. Many of these records are well constrained chronologically and include a variety of proxies. In addition to the Holocene HSG records (Bond et al., 1997, 2001), proxies for air temperature (Overpeck, 1987; Schulz and Paul, 2002; Viau et al., 2006), sea surface temperature (deMenocal et al., 2000), lake productivity (Hu et al., 2003), precipitation (Langdon et al., 2003), and deep ocean circulation (Bianchi and McCave, 1999; Chapman and Shackleton, 2000) all appear to exhibit periodic variability. Several of these records apparently vary on a 1500-year time
frame (Bond et al., 1997; Bianchi and McCave, 1999; deMenocal et al., 2000; Bond et al., 2001), but evidence is mounting that a 1000 ± 100 year period of variability is more well developed during the Holocene (Viau et al., 2006).

4.2 Objectives

The objective of this study is to test the hypothesis that HSG ice-rafting events reflect variations in solar irradiance by examining the frequency spectra of the Site 609/U1308 HSG records of the past three glaciations.

4.3 Background

The Holocene record is based on the stacked record from two locations, V29-191 (54°16’N, 16°47’W; 2370 mbsl) and GGC-22 (44°18’N, 46°16’W; 3958 mbsl). Prior to stacking, a multicorer (MC) record from the same site was spliced into the gravity and piston core records to extend the time series to the present (Bond et al., 2001). The resulting record ranges from core top to 11,550 calendar years BP 1950, is interpolated to a 70 year time step, and is comprised of 166 data points. HSG data varies between 1 to 18%, with a mean and standard deviation of 8.5 and 3.7, respectively (Figure 48). While this record was described as indicative of the 1500-year cycle, no interval counts were reported (Bond et al., 2001). In fact, it cannot be ruled out that the original HSG records which indicated the “1500-year cycle” persisted into and during the Holocene (Bond et al., 1997) are correlated with a ~ 900-year cycle in Greenland air temperature (Schulz and Paul, 2002) and therefore do not actually exhibit 1500-year cyclicity at all.
A) Stacked HSG record for the Holocene (MIS 1) from Sites MC52-V29191 and MC21-GGC22, with radiocarbon dates shown in blue and green, respectively (Bond et al., 2001). Data compiled by M.N. Evans and R. Muscheler from Bond’s original data files and comparison with published paper figures (Bond et al., 2008). The original figure, Fig. 2g in Bond et al. (2001), indicated that maximum values do not exceed 15% HSG, which differs from these data. B) Distribution of HSG data. C) MTM Spectral analysis reveals 99%-confident peaks (upper dashed line) are at approximately 970 and 540 years. Blackman-Tukey spectrum (gray line) is scaled to MTM units.
Core NEAP-15K from the Gardar Drift, sited to monitor changes in Iceland-Scotland Overflow Water (ISOW), exhibits periodic grain size (sortable silt) variations at 1500 years (Bianchi and McCave, 1999) but at ~ 1000 and 550 years in sediment lightness (L*, a proxy for calcium carbonate content in North Atlantic sediments) (Chapman and Shackleton, 2000). Because reductions in ISOW flow, as indicated by lower grain-size, are coincident with lightness minima (Chapman and Shackleton, 2000), it is likely that spectral results from one or both of these records are spurious. Reevaluation of the sortable silt record demonstrate that its 1500-year spectral peak is not statistically significant at a 95% level (Schulz and Paul, 2002). Schulz and Paul (2002) also pointed out that, if the Holocene HSG drift-ice record also exhibits a 1500-year cyclicity (Bond et al., 1997), then it should be coherent with the grain-size proxy for NADW flow. However, this is not the case.

4.3.1 Cosmogenic Nuclides

Changes in inferred production rates of long-lived cosmogenic radionuclides has been used by a number of studies to link Holocene climate variability to changes in solar irradiance (Chapman and Shackleton, 2000; deMenocal et al., 2000; Bond et al., 2001; Hodell et al., 2001; Neff et al., 2001; Hu et al., 2003; Viau et al., 2006). Cosmogenic nuclides, including $^{10}\text{Be}$, $^{14}\text{C}$, and $^{36}\text{Cl}$ are produced through the interaction of stable isotopes with galactic cosmic ray (GCR) particles, which consist primarily of protons and $\alpha$-particles, in the upper atmosphere (Lal and Peters, 1967). Accumulation of cosmogenic nuclides in natural archives, such as ice cores and tree rings is a function of transport and
deposition. Thus, inferring a rate of production from the concentration of a particular nuclide requires knowledge regarding its geochemical behavior; \(^{14}\text{C}\) is oxidized to \(\text{CO}_2\) and \(^{10}\text{Be}\) sorbs onto aerosols (Masarik and Beer, 1999). Changes in production may result from changes in GCR flux, as well as the intensity of the Earth’s and Sun’s magnetic fields. While the flux is inferred to be relatively constant (Vogt et al., 1990), the magnetic fields, which deflect GCRs, are not.

### 4.3.1.1 Geomagnetic Intensity Modulation of GCR Flux

The modulation of GCRs due to the Earth’s magnetic field occurs in space, where the shielding effects due to non-dipole fields are weak, and thus relies primarily on the dipole moment (Muscheler et al., 2005). GCR flux to the atmosphere increases during times of low geomagnetic intensity, such as the Laschamp Event (~ 40 ka). Additionally, when intensity is low, modulation of GCR flux is more sensitive to changes in field strength (Wagner et al., 2000). Latitude also affects modulation of GCRs, with negligible geomagnetic effects at the poles; magnetic field lines are parallel to the earth at the equator, providing maximum shielding, but perpendicular at the poles.

Based on archeological work, it has long been assumed that the geomagnetic field intensity only exhibits low-frequency variability (McElhinny and Senanayake, 1982; Yang et al., 2000; Teanby and Gubbins, 2002) and can therefore be neglected during interpretation of cosmogenic nuclide production records, allowing sole attribution to solar changes (Bard et al., 1997; van Geel et al., 1999; Bond et al., 2001; Hu et al., 2003; Viau et al., 2006). However, relative paleomagnetic intensity (RPI) records from high-sedimentation rate Pleistocene
sediment cores suggest global variability on both millennial (Stoner et al., 2000) and centennial time scales, with spectral coherence between $^{14}$C production and RPI (St-Onge et al., 2003). The coherency may result from climate contamination of the RPI record due to incomplete removal of lithological signal. It is currently believed to be unlikely that geomagnetic intensity changes occur on time scales less than 500 years (Stoner, 2008). However, due to limitations in the sedimentary records, it is currently not possible to definitively remove and/or account for the affects of changing paleointensity on records of cosmogenic nuclide production (Snowball and Muscheler, 2007).

4.3.1.2 Solar Modulation of GCR Flux

Changes in solar activity, as indicated by changes in faculae and sunspots numbers, results in variations in the strength of the Sun’s magnetic field. Recent space-based observations confirm that the solar constant, the total amount of incoming solar electromagnetic radiation on a surface perpendicular to the sun, varied with solar activity over the last 11-year Schwabe sunspot cycle by about 1 W/m$^2$ (Fröhlich and Lean, 2004; Lean, 2005), which directly resulted in a surface temperature change of $\sim$ 0.1 K (Douglass and Clader, 2002; Gleisner and Thejll, 2003). Model results suggest that irradiance forcing will result in a direct response in parameters such as surface air temperature and monsoon precipitation, as well as a feedback-modified ocean response (Weber et al., 2004). In addition, it has been shown the the circum-subpolar North Atlantic region my also be influenced by solar UV-induced changes in stratosphere-troposphere coupling on the North Atlantic Oscillation (Shindell et al., 2001). However, a
physical mechanism to connect solar activity and irradiance changes remains unidentified (Foukal et al., 2006). It also should be noted that, due to the short duration of observational irradiance data, the possibility that a longer irradiance cycle is superimposed on the 11-year activity cycle but has been aliased cannot be ruled out (Foukal et al., 2004).

Thus, the modulation on GCR flux to the stratosphere can be modeled as a function of geo- and solar-magnetic field strength (Figure 49), though the links to solar irradiance changes remains tentative, due to both uncertainty in paleomagnetic intensity records and an unclear link between solar activity and irradiance. However, the best available data indicates variable solar output modulates climate on at least decadal to centennial scales.
Production of $^{14}$C as a function of solar modulation and geomagnetic field intensity, with maximum production occurring as each field approaches zero. Sensitivity to changes in combined field strength is increased when during low intensities. After (Beer, 2000; Snowball and Muscheler, 2007).

**4.3.1.3 Carbon-14**

Neutrons created during GCR particle collisions with the Earth’s atmosphere subsequently collide with stable nitrogen atoms ($^{14}$N), knocking out a proton, resulting in an unstable $^{14}$C isotope with a half life of 5370 years. $^{14}$C is oxidized to $^{14}$CO$_2$ and enters the carbon cycle, greatly complicating interpretations based on the ratio of $^{14}$C/$^{12}$C. An ocean reservoir effect (~ 400 years) results from radioactive decay; the ratio of $^{14}$C/$^{12}$C in dissolved carbon
species in ocean water is always less than that of atmospheric CO$_2$. Due to changes in the past atmospheric $^{14}$C/$^{12}$C ratio and changing partitioning between active reservoirs, “radiocarbon” years must be calibrated to calendar years (e.g, INTCAL04, Reimer et al., 2004).

The current, most well dated calibration database is the terrestrial component of INTCAL04, which is based on annually-resolved tree-ring chronologies to 12,400 calendar years BP (Reimer et al., 2004). Circulation and carbon cycle changes during the Younger Dryas (YD) introduced uncertainty in the calibration beyond ~ 11,500 calendar years BP. In fact, the onset, duration, and termination of the YD is not well constrained due to this, though “floating” tree ring chronologies that are not yet connected to the continuous, absolutely dated record appear useful in resolving this uncertainty (Muscheler et al., 2008).

After accounting for these effects, carbon-14 is described as $\Delta^{14}$C, which is the measured $^{14}$C activity ($^{14}$C/$^{12}$C ratio) corrected for decay and isotopic fractionation relative to a standard (Stuiver and Braziunas, 1993). Production rates are then calculated from residual $\Delta^{14}$C (low-frequency trend subtracted) and a carbon cycle model. Production rates based on the INTCAL98 data set (Stuiver et al., 1998), were further processed and smoothed (Figure 50) to compare with paleoclimate proxies, such as HSG (Bond et al., 2001).
The smoothed and detrended $^{14}$C production rate (Bond et al., 2001) (blue) is based on INTCAL98 and has been used in several studies (Hu et al., 2003; Viau et al., 2006). Production rates based on INTCAL04 (Muscheler et al., 2005) (red) are consistent with the detrended rate calculated from INTCAL98 (Bond et al., 2001) (black).

4.3.1.4 Beryllium-10

Beryllium-10 is formed in the upper atmosphere due to spallation of nitrogen and oxygen atoms and has a half life of $1.5 \times 10^6$ years. It adsorbs onto aerosols, which have a 1 to 2 year stratospheric residence time (Masarik and Beer, 1999), and is eventually deposited due to dry deposition and precipitation scavenging. While most $^{10}$Be (and $^{14}$C) is formed in the high latitudes (due to the geometry of the earth’s magnetic field lines), the relatively long stratospheric residence time of aerosols should result in a well-mixed, global pattern, though circulations change could perhaps introduce a spatial bias to the pattern of $^{10}$Be deposition. Modeling suggest that under current climatic conditions, $^{10}$Be
deposition at polar latitudes would increase 80% of the global rate due to a modest decrease in geomagnetic field intensity (Field et al., 2006).

The best dated, high-resolution $^{10}$Be records are available from ice cores (Yiou et al., 1985; Finkel and Nishiizumi, 1997; Yiou et al., 1997; Raisbeck et al., 2006), though marine sediment core records also exist (Christl et al., 2007). Climatic influences these records. In a purely dry deposition case, $^{10}$Be concentration in ice would exhibit an inverse relationship with ice accumulation rate, as is the case in Antarctica at both Vostok (Yiou et al., 1985) and Dome C (Raisbeck et al., 2006). At Summit, Greenland, $^{10}$Be flux is calculated based on the assumption of both wet and dry processes using a linear regression model between the wet and dry end member points (Alley et al., 1995), which produces results in agreement with $^{14}$C production rates (Finkel and Nishiizumi, 1997). This calculated flux is not correlated with $\delta^{18}$O, indicating no strong climate overprint on the delivery of aerosol-bound $^{10}$Be to Greenland (Muscheler et al., 2000).

The $^{10}$Be record of Bond (2001) was based primarily on GISP2, with spliced in GRIP data, but did not include the last ~ 3000 years of the Holocene (Figure 51). A newer composite record, based primarily on GRIP, extends to within ~ 300 years of the present and also agrees well with inferred $^{14}$C production (Muscheler et al., 2005) (Figure 52).
The \(^{10}\)Be flux (Bond et al., 2001) has been extended to near present and with combined GRIP and GISP2 data (Muscheler et al., 2005).

Calculated production rates for both \(^{10}\)Be flux (black) \(^{14}\)C production (red) agree though the Holocene (Muscheler et al., 2005).
4.3.2 Indications of Solar Forcing

Bond et al. (2001) used smoothed and filtered records to illustrate a correlation in time between HSG and cosmogenic nuclides ($^{10}$Be $r^2 = 0.31$) ($^{14}$C $r^2 = 0.19$), thus attributing the “1500-year cycle” to solar forcing (Figure 53). As mentioned in the text of the Bond et al (2001) manuscript, the correlation is between centennial scale cycles in both records that are shown in supplementary online information and notes to exhibit a 950- and 550-year periodicity (Bond et al., 2001) (See also Figure 48, Page 88). Here, this correlation is reconsidered in the frequency domain using the most recently available nuclide production records with no additional processing (Muscheler et al., 2005) (Figure 54).

**Figure 53** | Time-Domain Correlation of HSG with Cosmogenic Nuclides

Unsmoothed stacked HSG (MC52-V29191+MC21-GGC22) compared with cosmogenic nuclide records (Bond et al., 2001). $^{10}$Be $r^2 = 0.31$; $^{14}$C $r^2 = 0.19$. 
The high coherence (99% confidence of non-zero coherence) (Figure 54a) and approximate in-phase relationship between HSG and INTCAL04-derived $^{14}$C production, as well as with $^{10}$Be flux (Figure 54b), argues for a primary statistical link at 950 years. Only the $^{14}$C production records exhibits relatively high spectral density at ~ 550 years, at which none of these records are highly coherent, which indicates the 950-year and 550-year fluctuations may be the result of different processes. There is no evidence for 950-year fluctuations in paleomagnetic intensity records (St-Onge et al., 2003; Snowball and Muscheler,
Since there is no climate correlation of $\delta^{18}O$ with $^{10}$Be flux in Greenland ice (Muscheler et al., 2000), it is unlikely that ocean circulation changes result in the 950-year period, though this is close to the characteristic time scale of ocean overturning circulation.

Though not definitive, this provides circumstantial evidence linking the 950-year period to solar forcing. As the 550-year fluctuation is most strongly expressed in the $^{14}$C production record, it may more likely be of an internal origin, perhaps due to changes in the carbon cycle related to ocean circulation (Stuiver and Braziunas, 1993). However, since these relationships are, thus far, identified in only one discrete climatic interval, the possibility that they arise simply by random chance cannot be excluded. Therefore, the HSG record of the past three glaciation at the same location, Site 609/U1308, are considered. Similar relationships in four discreet climatic intervals would indicate that these observations are very unlikely to be simple, stochastic flukes.

### 4.4 Potential Solar Forcing of Millennial Climate Change During the Last Three Glaciations

The Greenland $^{10}$Be flux record is now extended to 60 ka (Muscheler et al., 2005) (Figure 55), allowing comparison with the the Site 609 HSG record (Bond et al., 1999) updated with the NGRIP GICC05 chronology (See Chapter 2, “DSDP Site 609 Chronology: Implications for HSG Cyclicity”, Page 18). Bandpass filtering around a 1/950-year frequency reveals a relatively constant phasing and similar amplitude modulation (Figure 56). Although phasing is constant, the records are $\sim 180^\circ$ out of phase; offsetting the GRIP $^{10}$Be flux record by 500 years results in a visual correspondence between 15 and 60 ka. The original Site 609
chronology based on correlation of SST to GISP2 δ¹⁸O does not exhibit a similar correspondence (not shown); it neither exhibits similar amplitude modulation nor a consistent phasing relationship with ¹⁰Be flux.

**Figure 55** | Greenland Be-10 Flux for the Last 60,000 Years

Combined GRIP and GISP2 ¹⁰Be flux for the past 60 ka plotted on GRIP ss09sea chronology (Muscheler et al., 2005). Peak production at ~ 40 ka corresponds to the Laschamp paleointensity minimum.
Figure 56 | MIS 2 - 4 HSG and Be-10 Flux Bandpass-Filtered at 950 Years

Combined GRIP and GISP2 $^{10}$Be flux (Muscheler et al., 2005) and MIS 2 -4 HSG (Bond et al., 1999) on GICC05 chronology bandpass filtered at a 1/0.95 ky frequency and 0.02 ky$^{-1}$ bandwidth. GRIP data are offset by 500 years.

Inspection of HSG spectra for the last three glaciations indicates millennial-band periods in Stages 2 - 4 and 8 at ~ 1180 - 1200 and ~ 1415 -1450 years, relative to an AR(1) (red noise) background, but none at these bands in Stage 6 or the Holocene (Figure 57). Spectral analysis is performed using the SSA-MTM Toolkit (Allen et al., 2008), a standard tool in the field of paleoceanography and paleoclimatology. The Holocene contains a broad peak from ~ 1/1100 to 1/900 years. The Stage 8 data therefore indicate that the ~ 1450 and ~ 1200-year oscillations may be characteristic time scales of some, but not all, glacial oscillations. A result of at least equal importance is the presence of a significant 950 ± 50 year period in the Stages 2-4, 6, and 8 HSG records, the presence of which is not limited to Site 609/U1308. This band is also present within the Holocene stacked HSG record, NGRIP δ$^{18}$O during both the Holocene
and last glaciation, and numerous other Holocene records (See “Introduction”, Page 86).

**Figure 57 | Reproducible Ice-Rafting Frequencies to 300 ka**

A) HSG MTM power spectrum (3 tapers, 0 - 1/500 y frequency) (red) for, from top to bottom, the Holocene (0 - 11.5 ka, 70 yΔt), MIS 6 (130 - 190 ka, 240 yΔt), MIS 8 (243 - 300 ka, 150 yΔt), MIS 2 - 4 (15 - 70 ka, 190 yΔt), and for NGIP δ¹⁸O (blue; 14 - 71 ka, 50yΔt; GICC05 chronology to 60 ka). Three 1000-year wide bands are highlighted, 1500 - 1400 (green), 1200 - 1100 (blue), and 1000 - 900 (yellow). Upper and lower dashed lines for each spectrum indicates 99% and 95% confidence lines, respectively. B) Blackman-Tukey cross-spectral analysis (52 lags, 0.14 bandwidth, 99% confidence of non-zero confidence > 0.61) of MIS 2 - 4 HSG (updated NGRIP chronology of Chapter 2 (Page 18)) and GRIP (ss09sea chronology) between 15 and 60 ka. Vertical dashed line is at 1/950 year frequency. Other bands of high coherence include 1/1365 and 1/740 years.
As in the Holocene, HSG is coherent above a 99% confidence level with $^{10}$Be flux at a frequency of 1/950 years during MIS 2 - 4 (Figure 57), though greater chronological uncertainty makes the statistical link less powerful than that of the Holocene. Though chronological uncertainty is even greater for Stages 6 and 8, HSG records for these glaciations also contain spectral power at ~950 years. The reduced spectral amplitude of the 1/950-year peak during MIS 2 - 4 and 8 may be the result of the manifestation of a 1500-year cycle, detected by both spectral and interval counting methods.

No means is currently available for providing suborbital chronological constraint for Stage 6 and 8. Therefore, more advanced methods (e.g., Monte Carlo singular spectrum analysis) are required to further evaluate the significance of these spectral frequencies. Uncertainty in paleomagnetic reconstructions also hinder the separation of geomagnetic and solar effects on nuclide production. As such, results of this study cannot unequivocally refute the null hypothesis that the HSG proxy does not reflect variable solar irradiance. Solar forcing does however provide an explanation for climate variability in the 950-year band during the last three glaciations. The ~1450- or ~1200-year oscillations may still reflect some interactions of shorter period solar cycles with the climate system (Braun et al., 2005; Clemens, 2005).

4.5 Conclusions

The Holocene HSG record of Bond et al. (2001) is strongly linked to cosmogenic nuclide production at a 1/950-year frequency. This same 950-year spectral peak represents the only 99% confident frequency within the Stage 6
HSG record. Stages 2-4 and 8 also contain significant spectral power at this frequency. Therefore, solar forcing provides a potential, though not unequivocal explanation for the ~ 950-year period. The other repeatable oscillations (~ 1450 and ~ 1200 years) are also not inconsistent with a solar origin.
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Biography

Stephen Phillip Obrochta was born November 21, 1973 in Ft. Worth, Texas. He graduated from Booker T. Washington High School in Tulsa, Oklahoma as a National Merit Scholar Semifinalist (without ever knowing he was in the competition) in 1992 and then preceded to work as an auto mechanic. Stephen became interested in marine science after working as a divemaster, deckhand, and scuba instructor in south Florida. After graduating with high honors from Tulsa Junior College in 1995 with an Associate of Arts, he returned to Florida, earned a Bachelor of Science in marine science with honors from Eckerd College in 1997, continued to teach scuba diving, and eventually became distracted and moved to Japan where he taught high school English in a rural fishing village. After spending two years trading English lessons with the staff of a local dive shop in exchange for the opportunity to dive whenever the vessel had an open spot, as well as actually working at the local high school, Stephen returned to Florida. He then married Mariko Sakamoto, who provided endless support while he worked towards a Master of Science in geological oceanography at the University of South Florida College of Marine Science.

Stephen became involved in scientific ocean drilling during his tenure at USF, working on Ocean Drilling Program Leg 194. In 2004, after completing a thesis that employed a sedimentological approach to constrain the geologic age of the Australian Great Barrier Reef, Stephen matriculated at Duke University, forsook the warmth of lower latitudes, and sailed in the northern North Atlantic Ocean and Labrador Sea as a sedimentologist aboard the R/V JOIDES Resolution.
during Integrated Ocean Drilling Program Expedition 303, a product of which is this dissertation. After graduation, Stephen promptly took a post doctorate position at the University of Tokyo Ocean Research Institute.

During his undergraduate and graduate career, Stephen received several awards, though none too amazing. While at Eckerd College, he received an Eckerd College Honors Scholarship, several Howard Hughes Medical Institute undergraduate research grants, an Alyesworth Foundation for the Advancement of Marine Science scholarship, and the honorable mention for the best poster presentation at the Society for Sedimentary Geology Annual Meeting in 1998. Stephen then received two fellowships from the University of South Florida. As a doctoral candidate, Stephen received competitive awards for research abroad from the Duke University Asian Pacific Studies Institute, as well as the Graduate School. He was also presented an American Geophysical Union Outstanding Student Paper Award after the 2006 Fall Meeting for his presentation entitled “Was there a 1.5-k.y. cycle in hematite stained grains during the penultimate glaciation (MIS 6)?”. Finally, in 2007, Stephen was selected as a National Science Foundation East Asian Pacific Summer Institutes Fellow and spent what was supposed to be part of the summer but turned out to be half the year working on this dissertation at Akita University in northern Japan where he learned respect and appreciation for the Siberian High.