AN ABSTRACT OF THE DISSERTATION OF

Aaron M. Barth for the degree of Doctor of Philosophy in Geology presented on June 2, 2016.

Title: Geochemical and Geostatistical Analyses of Quaternary Climate Variability over Millennial-to-Orbital Timescales.

Abstract approved:

______________________________________________________

Peter U. Clark

The goal of dissertation research was to use geochemical, statistical and geological methods to constrain and understand climate variability over several different time scales. Specifically, I have addressed three questions regarding past climate change: (1) how does the record of Irish cirque glaciers constrain the dimensions of the Irish Ice Sheet during and since the Last Glacial Maximum (LGM); (2) what is the record of millennial-scale glacier variability in Ireland during the last glaciation; and (3) how did variability of various components of the climate system interact to contribute to the evolution of climate over the last 800,000 years.

The first chapter involves constraining the vertical and spatial extent of the Irish Ice Sheet (IIS). Reconstructions of the LGM IIS are widely debated, in large part due to limited age constraints on former ice margins and due to uncertainties in the origin of the trimlines used to identify vertical ice limits. The greatest differences exist in southwestern Ireland where reconstructions either have complete coverage by a contiguous IIS that extends onto the continental shelf or a separate, southern-sourced Kerry-Cork Ice Cap (KCIC) with more limited spatial and vertical extent. New $^{10}$Be surface exposure ages from two moraines in a cirque basin in this region provide a unique constraint on ice thickness for this region insofar as the presence of a cirque glacier at a given time clearly indicates that the site was not covered by the IIS. My new $^{10}$Be ages from these two moraines show that the central mountains in
southwestern Ireland were not covered by the IIS or a KCIC since at least 24.5±1.4 ka, thus supporting the more-limited reconstructions of the IIS at the LGM, indicating a reduced contribution to sea-level change and a smaller loading of the solid Earth, which is consistent with models of glacial isostatic adjustment to the IIS.

The second chapter presents research that has developed a record of millennial-scale variability in former Irish cirque glaciers between ~25 ka and 10 ka. Small alpine glaciers are sensitive to climate, and the paleo record of past small-glacier fluctuations offers an outstanding opportunity to use this glacier sensitivity for developing centennial- to millennial-scale records of climate variability. Because Ireland is immediately adjacent to, and downwind of, the North Atlantic, glacial records there are ideally located to record past climate changes associated with changes in North Atlantic Deep Water (NADW) formation and attendant feedbacks. I have developed a high-precision $^{10}$Be surface-exposure chronology of multiple moraines deposited by glaciers in eight cirque basins across Ireland to constraint this variability. The data show a remarkable record of persistent millennial-scale variability between 24.5±1.4 ka and 10.8±0.7 ka. Several of these events are associated with known climatic events during the last deglaciation such as onset of the Bølling-Allerød and end of the Younger Dryas. However, this persistent signal extends back to the Last Glacial Maximum (LGM), suggesting a previously unidentified mode of climate variability unrelated to large changes in the NADW. Multi-decadal to multi-centennial variability identified in Greenland ice cores present a mechanism for the variability recorded in the Irish glaciers.

The third chapter of this research involves characterizing and explaining climate variability at orbital timescales across the mid-Brunhes Transition (MBT; ~430 ka). The MBT involved a change in the amplitude of variability associated with cooler interglacials prior to 430 ka and warmer interglacials after. The key questions I address include determining whether other components of the climate system changed at this time, and identifying the mechanism for the MBT. Statistical tests of multiple proxies (sea-surface temperature, $\delta^{18}$O, $\delta^{13}$C, CO$_2$, CH$_4$, and dust) indicate that the MBT was largely a reorganization of the global climate system perhaps driven by an increase in interglacial CO$_2$ concentrations. Changes in marine $\delta^{13}$C may provide
insights into this change in the carbon cycle, perhaps associated with changes in global ocean circulation. In particular, there is a large positive $\partial^{13}C$ excursion during the interglacial immediately prior to the MBT, suggesting an enrichment of the Atlantic basin at this time relative to other interglacials of the past 800 kyr. Variability in the depth gradient of $\partial^{13}C$ from the North Atlantic show increased correlation with South Atlantic $\partial^{13}C$ records after the MBT, indicating more homogenous mixing of the northern- and southern-component water masses. This is accompanied by a greater difference in the $\partial^{13}C$ latitudinal gradient in the Atlantic basin prior to the MBT that is reduced afterwards.
Geochronological and Geostatistical Analyses of Quaternary Climate Variability over Millennial-to-Orbital Timescales

by
Aaron M. Barth

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Major Professor, representing Geology

Dean of the College of Earth, Ocean, and Atmospheric Sciences

Dean of the Graduate School

I understand that my dissertation will become part of the permanent collection of Oregon State University libraries. My signature below authorizes release of my dissertation to any reader upon request.

Aaron M. Barth, Author
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I have been very fortunate with the people who surround me. Getting to this point is not something I completed on my own, and I am very grateful for the people who guided and encouraged me.

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CONTRIBUTION OF AUTHORS

Chapter 2 – P.U. Clark co-wrote the manuscript, assisted with fieldwork, and conceived of the project. J. Clark conceived of the project and assisted with fieldwork. A.M. McCabe conceived of the project and assisted with fieldwork. M. Caffee oversaw the AMS procedures and measurements.

Chapter 3 – P.U. Clark co-wrote the manuscript, assisted with fieldwork, and conceived of the project. J. Clark conceived of the project and assisted with fieldwork. A.M. McCabe conceived of the project and assisted with fieldwork. J.K. Cuzzone and S.A. Marcott assisted with the chemical procedures. P. Dunlop assisted with field work. M. Caffee oversaw the AMS procedures and measurements.

Chapter 4 – P.U. Clark co-wrote the manuscript and conceived of the project. N.S. Bill assisted with data analysis. F. He assisted with data analysis. N. Pisias helped developed the statistical methods and assisted with data analysis.
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1 Geochemical and geostatistical analyses of Quaternary climate variability over millennial-to-orbital timescales

Chapter 1

Introduction

1.1 Forward

This dissertation represents a multidisciplinary effort that combines geology, geostatistics, and geochemistry to identify and characterize climate variability over the last 800,000 years. This research specifically addresses key objectives in paleoclimate research regarding geological records of climate system variability to provide insights into the mechanisms and rate of change that characterized Earth’s past climate. Through a coordination of efforts on cosmogenic data acquisition, data analysis, and geochemical proxy assessment, this research is able to address climate variability over millennial-to-orbital timescales. By examining the climate system on multiple timescales, this research addresses the response of key components of the Earth system to known forcings and identifies new aspects of climate change that were unforced.

1.2 Project objectives

In Chapter 2, vertical and spatial constraints are put on the dimensions of the Irish Ice Sheet (IIS) during the Last Glacial Maximum (LGM). The LGM configuration of the IIS remains widely debated. Differing methods used to constrain the LGM IIS extent, particularly the thickness, have led to large uncertainties in the reconstructions. Three reconstructions of the LGM IIS can be identified: (1) a minimal version with restricted, largely terrestrial margins and a separate Kerry-Cork Ice Cap (KCIC) (Ballantyne et al., 2011; Bowen et al., 1986; McCabe, 1987), (2) an intermediate version which is also relatively thin and with a separate KCIC but with more extensive marine margins (Brooks et al., 2008; Shennan et al., 2006), and (3) a
maximum version with complete coverage of the landscape and margins that extended far onto the continental shelf (Clark et al., 2012; Greenwood and Clark, 2009; Sejrup et al., 2005). The greatest differences between these reconstructions exist in southwestern Ireland where the maximum reconstructions show complete coverage by a northern-sourced IIS, and the intermediate and minimal reconstructions suggest a separate southern-sourced KCIC. New $^{10}$Be ages from two moraines in a cirque basin found in MacGillycuddy’s Reeks show that the central mountains in southwestern Ireland were not covered by the IIS or a KCIC since at least $24.5\pm1.4$ ka, thus supporting the more-limited reconstructions of the IIS at the LGM. This result has implications for the IIS with respect to its smaller contribution to sea-level change and smaller loading for glacial isostatic adjustment models.

In Chapter 3, a record of millennial-scale deglaciation from cirque basins in Ireland is presented as a means to address the abrupt climate changes experienced in the North Atlantic region during the last deglaciation. Because of its location immediately adjacent to, and downwind of, North Atlantic Deep Water (NADW) formation sites, Irish cirque glaciers present an ideal opportunity to constrain high-frequency climate variability in the region (Liu et al., 2009). A $^{10}$Be chronology of 80 samples compiled from 15 moraines found in eight cirque basins across mostly western Ireland is developed that represents a composite record of cirque glaciation. This chronology demonstrates persistent millennial-scale cirque glacier fluctuations between $24.5\pm1.4$ ka to $10.8\pm0.7$ ka. We find that some fraction of the signal can be explained by a forced response from variability in the Atlantic Meridional Overturning Circulation (AMOC) and its associated changes in heat transport, which would have affected the climate of Ireland (Liu et al., 2009). In contrast, many of the moraine ages occur during a period of reduced AMOC and cooler temperatures. We attribute this signal of variability during a time of reduced climate forcing as being driven by increased climate variability that has been shown to drive kilometer-scale changes in glacier length (Roe and O’neal, 2009).

In Chapter 4, a global distribution of multiple paleoclimate records are analyzed using geostatistical methods to address the Mid-Brunhes Transition (~430 ka; MBT), which is a proposed increase in glacial-interglacial cycle amplitude
(Jansen et al., 1986). Records of sea-surface temperature (SST), benthic carbon isotopes ($\delta^{13}C$), and dust accumulation are analyzed using empirical orthogonal function analysis to objectively characterize the dominant modes of variability over the last 800,000 years. Variance calculations confirm a significant increase across the MBT in several climate proxies, consistent with previous work identifying the MBT. Furthermore, these tests demonstrate that the MBT was a global event. Principal components of $\delta^{13}C$ exhibit a significant carbon isotope excursion during MIS 13 (~500 ka) that has been attributed to a build up of terrestrial biomass from increased Northern Hemisphere summer Asian monsoon strength and precipitation (Guo et al., 2009). We suggest that a reduction in ice volume and a stronger Asian summer monsoon during the preceding MIS 14 may also have contributed to this increase in biomass (Shakun et al., 2015). Additionally, regional records of $\delta^{13}C$ identify changes in the dominant water masses of the paleo-Atlantic Ocean that may have implications for dissolved inorganic carbon storage and atmospheric CO$_2$ concentrations. These characterizations suggest that the MBT may not have been a singular event, but instead involved a sequence of events that transitioned to a different mode of the climate system.

1.3 References


Chapter 2

Last Glacial Maximum cirque glaciation in Ireland and implications for reconstructions of the Irish Ice Sheet

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Abstract

Reconstructions of the extent and height of the Irish Ice Sheet (IIS) during the Last Glacial Maximum (LGM, ~19-26 ka) are widely debated, in large part due to limited age constraints on former ice margins and due to uncertainties in the origin of the trimlines. A key area is southwestern Ireland, where various LGM reconstructions range from complete coverage by a contiguous IIS that extends to the continental shelf edge to a separate, more restricted southern-sourced Kerry-Cork Ice Cap (KCIC). We present new $^{10}$Be surface exposure ages from two moraines in a cirque basin in the Macgillycuddy’s Reeks that provide a unique and unequivocal constraint on ice thickness for this region. Nine $^{10}$Be ages from an outer moraine yield a mean age of 24.5±1.4 ka while six ages from an inner moraine yield a mean age of 20.4±1.2 ka. These ages show that the northern flanks of the Macgillycuddy’s Reeks were not covered by the IIS or a KCIC since at least 24.5±1.4 ka. If there was more extensive ice coverage over the Macgillycuddy’s Reeks during the LGM, it occurred prior to our oldest ages.

Introduction

Establishing the impact of an ice sheet on regional and global climate (Hostetler et al., 2000; Braconnot et al., 2012) and sea level (Lambeck et al., 2014; Peltier et al., 2015) requires an accurate reconstruction of the extent and the height (or thickness) of the ice sheet. Reconstructing the extent of a former ice margin is generally straightforward in terrestrial environments but more difficult where the margins extended offshore, particularly with respect to establishing their age (Weber et al., 2011; Hillenbrand et al., 2014). Reconstructing the height of former ice sheets is more challenging. Direct evidence comes from trimlines, formed where underlying topography projected above the ice surface as nunataks, but their interpretation is often controversial (Carlson and Clark, 2012). Modeling the response of the solid Earth to ice-loading history so as to match relative sea-level (RSL) histories or GPS data is commonly used to estimate ice thickness (Lambeck et al., 2014; Peltier et al., 2015), but the absence of constraining RSL data in many regions, the short length of
most RSL records relative to the total deglacial history, and limited constraints on the properties of the solid Earth that influence isostatic rebound introduce uncertainties in these reconstructions.

At the Last Glacial Maximum (LGM, 19 – 26 ka) (Clark et al., 2009), Ireland was predominantly covered by the Irish Ice Sheet (IIS), which was among the smallest of the LGM ice sheets. Despite its small size, an accurate reconstruction of the LGM IIS is critical for characterizing its response to climate forcing, particularly the abrupt changes in the North Atlantic (J. Clark et al., 2012). The LGM configuration of the IIS, however, remains widely debated. To a large extent, this debate reflects the differing methods used to reconstruct the IIS and their associated uncertainties, making it a microcosm of the issues involved in reconstructing other, larger ice sheets. This includes uncertainties in establishing the extent of the LGM margin on the continental shelf (Greenwood and Clark, 2009; C. Clark et al., 2012) and in determining its thickness from trimline data (Rae et al., 2004; Ballantyne et al., 2011) and from modeling of post-glacial rebound (Lambeck, 1996; Shennan et al., 2006; Brooks et al., 2008). In general, however, we can identify three reconstructions of the LGM IIS (Fig. 2.1): (1) a minimal version with restricted, largely terrestrial margins and a separate KCIC (Bowen, 1986; McCabe, 1987; Ballantyne et al., 2011), (2) an intermediate version which is also relatively thin and with a separate KCIC but has more extensive marine margins (Shennan et al., 2006; Brooks et al., 2008), and (3) a maximum version with complete coverage of the landscape and margins that extended far onto the continental shelf (Sejrup et al., 2005; Greenwood and Clark, 2009; C. Clark et al., 2012). All these reconstructions agree in showing central and northern Ireland being completely covered by the IIS; the main differences are in the extent of the western and southern margins and whether southwestern Ireland was the center of an independent ice cap (the Kerry-Cork Ice Cap, or KCIC) or was completely overridden by a contiguous IIS (Fig. 2.1).

Here we present new $^{10}$Be surface exposure ages from cirque moraines in southwestern Ireland. The existence of cirque moraines indicates that they have not been covered by an ice sheet since their formation. Determining their age places a limiting minimum age on when an ice sheet could have last covered the mountains,
Figure 2.1. Last Glacial Maximum ice sheet reconstructions from A) Bowen et al. (1986), B) Ballantyne et al. (2011), C) Brooks et al. (2008) (for 21 ka), and D) Greenwood and Clark (2009) (for 24 ka) showing varying interpretations of the IIS extent. White shaded areas indicate ice cover. Thick black lines represent ice divides. Thin black lines represent ice thickness contours – numbers indicate topography corrected ice-sheet thickness. Gray lines represent proposed ice flow lines. The location of the Alohart cirque is indicated by the circle on each map.
thus helping to distinguish among the various reconstructions for this region (Fig. 2.1). In particular, these ages provide a unique and unequivocal constraint on ice-sheet thickness in this mountainous region, address the issue of existence of a separate KCIC and the possibility for ice extent offshore, and finally constrain the dimensions of the IIS.

2.3 Setting
The Macgillycuddy’s Reeks (51°58’ – 52°03’N; 09°50’ – 09°34’W) are located on the Iveragh Peninsula in southwestern Ireland. This region exhibits classic alpine glacial topography, and includes the tallest peak in Ireland (Carrauntoohil at 1030 m). According to the minimal reconstructions, this mountain range is near the confluence of the northern-sourced IIS and the southern KCIC, whereas according to the maximum reconstruction, it was completely overridden by a southwest-flowing IIS, thus providing a strategic location for addressing the proposed reconstructions (Fig. 2.1).

The northern slopes of the Macgillycuddy’s Reeks contain multiple cirque basins formed by local glaciation (Anderson et al., 1998). Harrison et al. (2010) reported $^{10}$Be ages from two moraines deposited by a cirque and valley glacier in the Gaddagh valley ~4 km west of our study site (Fig. 2.2). Given the improved understanding of the $^{10}$Be production rate, we have recalculated these ages using the same production rate as for our new ages (see Methods). Since the production rate is now lower, the ages are older, by up to 3 kyr, than those reported by Harrison et al. (2010), with ages on the outermost moraine being 19.4±1.6 ka (GV1) and 25.9±1.9 ka (GV2), and ages on the inner moraine being 16.8±1.9 ka (GV4) and 17.1±1.1 ka (GV3).

The southern slopes of the Macgillycuddy’s Reeks are ice-scoured and suggest erosion from more contiguous southern-sourced ice (Warren, 1979). The minimal LGM reconstructions that suggest the existence of a KCIC propose an ice divide ~20 km south of the Reeks with northward flowing ice that reached ~700 m along the southern flanks of the Reeks before diverging around the mountains and converging on the northern side of the range, leaving the highest elevations ice-free as nunataks (Fig. 2.2) (Wright, 1927; Warren, 1979; Ballantyne et al., 2011).
Figure 2.2. Top panel) Map of the MacGillycuddy’s Reeks redrawn from Warren (1979) highlighting the important locations and geomorphic features including moraines, terraces, and cirques. Bottom panel) Elevation profile of the MacGillycuddy’s Reeks with the approximate Kerry-Cork Ice Cap surface build-up along the southern slopes and thinning as it diverts around the mountain range. Base map from Google Maps.
minimal reconstruction does not include the possibility of contemporaneous cirque glaciation in the Reeks.

We dated moraines formed by a cirque glacier in the Alohart valley located on the eastern side of the Macgillycuddy’s Reeks, ~2 km west of the Gap of Dunloe (Fig. 2.2). Alohart is the lowest of the cirque basins in the Reeks (373 m cirque-floor elevation). A col located at the top of the cirque headwall reaches an elevation of 640 m. The cirque basin contains two distinct moraines (Fig. 2.4). The elongate outer moraine extends ~1750 m beyond the headwall and is characterized by lateral moraines with broad crests and steeply sloping sides and a terminal moraine with more gentle slopes. The arcuate inner moraine extends ~750 m beyond the headwall and also exhibits lateral moraines with steep proximal slopes and a terminal moraine with more gentle slopes. The crests of the lateral portions of the inner moraine are separated from the crests of the lateral portions of the outer moraine by a well-defined topographic low. The elevation difference from the base of the cirque headwall to the toe of the outer terminal moraine is ~250 m. Each moraine contains abundant, locally sourced quartzolithic sandstone boulders of the Devonian Old Red Sandstone. Warren (1979) proposed that the high-relief, steeply sloping Gearha moraine that crosses the valley ~2 km north of the Alohart moraines was deposited as a lateral moraine by the LGM KCIC where it encircled the northern portion of the Reeks.

2.4 Methods

We collected samples from 16 boulders from the two Alohart moraines for $^{10}$Be cosmogenic surface exposure dating (Fig. 2.3). Each boulder was at least 0.5 m high above the ground surface, thus limiting the possibility of post-depositional exhumation, and located near or on the crest of the moraine, thus limiting the possibility of post-depositional movement. Sampled boulders had faceted surfaces indicating glacial erosion during transport. Quartz veins, when present, were low relief (< 1 cm) suggesting little-to-no post-depositional erosion of the sampled surface. Our calculations do not account for possible erosion, but sensitivity tests suggest that a conservative erosion estimate of 1 mm ka$^{-1}$ (Ballantyne and Stone, 2012) would increase our ages by < 2%. Samples were from the upper 2 cm of each
boulder’s top surface. Samples were processed for $^{10}\text{Be}/^{9}\text{Be}$ measurements at the Oregon State University Cosmogenic Isotope Laboratory following the procedures of Licciardi (2000), and AMS measurements were made at the Purdue Rare Isotope Measurement (PRIME) Laboratory.

The $^{10}\text{Be}$ ages were calculated with the CRONUS-Earth online calculator (Balco et al., 2008) using the northeast North American (NENA) production rate (Balco et al., 2009) and the time-dependent scaling scheme of Lal (1991) and Stone (2000). We use the NENA production rate because it has greater latitudinal similarity with the our field site than the Arctic production rate (Young et al., 2013); calculated ages using the two rates have an average difference of <1.5% (Table A2). We also use the NENA production rate rather than recent production rates derived from Scotland (Ballantyne and Stone, 2012; Small and Fabel, 2015) because of greater confidence in the geologic constraints used to derive the NENA value. We note that although using the higher Scottish production rates reduces our ages by ~8%, they do not change our conclusions (see Supplementary Materials). Shielding factors for each sample were calculated using the CRONUS-geometric shielding calculator (Table A1). The moraine ages are reported using the arithmetic mean age of each sample set and the standard error of the distribution with the production rate uncertainty added in quadrature. Mean ages are interpreted as reflecting the end of moraine construction, and therefore the onset of deglaciation (Licciardi, 2009).

2.5 Results

Nine $^{10}\text{Be}$ ages from the outer moraine range from 21.1±0.8 ka to 27.7 ± 0.8 ka (Fig. 2.5A). Based on Chauvenet’s criterion, there are no statistical outliers. The mean age and uncertainty of the outer moraine is 24.5±1.4 ka. Six of the seven ages from the inner moraine range from 18.8±1.1 ka to 22.9±0.7 ka (Fig. 2.5B). One sample from this moraine is 55.1±1.4 ka and is an outlier based on Chauvenet’s criterion, likely due to inheritance. The mean age and uncertainty of the six remaining samples is 20.4±1.2 ka. A lack of moraines up-ice from the inner moraine suggests that this age represents the final deglaciation of the Alohart cirque basin. The spread in the distribution of our ages from each moraine (Fig. 2.5) is comparable to that
Figure 2.3. Glacially deposited boulders along the crests of the inner (A and B) and outer (C and D) moraines in Alohart cirque basin that were sampled for surface exposure dating.
below, is that their reconstruction predates 24.5±1.4 ka, with cirque glaciation developing after the KCIC had thinned and retreated from the Alohart drainage.

Although they favored an englacial origin for the trimlines, Ballantyne et al. (2011) also reconstructed a regional ice surface based on the alternative interpretation that trimlines record the former ice surface. Figure 2.6 shows their reconstruction but modified to include the constraints from our Alohart cirque ages, assuming that their ice-surface reconstruction is contemporaneous with or younger than the 24.5±1.4 ka age from Alohart. These data support the hypothesis by Warren (1979) that the southern-sourced KCIC was diverted around the Macgillycuddy’s Reeks, leaving the mountains ice free and forming piedmont lobes in the lowlands to the north.

Because our results demonstrate that the northern flanks of the Macgillycuddy’s Reeks above an elevation of ~200 m were not covered by the IIS nor the KCIC since at least 24.5±1.4 ka, those maximum reconstructions showing a contiguous IIS covering southwestern Ireland at 23 ka (C. Clark et al., 2012) or 24 ka (Greenwood and Clark, 2009) (Fig. 2.1D) are too large. Instead, our results are consistent with the minimal ice-sheet reconstructions for Ireland showing a separate KCIC since 24.5±1.4 ka (Fig. 2.1), albeit with less ice than previously inferred. Our results are also consistent with an intermediate reconstruction (BIM-1 model of Shennan et al., 2006) that shows a thin KCIC (ice thickness <125 m) over the Macgillycuddy’s Reeks at 24 ka, but suggests that another, intermediate reconstruction with ice thicknesses >375 m over the Alohart site at 24 ka and 21 ka (Brooks et al., 2008) is too thick (Fig. 2.1C).

One of the critical records for constraining the timing of the maximum Irish and British ice-sheet extent on the western continental shelf comes from the Barra Fan, where the first occurrence of glacimarine turbidites at ~27 ka is interpreted as marking the arrival of the ice-sheet margin to its LGM limit (Kroon et al., 2000; Wilson et al., 2002). Turbidite sedimentation ended ~23 ka, suggesting onset of retreat of the LGM margin at that time. On the Irish Sea coast, a calibrated 14C age on reworked shells from till on the Ards Peninsula suggests ice advance across the site sometime after 28.6 ka (Hill and Prior, 1968), whereas a limiting calibrated 14C age from a marine core suggests deglaciation of the Irish Sea by ~23.3 ka (Kershaw,
Figure 2.4. Contour map of the Alohart cirque basin with location of dated samples indicated by a white circle and their respective ages shown in text boxes. Italicized age represents a statistical outlier that has been associated with cosmogenic nuclide inheritance. Moraine crests are indicated by the white lines with arrows indicating the proximal ice-contact surface. Base map from Apple Maps.
found in other dating studies of glacier moraines (Putnam et al., 2013; Rood et al., 2011), and indicates that the various sources of geologic uncertainty (exhumation, erosion, and inheritance) that cannot be identified when sampling are insignificant. On the other hand, differences resulting from large inheritance are not expected to be systematic and will introduce a large spread in the age distribution with significant differences between ages (Bentley et al., 2010). Our one sample from the inner moraine that is ~30 kyr older than the other moraine ages is a typical example of an outlier due to inheritance.

2.6 Discussion

Our new $^{10}$Be ages indicate that a cirque glacier occupied the Alohart drainage and deposited moraines at 24.5±1.4 ka and 20.4±1.2 ka. The two ages on the inner moraine from the Gaddagh valley 4 km to the west (Harrison et al., 2010) (Fig. 2.2) extend this interval of cirque glaciation in the Reeks to ~17 ka. The two ages from the outer moraine from the Gaddagh valley are similar to our two age populations from Alohart. We do not know the context for the two boulders, but we speculate they may be recording the same two events as at Alohart. We were unable to find any additional suitable boulders for sampling from the associated outer Gaddagh Valley moraine.

Trimlines and ice-flow features in the mountainous region of southwestern Ireland have long been used to argue for a KCIC (Wright, 1927; Farrington, 1954; Warren, 1979; Rae et al., 2004). Ballantyne et al. (2011) considered two alternative hypotheses for the trimlines: that they represent a former ice surface, or that they record an englacial transition from erosive warm-based ice below to non-erosive cold-based ice above. Ballantyne et al. (2011) favored the second englacial interpretation and reconstructed the ice surface with an elevation of 500 m north of the Macgillycuddy’s Reeks, therefore requiring that the Alohart cirque was overridden by the KCIC. Our data, however, clearly demonstrate that the KCIC did not override the Alohart cirque within the last 24.5±1.4 ka, suggesting that if their reconstruction is for a KCIC for that time or later, its elevations north of the Macgillycuddy’s Reeks are too high. An alternative hypothesis, discussed further
Figure 2.5. A) Probability distribution of the nine samples located on the outer moraine. B) Probability distribution of the six samples located on the inner moraine. Vertical line represents the reported arithmetic mean. Gray bar represents the combined analytical and production rate uncertainty.
Figure 2.6. Map of the proposed contours of minimum ice surface altitude for the Kerry-Cork Ice Cap redrawn from Ballantyne et al. (2011) and modified to account for our ages from Alohart cirque basin. Arrows indicate generalized directions of ice movement. Triangles mark glaciated summits with their respective elevations in meters. Circle marks the location of the Alohart valley. Base map from Google Maps.
1986). Agreement of these age constraints from the western and eastern IIS margins thus suggests that the LGM IIS was ~27-23 ka (J. Clark et al., 2012).

We cannot exclude reconstructions in which a larger KCIC or the IIS covered the Reeks before 24.5±1.4 ka, generally corresponding to the end of the LGM IIS. As discussed above, the KCIC reconstruction by Ballantyne et al. (2011) may predate cirque glaciation. Alternatively, Greenwood and Clark (2009) reconstructed an IIS that covered southwestern Ireland at 28 ka. In either case, if the Alohart site was ice covered, our age for the outer moraine (24.5±1.4 ka) requires substantial thinning and retreat of any ice cover by that time, and is thus consistent with the end of the LGM IIS being at ~23 ka (J. Clark et al., 2012).

2.7 Conclusions

We present new 10Be ages that identify deposition of two moraines by a cirque glacier in the Alohart drainage of the Macgillycuddy’s Reeks, southwestern Ireland, at 24.5±1.4 ka and 20.4±1.2 ka. Two ages from a moraine 4 km to the west extend the interval of cirque glaciation within the Macgillycuddy’s Reeks to ~17 ka (Harrison et al., 2010). The ages on the oldest Alohart moraine demonstrate that the Macgillycuddy’s Reeks have not been covered by either the IIS or KCIC since at least 24.5±1.4 ka. Reconstructions of an extensive IIS covering southwestern Ireland and extending onto the adjacent continental shelf at 23 or 24 ka are thus too large, whereas reconstructions of a KCIC covering the Macgillycuddy’s Reeks need to be reduced, unless that reconstruction relates to an earlier phase of the LGM. Our results are consistent with reconstructions of a separate KCIC that was diverted around the Macgillycuddy’s Reeks to form piedmont lobes in the lowlands to the north. Complete coverage of the Reeks during an early phase of the LGM by either the IIS or the KCIC remains a possibility, although substantial thinning and ice-margin retreat would be required by the time the oldest moraine was deposited in the Alohart cirque basin 24.5±1.4 ka.

2.8 Acknowledgements
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2.9 References


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Chapter 3

Persistent millennial-scale cirque-glacier fluctuations in Ireland between 24,000 and 10,000 years ago

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3.1 Abstract

The abrupt climate changes of the last deglaciation originated from changes in the Atlantic meridional overturning circulation (AMOC) and associated changes in ocean heat transport, likely with a strong feedback from changes in sea ice (Clark et al., 2002; Liu et al., 2009). Because of its location immediately adjacent to, and downwind of, the North Atlantic Ocean, glacial records from Ireland are strategically located to monitor the full climatic effects of abrupt changes in the AMOC. Given their short response times, former Irish cirque glaciers should provide a robust record of such climate variability, but dating their fluctuations remains poorly constrained (Ballantyne et al., 2008; Harrison et al., 2010). Here we report $^{10}$Be ages on Irish cirque moraines that show a persistent millennial-scale signal throughout the interval from 24.5±1.4 ka to 10.8±0.7 ka. Many of our new moraine ages are associated with known climatic events driven by changes in the AMOC, suggesting a forced response. On the other hand, our ages also show that a millennial-scale signal persists during an interval (~21 to 15 ka) of reduced AMOC and surface air temperatures. This signal is consistent with modeling results that identify kilometer-scale glacier fluctuations on centennial-to-millennial timescales solely in response to high-frequency natural variability (Roe and O'neal, 2009).

3.2 Introduction

High-resolution cosmogenic $^{10}$Be chronologies of small-glacier fluctuations have provided important new insights into climate variability and its causes on centennial-to-millennial timescales, particularly in regions where development of other climate proxies is limited (Licciardi et al., 2009). In general, these ages constrain glacier fluctuations that can be associated with the well-known millennial- and orbital-scale variability of the last 20,000 years (Shakun et al., 2015). Recently developed chronologies for the Irish Ice Sheet (IIS), however, indicate that it experienced higher frequency fluctuations than would be inferred from the canonical deglacial climate variability associated with changes in the Atlantic meridional overturning circulation (AMOC)(Clark et al., 2012). Many cirque basins in Ireland are fronted by two or more moraines (Colhoun and Synge, 1980; Harrison et al.,
Figure 3.1 – Site locations. Locations of the cirque basins from this study. Numbers correlate to location name. 1 – Alohart, 2 – Carrawaystick, 3 – Lougaharry, 4 – Sruhauncullinmore, 5 – Corranabinna, 6 – Accorymore, 7 – Bunnafreva, 8 – Glascirns Hill.
suggesting that they may preserve a similar record of climate variability, but their ages remain largely unconstrained.

We used $^{10}$Be to date 80 boulders from 15 moraines in eight cirque basins that are located from southern (52.0° N) to northern (54.8° N) Ireland, with all but one cirque basin being on or near the west coast (Fig. 3.1). We calculated $^{10}$Be ages with the CRONUS-Earth online calculator (v. 2) (Balco et al., 2008) using the northeast North American production rate (Balco et al., 2009) and the time-dependent scaling scheme of Lal (1991) (Lal, 1991) and Stone (2000) (Stone, 2000). We report the arithmetic mean or error-weighted mean age of each sample set depending on whether the geologic uncertainty (standard deviation of the sample ages) or the analytical uncertainty is larger; production rate uncertainty is added in quadrature (Methods). Mean $^{10}$Be ages are interpreted as reflecting the onset of deglaciation from the moraines (Licciardi et al., 2009).

3.3 Results

Given the low probability of moraine preservation in any given basin (Gibbons et al., 1984; Roe and O’neal, 2009), we consider our moraine ages collectively to represent a composite record of cirque-glacier variability. In this regard, moraine ages, which range from 24.5±1.4 ka to 10.8±0.7 ka, show a persistent millennial-scale signal throughout this interval (Figs. 3.2a, B1). We note that there are no regionally coherent or cirque morphologic (e.g., elevation, size, orientation) patterns that would explain the regional distribution of moraine ages (Fig. B2). The oldest moraine age (24.5±1.4 ka) is from our southernmost site, with an inner moraine from the same cirque basin having an age of 20.4±1.2 ka (Table B1). These ages indicate that the IIS did not cover the site during the Last Glacial Maximum (LGM), or that the LGM IIS deglaciated the site prior to ~24.5 ka (Barth et al., 2016). Otherwise, all remaining cirques were likely covered by the IIS at the LGM, and glaciers only occupied the cirques after ice-sheet retreat, with our oldest ages from each cirque providing minimum-limiting ages for IIS deglaciation. On Achill Island, western Ireland, ages on the middle and innermost moraines from Bunnafreva Lough are 20.7±1.1 ka and 18.9±1.0 ka, respectively, whereas ages on the two moraines
Figure 3.2 – Deglacial chronology and known climatic forcings. a, 10Be moraine ages for cirque deglaciation. Reported ages follow the methodology outlined in the text. b, Pa/Th records from Böhm et al. (2015) (blue) and McManus et al. (2004) (green) showing the relative strength of Atlantic Meridional Overturning Circulation (AMOC). c, Greenland Summit ice core δ18O from GRIP (black).
from Lough Accorymore are 18.4±1.0 ka and 17.3±0.9 ka. Thirty km inland from Achill Island, two moraines in the Corranabinna cirque are 19.5±1.1 ka and 15.4±1.0 ka. Moraine ages from our northernmost site (Glascairns Hill) are 17.4±0.9 ka, 15.3±0.8 ka, and 12.2±0.8 ka. Three moraines from two cirque basins in western Ireland (Lougaharry and Sruhauncullinmore) have ages of 14.2±0.8 ka, 12.9±1.0 ka, and 11.5±0.7 ka. The youngest moraine age (10.8±0.7 ka) comes from the eastern site (Carrawaystick) in the Wicklow Mountains.

3.4 Discussion

Our initial hypothesis had been that Irish cirque glaciers would record the large and abrupt warming events associated with the onset of the Bølling (14.7 ka) and the end of the Younger Dryas (YD; 11.7 ka) (Fig. 3.2). This is supported by the moraine ages at Lough Lougaharry (14.2±0.8 ka) and Sruhauncullinmore (11.5±0.7 ka). We suggest that the oldest moraine age in our composite record identifying retreat at 24.5±1.4 ka is also associated with the resumption of the AMOC and attendant warming at the end of Heinrich Stadial 2 (24.3 ka) (Fig. 3.2b, 3.2c). Similarly, moraine ages that occur during the Allerød (12.9±1.0 ka) and after the YD (10.8±0.7 ka) may reflect a forced response to centennial-scale North Atlantic climate changes at these times (Inter-Allerød cold period at 13 ka, cold event at 10.3 ka (Bond et al., 1997)), although our age uncertainties preclude a robust correlation. Our moraine age suggesting retreat during the Younger Dryas (12.2±0.8 ka) may be plausibly related to a climate oscillation at this time seen in a number of high-resolution records (Carlson et al., 2007), or possibly increased climate variability with increased incursions of warm air across the Nordic Seas and northern Europe beginning ~12.2 ka (Bakke et al., 2009).

Our remaining moraine ages identify a millennial-scale signal that persists during an interval (~21 ka to 15 ka) of reduced AMOC and surface air temperatures in the North Atlantic region (Buizert et al., 2014) (Fig. 3.2c) that should favor a positive surface mass balance (SMB) for glaciers across Ireland (Liu et al., 2009). We note that this signal persists despite increasing summer insolation and greenhouse gases across much of this interval (Fig. 3.3). Notably, our moraine ages during this
period are in good agreement with well-dated recessional phases of the IIS (Clark et al., 2012): onset of the Cooley Point Interstadial ≥20 ka (moraine ages of 20.7±1.1 ka, 20.4±1.2 ka, 19.5±1.1 ka); IIS retreat from western Ireland at 18.3±1.0 ka (18.9±1.0 ka, 18.4±1.0 ka); onset of the Linns Interstadial ≥17.0 ka (17.3±0.9 ka, 17.4±0.9 ka); and onset of the Rough Island Interstadial ~15.5 ka (15.4±1.0 ka, 15.3±0.8 ka). This shared variance further points to a previously unrecognized forcing that influenced the SMB of the IIS as well as Irish cirque glaciers.

Although there are no large abrupt changes in North Atlantic climate during this cold climate state, δ¹⁸O records from Greenland ice cores identify considerable high-frequency temperature variability (Fig. 3.2c). One model has shown that, in the absence of any forced climate change, such variability can lead to kilometer-scale changes in glacier length on centennial to millennial timescales (Roe and O'neal, 2009). Specifically, when forced by normally distributed random variations in precipitation and melt-season temperature, a glacier with a given geometry and a characteristic response time will experience normally distributed glacier length fluctuations as it integrates the SMB anomalies. Under such a forcing, therefore, a typical glacier will spend 30% of its time fluctuating order hundred meters at ±1σ, and 0.3% of its time fluctuating order a kilometer at ±3σ. Behavior of this sort means that each subsequent readvance of a larger size overrides and thus destroys the moraine evidence of the higher frequency shorter readvances, leading to the greatest preservation potential of moraines recording the largest kilometer-scale readvances (Gibbons et al., 1984; Roe and O'neal, 2009).

These model results suggest that the millennial-scale signal of our moraine ages during the relatively stationary cold period between the LGM and onset of the Bølling may be due entirely to natural variability. Here we evaluate this possible forcing by quantifying the variability recorded in three Greenland δ¹⁸O records (GISP2, GRIP, and NGRIP), each of which has a sampling resolution of 20 years (Methods) (Fig. 3.3). Whereas variability in all of these records is normally distributed, Figure 3 shows that periods of higher/lower variability relative to the mean state occur on multi-centennial to millennial time scales (Fig. B3). According to the glacier model (Roe and O'neal, 2009), we thus expect that the periods of higher
Figure 3.3 – North Atlantic climate variance. a, $^{10}$Be moraine ages for cirque deglaciation. Reported ages follow the methodology outline in the text. b-d, Variance records from Greenland ice cores $\partial^{18}$O (NGRIP, GRIP, and GISP2) indicating periods of higher (red) or lower (blue) variance versus the mean state. e, Mean insolation
at 53ºN. f, Atmospheric CO2 concentration from the West Antarctic Ice Sheet Divide Ice Core.
variability would result in even greater glacier length fluctuations than during periods of lower variability, thus further developing a robust millennial-scale signal of moraine preservation such as seen in our record.

3.5 Conclusions

In summary, our new Irish cirque glacier chronology preserves a remarkable millennial-scale signal of glacier fluctuations that persists from the LGM until the early Holocene. A large fraction of this signal can be explained by a forced response to changes in the AMOC and associated abrupt climate changes that would have had a particularly strong impact on the SMB of Irish cirque glaciers (Liu et al., 2009). In contrast, many of our moraine ages record a millennial-scale signal that persists during a period from 21 ka to 15 ka associated with a reduced AMOC and cold climate state. Moreover, this signal persists despite increasing radiative forcing from increasing summer insolation and greenhouse gas concentrations during this interval. We conclude that our signal is consistent with forcing from high-frequency natural variability alone, with periods of enhanced variability at multi-centennial to millennial timescales amplifying glacier response to produce a robust millennial-scale moraine record.

3.6 Methods

3.6.1 Sample collection

We collected samples from 119 boulders from 18 moraines for $^{10}$Be cosmogenic surface exposure dating. Each processed boulder was at least 0.3 m high above the ground surface, thus limiting the possibility of post-depositional exhumation. Samples were located near or on the crest of the moraine, thus limiting the possibility of post-depositional movement. Sampled boulders had faceted surfaces indicating glacial erosion during transport. Quartz veins, when present, were low relief (< 1 cm) suggesting little-to-no post-depositional erosion of the sampled surface. Areas of the boulder surface that exhibited pitting were avoided. Our calculations do not account for possible erosion, but sensitivity tests suggest that a conservative erosion estimate of 1 mm ka$^{-1}$ would increase our ages by < 2% (Ref. 1).
Samples were collected using a hammer and chisel from the upper 2 cm of each boulder’s top surface. Geographic location and elevation were recorded using a handheld GPS and compared against topographic maps of the region.

3.6.2 Laboratory methods

Samples were processed for $^{10}\text{Be}/^{9}\text{Be}$ measurements at the Oregon State University Cosmogenic Isotope Laboratory following the procedures of Licciardi (2000) (Licciardi, 2001) and Marcott (2011) (Marcott, 2011). Rock samples were crushed and sieved to isolate the 250-710 µm fraction. Quartz separation was achieved through initial magnetic separation and subsequent leaching of the sample in a dilute solution of 2% HF and 2% HNO₃. Samples were tested for quartz purity with an ICP-OES at the University of Colorado – Boulder. Anion exchange, cation exchange, and pH adjustment steps were performed to isolate the beryllium (Kohl and Nishiizumi, 1992). A reference $^{9}\text{Be}$ spike with a concentration of 358 ppm was added to each sample. During each sample batch (usually nine samples), a procedural blank was processed to assess background levels of contamination in the cosmogenic lab. $^{10}\text{Be}/^{9}\text{Be}$ ratios were measured by accelerator mass spectrometry at the PRIME Laboratory at Purdue University.

3.6.3 Age calculations and reporting

The $^{10}\text{Be}$ ages were calculated with the CRONUS-Earth online calculator (Balco et al., 2008) using the northeast North American (NENA) production rate (Balco et al., 2009) and the time-dependent scaling scheme of Lal (1991) (Lal, 1991) and Stone (2000) (Stone, 2000). We use the NENA production rate because it has greater latitudinal similarity with the our field site than the Arctic production rate (Young et al., 2013); calculated ages using the two rates have an average difference of <1.5%. We also use the NENA production rate rather than recent production rates derived from Scotland (Ballantyne and Stone, 2012; Small and Fabel, 2015) because of greater confidence in the constraints used to derive the NENA value. We note that although using the higher Scottish production rates reduces our ages by ~8%, they do not change our conclusions. Shielding factors for each sample were calculated using
the CRONUS-geometric shielding calculator using field measurements (Table B1). The moraine ages are reported using either the arithmetic mean or error-weighted mean age of each sample set depending on whether the geologic uncertainty (standard deviation of the sample ages) or the analytical uncertainty is larger, respectively. Choosing the larger of the two presents the more conservative option. When the arithmetic mean age is reported the age uncertainty is reported as the standard error of the distribution with the production rate uncertainty added in quadrature. For error-weighted mean ages we report the error-weighted uncertainty with the production rate uncertainty added in quadrature. Mean ages are interpreted as reflecting the end of moraine construction, and therefore the onset of deglaciation (Licciardi et al., 2009).

3.6.4 Greenland variance calculations

Variance of $\delta^{18}O$ for each record of the Greenland ice cores NGRIP (Andersen et al., 2007b), GRIP (Andersen et al., 2007a), and GISP2 (Rasmussen et al., 2007)) was calculated using a 9-point moving window. Each record was smoothed using a 17-point moving average. The mean value for each record spanning the time interval of 22.99 to 15.01 ka was subtracted from each data point to create a time series of variance relative to the mean state.
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Chapter 4

Climate evolution across the Mid-Brunhes Transition

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4.1 Abstract

The Mid-Brunhes Transition (~430 ka; MBT) has been described as an increase in the glacial-interglacial climate cycles of the past 800,000 years. Temperature records from ice cores indicate that prior to 430 ka the interglacials were in fact cooler than those after the MBT. Additional records of benthic oxygen isotopes suggest that the interglacials from 430-800 ka experienced higher ice volumes relative to the younger interglacials. Yet to be answered though is whether the MBT was a global phenomenon or regional. Here we characterize the climate system across the MBT through geostatistical analyses of multiple climate proxies including sea-surface temperatures, benthic carbon isotopes, and dust accumulation. Our results demonstrate that the MBT was in fact a global event with a significant increase in climate variance. Furthermore, a sequence of events leading up to the MBT starting at MIS 14 may suggest a more complex transition.

4.2 Introduction

The last 800 kyr of the Pleistocene epoch is characterized by the emergence of dominant ~100-kyr glacial-interglacial cycles (Clark et al., 2006; Raymo et al., 1997). These climate cycles typically have long glacial periods punctuated by short, warm interglacials. Since ~430 ka (i.e., starting with Marine Isotope Stage (MIS) 11), the interglacials have experienced warmer temperatures (Jouzel et al., 2007; Lisiecki and Raymo, 2005) and higher concentrations of atmospheric carbon dioxide (CO$_{2\text{atm}}$) (Luthi et al., 2008) relative to the earlier interglacials of the last 800 kyr. The transition to higher amplitude interglacials has also been recognized in deep-sea records of $\delta^{18}$O measured in benthic foraminifera (Lisiecki and Raymo, 2005) that identify smaller volumes of ice and/or warmer deep-ocean temperatures in the post-MBT interglacial climate.

This shift of the climate system at ~430 ka thus suggests a change in the 100-kyr cycles at a time of minimal change in orbital forcing. Although originally described as a singular event and referred to as the Mid-Brunhes Event (MBE) (Jansen et al., 1986), Yin [2013] suggested that it is more appropriately considered as a transition between two distinct climate states, thus referring to it as the Mid-
Brunhes Transition (MBT). Attempts have been made to provide a mechanism for this change by invoking changes to Antarctic Bottom Water (AABW) formation through mechanisms such as insolation induced feedbacks on sea ice and surface water density to create a more vigorous pre-MBT AABW formation (Yin, 2013), or latitudinal shifts in the position of the Southern Ocean westerlies increasing upwelling of respired carbon in the post-MBT deep ocean (Kemp et al., 2010). Several questions remain, however, such as (1) how and when was the MBT expressed in different components of the climate system and (2) was this a global or regional transition. Here we provide a statistically robust characterization of changes occurring over the last 800 kyr as recorded by a variety of paleoclimatic proxies with broad spatial coverage. Our results identify a more complex sequence of changes in various components of the climate system before the MBT, suggesting a more complex explanation for it.

4.3 Methods

We compiled a number of published proxy records of sea-surface temperature (SST), benthic marine carbon isotopes ratios ($\delta^{13}C$), and dust accumulation (Dust) (Fig. 4.1). Most proxy records span the time period of 8-800 ka in order to characterize the climate system before and after the MBT (~430 ka). Each data set has an average temporal resolution higher than 5 kyr, does not include any large gaps, spans the entire time period of consideration to limit biasing of the younger parts of the record, and is on the LR04 age model (Lisiecki and Raymo, 2005) or contains the necessary information (depth or benthic $\delta^{18}O$) to be placed on it. Each record was then interpolated to a time step ($\Delta t$) of 2 kyr. With each record having an average resolution higher than 5 kyr, this $\Delta t$ allows for the preservation of higher-frequency variability while limiting the number of interpolated data points.

Each compiled proxy data set was analyzed using empirical orthogonal function analysis (EOF) to statistically characterize the dominant modes of variability and robustly demonstrate global and regional signals. Spectral analyses of each resulting principal component (PC) were used to characterize their periodicity, phase, and amplitude.
Figure 4.1 – Site locations. Map indicating the locations of the cores used in this research. Each circle represents a different proxy record. Red – sea-surface temperatures. Green – benthic $\delta^{13}C$. Orange – dust. Black – bottom-water temperature. White – greenhouse gases.
Sea-surface temperature

There are 11 SST records that met our criteria for the 800-kyr time period, but we increased this to 15 records by shortening the time period 8 – 750 ka. Inclusion of these four shorter records does not change our conclusions. The SST records cover the Pacific (n = 9), Atlantic (n = 5), and Indian (n = 1) Oceans (Table C1). Records with initial age models that were not on LR04 were placed on it in one of two ways. If the original data had depth data, the SST record was placed on LR04 using the ager script in MatLAB as part of the Arand software package (Howell et al., 2006). When only $\delta^{18}O$ were available, the SST records were placed on LR04 using the Analyseries version 2.0 software (Paillard et al., 1996).

Carbon isotopes ($\delta^{13}C$)

We used the benthic $\delta^{13}C$ records compiled by Lisiecki (2014) (Lisiecki, 2014). Our initial analysis was conducted on the global set of all $\delta^{13}C$ records (n = 18; Fig. 4.1). Additionally, we separated the $\delta^{13}C$ records by basin focusing on the Atlantic (n = 14) and the Pacific (n = 4), thus distinguishing differences in the dominant water masses within each basin and removing the muting effect of the more negative Pacific values on the more positive Atlantic. We also looked at regional and depth stacks of the $\delta^{13}C$ records in the Atlantic to characterize changes in the dominant water masses on orbital time scales. Regional stacks were broken into North Atlantic (> 20° N; n = 4), Equatorial Atlantic (20° S to 20° N; n = 14), and South Atlantic (> 20° S; n = 8). Additional stacks were created based on depth to analyze the influences of water masses. These stacks include the deep North Atlantic (depth > 2000 m; n = 4) and intermediate North Atlantic (depth < 2000 m; n = 3). Each stacked record was linearly interpolated to a 2-kyr-time step. All included records were averaged to create the stack.

Dust

We use nine proxy records of dust that span the entire 800-kyr time period. We also separately analyze dust records from the Northern (n = 4) and Southern (n = 5) hemispheres to characterize hemispheric differences. The various proxies for
Figure 4.2 – Principal components. Plots of the first (PC1; black) and second (PC2; red) principal components from our EOF analysis of each climate variable. Percent variance explained by each PC represented by the numbers with the corresponding color. a, Dust records. b, Sea-surface temperatures. c, δ13C of the Atlantic. d, δ13C of the Pacific. e, Global δ13C.
Figure 4.3 – Spectral power analyses. Power spectral density plots of major climate variables. Blue plot represents the pre-MBT time interval (800-450 ka). Red plot represents the post-MBT (350-8 ka). Vertical lines represent the dominant Milankovitch periods at 100-, 41-, and 23-kyr. a, EPICA Dome C CO2. b, Dust PC1. c, Atlantic δ13C PC1. d, Sea-surface temperatures PC2. e, Sea-surface temperatures PC1.
Figure 4.4 – Wavelet analysis. Wavelets of four of the first principal components. a, Sea-surface temperatures. b, Dust records. c, $\delta^{13}C$ of the Atlantic. d, Global $\delta^{13}C$. Red colors represent higher spectral power. Blue colors represent lower spectral power. Statistical significance highlighted by the thin black line. Milankovitch periods highlighted by the dashed horizontal lines.
“dust” include Fe mass accumulation rates, weight percent of terrigenous material and Fe, flux of lithogenic grains, and grain size analysis. We standardized each record before analysis to account for these various proxy types and their range in values, thus allowing for comparison of their relative amplitudes of variation. Not all records used were on the LR04 age model, and a lack of necessary data prevented us from putting a few records on the same age model. For this reason, certain analyses of the dust records are not considered robust including phase/lag relationships. Nevertheless, the overall variance in the pre- and post-MBT records is likely preserved.

Empirical Orthogonal Function analysis (EOF)

We used EOF analysis to objectively characterize the climate variability recorded by all proxies across the MBT. The records for SST and \( \delta^{13}C \) were kept in their original values of degrees and per mil, respectively, to preserve the original variance. Dust records were standardized to a mean value of zero and unit variance so that each record provided equal weight to the EOF. In most analyses, the first EOF represents a majority of the variance, ranging from 49-80%, while the second EOF accounts for 7-20%. Statistical significance of all EOFs was determined through segmented linear regression analysis. All resulting break points occur on or after the second EOF and are thus considered significant. For all four proxy types analyzed, the first two EOFs explain 70% or more of the total variance with the first EOF representative of at least half of the total.

Spectral analysis

We used the Blackman-Tukey technique in the Arand software package for spectral analysis of each PC. Multiple tests were conducted for the time slices 8-800 ka, 450-800 ka, and 8-350 ka. These intervals characterize the dominant frequency of variability over the entire 800-kyr record, and for the pre- and post-MBT intervals. The removal of the 350-450 ka interval limited the influence of MIS 11, MIS 12, and Termination V (T5) as these were shown to potentially bias the spectral power. Furthermore, these selected intervals result in time series of equal length to limit
biasing of longer records. Additional tests were conducted using wavelet analyses that characterize the change in spectral power as a time series. Complementary spectral analyses were conducted on atmospheric CO₂ (CO₂ atm) and atmospheric methane (CH₄) taken from the EPICA Dome C ice core (Jouzel et al., 2007; members, 2004), and benthic δ¹⁸O using the LR04 stack (Lisiecki and Raymo, 2005). Cross-spectral analyses were conducted for the resultant PCs against mean insolation values to determine phase and coherency of each. Mean insolation values were calculated for each of the dominant periodicities (eccentricity, obliquity, and precession) with the data derived from Analyseries (Laskar et al., 2004; Paillard et al., 1996).

Variance tests

We used f-tests to test for variance changes across the MBT for each principal component from the EOF analysis as well as for atmospheric carbon dioxide and methane and benthic δ¹⁸O values. If the resulting variance values reject the null hypothesis of no statistical difference, then they are determined to have identified a significant change in variance across the MBT. We interpret the change in variance to reflect differences in amplitude of each climate signal.

4.4 Results

Sea-surface temperatures

EOF analysis of global SST over the last 758 kyr results in two statistically significant principal components (Fig. 4.2b). The first and second principal components (PC1 and PC2, respectively) account for 69% of the total variance with PC1 explaining 49% alone. Factor loadings indicate that each record positively contributed to PC1 with a larger contribution coming from high-latitude records. Thus, PC1 is representative of a global SST signal. PC1 demonstrates a stepwise increase in interglacial SST values starting with MIS 11, which is one of the properties used to define the MBT. The highest spectral density is in the 100-kyr-frequency band throughout the entire time period (Fig. 4.3). Wavelet analysis (Fig. 4.4a) shows a significant increase in the 100-kyr-frequency band 580 ka that reaches
its maximum spectral power during MIS 11 and persists throughout the remainder of the interval. Figure 4.3e also shows that the 100-kyr spectral power increases across the MBT. Additionally, variance tests reveal a significant increase in variability from the pre- to post-MBT SSTs (Table C2). These results thus confirm that there was a stepwise global transition of SST from lower to higher variability in the 100-kyr glacial-interglacial cycles as previously inferred from individual records.

Variance calculations on proxies of bottom water temperature (Elderfield et al., 2010) and on the Antarctic EPICA ice-core deuterium record (members, 2004), a measure of Antarctic atmospheric temperature, indicate statistically significant increases in variance across the MBT. In both proxies, the time series indicate an increase of interglacial temperature values while showing no significant change to the lower limit glacial values, similar to PC1 of SSTs (Fig. 4.5).

**Dust**

The EOF analysis of the global dust records yielded two statistically significant principal components with PC1 representing 56% of the total variance and PC2 15% (Fig. 4.2a). All records but the one from the Chinese Loess Plateau (CLP) reflect increased dust due to increased aridity or wind strength during glaciations, whereas the CLP reflects increased summer Asian monsoon strength, which is an interglacial signal (Sun and An, 2005). Accordingly, factor loadings for the dust records are all positive for PC1 except for CLP. Similar to the SST PC1, there is a stepwise change in the amplitude in the dust PC1, but this increase occurs during MIS 12 rather than MIS 11, reflecting increased amplitude of the glacial signal rather than the increased interglacial SST signal.

Separating the records by hemisphere shows that the increase in glacial amplitude starting at MIS 12 occurs in the southern PC1 but not in the northern PC1 (Fig. 4.6). Similarly, the signal during MIS 14 present in the global PC1 is absent in the northern PC1, suggesting that the northern control on dust accumulation was skipped during that glacial.

Spectral analysis of the global PC1 indicates dominant power in the 100-kyr frequency band that increases in spectral power across the MBT (Fig. 4.3).
Furthermore, wavelet analysis of PC1 demonstrates an increase in the spectral power of the 100-kyr band at ~600 ka with its highest power during MIS 11 (Fig. 4.4b), similar to the SST PC1. The 100-kyr frequency remains statistically significant throughout the interval of 100-600 ka. Additional variance tests confirm an increase in dust variance across the MBT (Table C2).

$\delta^{13}C$

The global $\delta^{13}C$ ($\delta^{13}C_G$) PC1 yields a first principal component that explains 58% of the total variance (Fig. 4.2e). EOF analysis of $\delta^{13}C$ records from the Atlantic basin ($\delta^{13}C_{ATL}$) yields two statistically significant PCs with PC1 and PC2 explaining 58% and 13% of the total variance, respectively (Fig. 4.2c). Both the global and Atlantic PC1 exhibit a strong 100-kyr frequency that is persistent through most of the time interval (680-180 ka) (Fig. 4.4). However, unlike SST and dust, $\delta^{13}C$ demonstrates a stronger 100-kyr power prior to the MIS 11 that has its highest power throughout MIS 13 and 12 (510-460 ka). Furthermore, Fig. 4.3 shows a decrease in spectral power of the 100-kyr-frequency band from pre- to post-MBT. F-tests for the pre- and post-MBT intervals show that $\delta^{13}C_G$ and $\delta^{13}C_{ATL}$ are statistically different with the pre-MBT $\delta^{13}C_{ATL}$ exhibiting higher variance (Table C2).

Factor scores for $\delta^{13}C_{ATL}$ PC1 are all positive suggesting that the time series is representative of the entire Atlantic basin. In contrast, $\delta^{13}C_{ATL}$ PC2 yields negative values for all but the intermediate North Atlantic records and does not show strong 100-kyr spectral power. Glacial shoaling reduces the penetration of NADW formation to ~2000 m in the North Atlantic (Curry and Oppo, 2005). The sites with positive factor scores in PC2 are located at depths < 2000 m, and therefore each site should remain consistently bathed in NADW through glacial-interglacial cycles. We thus interpret PC2 as a record of changes in the isotopic values of the North Atlantic carbon reservoir. Similar to the $\delta^{13}C_{ATL}$ PC1, f-tests of PC2 indicate an increase in variance from pre- to post-MBT.

EOF analysis of $\delta^{13}C$ records from the Pacific ($\delta^{13}C_{PAC}$) yields one statistically significant principal component (PC1 = 81% total variance) (Fig. 4.2d). $\delta^{13}C_{PAC}$ PC1 is similar to the global and $\delta^{13}C_{ATL}$ PC1s in spectral analyses and
Figure 4.5 – Temperature records. a, Deuterium-based temperature record from EPICA Dome C in Antarctica (light yellow; Jouzel et al., 2007). The darker yellow line is a 15-point moving average. b, The first principal component of our sea-surface temperature analysis (green). c, Bottom water temperature derived from Mg/Ca measurements at ODP 1123 (light blue; Elderfield et al., 2012). Dark blue line is a 15-point moving average.
Figure 4.6 – Dust principal components. The first principal components of our dust analysis for the global (yellow), north (light blue), and south (dark blue) records. Vertical gray boxes highlight specific glacial (dark gray) and interglacial (light gray) periods. The numbers indicate the associated Marine Isotope Stage of each box.
Figure 4.7 – Atlantic δ13C proxy comparison. Comparison of the global δ13C first principal component (PC1; black) compared against a, EPICA Dome C CO2 (yellow; EPICA community members, 2007, Lüthi et al., 2008) and b, sea-surface temperature PC1 from this research (blue).
variance tests. The only difference between the three PC1s is less variance recorded in $\partial^{13}C_{\text{PAC}}$. We interpret this muted signal to be a result of three contributions: the large size of the Pacific relative to the Atlantic, less mixing between water mass end members such as the positive NADW and more negative AABW, and ocean circulation aging the carbon isotopes over time leading to more homogenized water masses in the Pacific.

During MIS 13, all three $\partial^{13}C$ PC1s (global, Atlantic, and Pacific) demonstrate high positive values. This excursion, first recognized in individual records by Raymo et al. (1997), stands out relative to other interglacial values recorded throughout the last 800 kyr. The MIS 13 excursion is even more apparent when compared against other proxy records such as CO$_2$ (Fig. 4.7a) and SST (Fig. 4.7b). This high amplitude change in $\partial^{13}C$ values is similar to the changes recorded in other proxies during MIS 11, yet precedes the MBT by one glacial cycle.

$\partial^{13}C$ gradients

Figure 4.8 shows regional stacks of $\partial^{13}C$ from the deep and intermediate North Atlantic and the deep South Atlantic. As discussed, the intermediate North Atlantic (INA) signal is predominantly controlled by changes in the carbon reservoir over orbital time scales. In contrast, the deep North Atlantic (DNA) is controlled by changes in the relative influence of NADW and AABW as well as any $\partial^{13}C$ changes recorded in the values of the NADW (i.e., INA). Subtracting the INA from the DNA record therefore removes the influence of reservoir changes, with the residual time series then reflecting only the relative influences of AABW and NADW in the North Atlantic. This is supported by comparing the North Atlantic depth gradient time series against the South Atlantic stack (Fig. 4.9). Both time series demonstrate good correlation for the entire time interval ($r^2 = 0.58$, but even more striking is the similarity in $\partial^{13}C$ values, with both time series showing similar variability and range in $\partial^{13}C$ space. Moreover, the correlation between the two records increases starting at MIS 15.

The depth gradient no longer shows the MIS 13 excursion that was present in the original DNA stack (Fig. 4.9), suggesting that the excursion is likely a change in
Figure 4.8 – Regional $\delta^{13}C$ stacks. Stacked records of benthic $\delta^{13}C$ separated into three regions: Intermediate North Atlantic (orange), Deep North Atlantic (blue), and Deep South Atlantic (black). All plots shown in $\delta^{13}C$ space to highlight different isotopic values.
Figure 4.9 – Depth gradient and South Atlantic δ13C. Depth gradient of the North Atlantic δ13C records (Deep minus Intermediate; green) compared with the Deep South Atlantic δ13C stack (black). Each record has been smoothed using a 9-point moving average. Both are plotted in δ13C space to highlight the similarity in values once the isotopic influence of the Intermediate North Atlantic is removed. Horizontal lines indicate a period of lower correlation ($r^2 = 0.63$) prior to MIS 15, and higher correlation ($r^2 = 0.77$) after the MIS 15.
the carbon reservoir (represented by the INA) and not related to ocean circulation. Figure 4.10 shows contour $\delta^{13}C$ plots of the Atlantic basin for MIS 13 and MIS 5e. The MIS 13 plot clearly shows an enrichment of the entire basin relative to average post-MBT interglacial conditions. The global $\delta^{13}C$ PC1 also shows the MIS 13 $\delta^{13}C$ excursion, suggesting that it reflects a change in the global carbon reservoir.

We next evaluated the gradient between the South Atlantic signal and the DNA signal in order to further assess the relative influence of the more negative AABW $\delta^{13}C$ values on $\delta^{13}C$ values in the North Atlantic (Fig. 4.11). The closer the gradient is to zero, the more similar the two water masses are to each other. Lisiecki (2014) interpreted weaker gradients during glaciations to reflect shoaling of NADW and greater penetration of AABW, which could result from reduced NADW formation or stronger AABW formation. Figure 4.8b shows a noticeable stepwise drop in mean values during MIS 12, with greater similarity between the North Atlantic and South Atlantic glacial and interglacial values afterwards.

4.5 Discussion

These new analyses demonstrate that there was a statistically significant increase in variance at 430 ka in global SSTs, atmospheric CO$_2$ and CH$_4$, bottom-water temperature, and Antarctic temperature. This increase, which corresponds to the MBT as conventionally defined, is also associated with an increase in spectral power in the 100-kyr frequency band. While these proxies indicate an increase in interglacial values starting with MIS 11, the dust analyses suggest that this transition to greater variability was experienced in the Southern Hemisphere in the glacial periods starting in MIS 12.

MIS 13 carbon isotope excursion

Our record of $\delta^{13}C_G$ shows a high correlation with the GHG records (CO$_2$ and CH$_4$) for a majority of the last 800 kyr (Fig. 4.7) other than during MIS 13, when CO$_{2\text{atm}}$ levels were still at pre-MBT levels while the $\delta^{13}C_{\text{ATL}}$ shows an anomalously high enrichment relative to other interglacial values. To further illustrate this point, Figure 4.10 shows that the Atlantic basin was enriched in $\delta^{13}C$ during MIS 13 relative
Figure 4.10 – MIS 13 and 5e contour plots of δ13C. Contour plots of the δ13C values in the North Atlantic basin for the interglacials MIS 13 and MIS 5e. Red colors represent more positive, enriched values. Blue colors represent lower, depleted values. Plot created using Ocean Data Viewer.
Figure 4.11 – Latitudinal $\delta^{13}C$ gradient. a, North Atlantic regional $\delta^{13}C$ stack plotted in $\delta^{13}C$ space (dark blue). b, Latitudinal gradient of Atlantic $\delta^{13}C$ regional stacks (North Atlantic minus South Atlantic; light blue). Lower values demonstrate increased similarity between the records. Stepwise change in the gradient highlighted by the gray box. c, South Atlantic regional $\delta^{13}C$ stack plotted in $\delta^{13}C$ space (black).
to the MIS 5e, further suggesting that relatively little light carbon was present in the oceans during MIS 13.

The proxy Ba/Fe from the Antarctic Zone (AZ; ODP 1094) records the sedimentary concentration of biogenic Ba and is thus a proxy of organic matter flux to the deep ocean (Jaccard et al., 2013), whereas alkenone concentrations from the Subantarctic Zone (SAZ; ODP 1090) indicate export productivity to the deep ocean in the region north of the Polar Front (Martínez-García et al., 2009). Based on these proxies, Jaccard et al. (2013) argued that there were two modes of export productivity in the Southern Ocean (SO), where high/low values occur in the AZ during interglacials/glacials, and low/high values occur in the SAZ during interglacials/glacials. They interpreted the increase in SAZ export productivity as due to iron fertilization from increased dust accumulation in the SAZ associated with intensified SO westerlies during glacial periods. Our Southern Hemisphere dust PC1 record supports this hypothesis in showing that high values of dust accumulation correlate well with increased values of SAZ export productivity over the last 800 kyr (Fig. 4.12). We note, however, that the increase in dust starting at MIS 12 does not have an associated decrease in glacial CO$_{2\text{atm}}$ values, suggesting that while iron fertilization may have increased SAZ export productivity, it apparently had little affect on CO$_{2\text{atm}}$ levels.

The antiphase relationship between export productivity between the SAZ and AZ, however, requires a mechanism to increase organic matter productivity during interglacials. In the modern SO, vertical mixing and upwelling drive the delivery of nutrient-rich waters necessary for biologic activity in the surface ocean. Wind-driven upwelling is associated with a poleward shift of SO westerlies during interglacials (Toggweiler et al., 2006). Thus, any reduction of upwelling would be a result of either a more northerly position or decrease in strength of the westerlies (Jaccard et al., 2013). Further reduction of nutrient-rich surface waters in the AZ during glacials could be a product of increased SO stratification, or a combination of both. We note, however, that Jaccard et al. (2013) find no AZ export productivity during MIS 13 whereas all other interglacials over the last 800 kyr show some evidence for it (Fig. 4.12c). This skipped interglacial in export productivity suggests either a change in the
Figure 4.12 – Dust and productivity. a, the first principal component of southern dust (blue). The y-axis has been inverted to show interglacial periods as up. b, Alkenone record of export productivity from the Subantarctic Zone (ODP 1090; Martinez-Garcia et al, 2009; red). c, Ba/Fe ratios show export productivity from the Antarctic Zone (ODP 1094; Jaccard et al., 2013; black). The vertical gray box highlights a period of little to no export productivity or dust during MIS 13.
Figure 4.13 – Marine isotope stages 15 to 13 and the carbon isotope excursion. a, First principal component of Atlantic δ13C (black). b, EPICA Dome C CO2 (yellow; EPICA community members, 2004, Lüthi et al., 2008). c, Detrended sea-level equivalent from Shakun et al., 2015 (light blue). Derived from δ18Osw calculations. Negative numbers indicate lower sea level and increased ice volume. d, Chinese Loess Plateau grain size indicating relative Asian summer monsoon strength (brown;
Sun et al., 2005). e, Quartz/Calcite ratios from site U1313 in the North Atlantic as a measure of ice-rafted debris (teal; Naafs et al., 2012). Dark gray bars highlight the interglacials (MIS 15 and MIS 13) between ~630 to ~470 ka. Light gray bar highlights MIS 14.
position/强度 of the SO westerlies, or stratification of the AZ limited the delivery of nutrient-rich deep waters to the surface.

The PC1s of δ^{13}C (global, Atlantic, and Pacific) demonstrate that the oceans were enriched in heavy carbon during MIS 13 relative to any other interglacial of the last 800 kyr. In contrast, atmospheric CO₂ concentrations were ~240 ppm during MIS 13, similar to other pre-MBT interglacial levels. Records of organic export productivity from the AZ and SAZ indicate no peaks during this interglacial. Therefore the question becomes: if the ocean is heavily enriched in δ^{13}C during MIS 13 while CO₂(atm) and export productivity remained at low levels, where is the light carbon?

Paleoclimate records from the CLP indicate increased precipitation during MIS 13 relative to the other interglacials (Liu, 1985; Yin and Guo, 2008). This increased precipitation has been attributed to increased monsoon activity that has been recognized throughout monsoonal areas of the Northern Hemisphere and persisted through MIS 15, 14, and 13 (Guo et al., 2009; Yin and Guo, 2008). Biogenic silica measurements from Lake Baikal exhibit continuously high terrestrial productivity throughout MIS 11 to MIS 15 (Prokopenko et al., 2002). Sea-level reconstructions indicate that ice volume during MIS 14 was considerably less relative to other glacial maxima of the last 800 kyr (Fig. 4.13c) (Elderfield et al., 2012; Shakun et al., 2015). Thus, the smaller ice sheets of MIS 14 would not have affected as much of the forested areas of the Northern Hemisphere. The increased monsoon activity and smaller ice volume during MIS 14 would have combined to increase land biomass that continued into MIS 13. The Northern Hemisphere thus had the potential to store light carbon in the terrestrial realm resulting in the enriched δ^{13}C MIS 13 signal seen in the ocean basins (Yin and Guo, 2008).

**NADW influence in the South Atlantic**

The latitudinal gradient between North and South Atlantic δ^{13}C records shows a stepwise decrease starting at T5 (~430 ka) (Fig. 4.11), indicating greater similarity between AABW and NADW water masses. Prior to the MBT, the higher gradient between the NA and SA may reflect either increased influence of the AABW in the
north or decreased influence of NADW in the south, whereas after the MBT, this relationship switches. As previously discussed, the depth $\delta^{13}C$ gradient from the North Atlantic (Fig. 4.9) reflects changes in NADW formation, and shows increased similarity through higher correlation between the gradient and the South Atlantic across the MBT.

Models of glacial-interglacial cycling of $CO_{2\text{atm}}$ have shown a dominant role of the Southern Ocean in storing and releasing dissolved inorganic carbon (DIC) in the deep Southern Ocean (Sigman et al., 2010). Deglacial release of $CO_2$ is thought to occur through increased upwelling and vertical mixing of AABW. Therefore, we suggest that $CO_{2\text{atm}}$ reflects DIC storage in the Southern Ocean and is thus a reflection of the volume of AABW (Ferrari et al., 2014). The relatively cool interglacials of the pre-MBT are associated with a persistently large volume occupied by AABW from 800-430 ka, whereas the post-MBT $CO_{2\text{atm}}$ record suggests a reduced volume of the interglacial AABW water mass. Glacial values of $CO_{2\text{atm}}$ remain reasonably constant throughout the last 800 kyr, suggesting that the change in relative AABW volume only occurs during interglacials (Yin, 2013).

Contour plots of average interglacial $\delta^{13}C$ values in the Atlantic indicate that the pre-MBT experienced penetration of AABW north of the equator (Fig. 4.14a). Removal of MIS 13 and its associated enriched carbon isotope excursion further highlights the greater influence of AABW in the pre-MBT interglacial Atlantic (Fig. 4.14b). By comparison, contour plots of the post-MBT Atlantic show a greater southern penetration of NADW during interglacials, indicating a reduced volume of AABW. This reorganization of the dominant interglacial water masses in the Atlantic basin across the MBT is consistent with greater release of deep-ocean $CO_2$ during the post-MBT interglacials.

A cross-spectral analysis of pre-MBT North and South Atlantic $\delta^{13}C$ stacks indicates in-phase coherency between the records in both the 100-kyr and 41-kyr frequencies. Similar tests for the post-MBT $\delta^{13}C$ stacks exhibit coherency in eccentricity, obliquity, and precession with the South Atlantic stack leading the North Atlantic by $\sim$23º (7 kyr) in eccentricity, $\sim$18º (2 kyr) in obliquity, and $\sim$36º (2 kyr) in precession (Fig. 15). All phase relationships overlap within uncertainty, suggesting
Figure 4.14 – Average interglacial $\delta^{13}C$ contours. Contour plots of the average interglacial $\delta^{13}C$ values in the Atlantic for a, pre-MBT included MIS 13, b, pre-MBT excluding MIS 13 (enriched carbon isotope excursion), and c, post-MBT. Red colors indicate higher $\delta^{13}C$ values. Blue colors indicate lower $\delta^{13}C$ values. Boundary between the two water masses (NADW and AABW) indicated at the 0.25‰ contour (Curry and Oppo, 2005).
Figure 4.15 – Post-MBT δ13C phase wheels. Phase wheels for each of the Milankovitch cycles (eccentricity, obliquity, and precession) between the North and South Atlantic δ13C regional stacks. The arrow at the top of each wheel shows an in-phase relationship of the South Atlantic with the North Atlantic in that frequency band. Values to the right indicate a lag by the South Atlantic relative to the North Atlantic. Values to the left indicate a lead by the South Atlantic relative to the North Atlantic. Purple shading indicates the uncertainty in each relationship. Dotted line highlights the 90% confidence interval for each frequency.
that the South Atlantic δ¹³C leads the North Atlantic δ¹³C by 2-7 kyr in the post-MBT. This lead by the South Atlantic is most apparent during terminations and is most likely related to deglacial mechanisms for ventilation of respired CO₂ from the deep such as enhanced wind-driven upwelling of the Southern Ocean or the melting of sea ice from the bipolar seesaw (Sigman et al., 2010). Such ventilation is expressed in the South Atlantic stack as an enriching of δ¹³C values. We suggest that the stepwise decrease in the latitudinal gradient coupled with the lead of the South Atlantic δ¹³C reflects a reduction in AABW formation across the MBT, whereby the pre-MBT was dominated by a reduced NADW and as a result further northward penetration of AABW, while the post-MBT saw a reduced AABW and evasion of CO₂ from the deep.

4.6 Conclusions
Statistical analyses of multiple climate proxies address the hypothesized transition of lower to higher amplitudes in glacial-interglacial values. EOF tests of sea-surface temperatures resulted in a global record of dominant SST variability that demonstrates a statistically significant increase in interglacial values across the MBT. This increase is reflected in other climate proxies such as CO₂\text{atm}, CH₄, and δ¹⁸O. Significant increases in the amplitude of marine dust accumulation records indicate that this phenomenon was similarly expressed in glacialis.

EOF analyses of basinally separated benthic carbon records highlight an anomalous enrichment of the ocean δ¹³C during MIS 13. This enrichment occurred at a time of “cool interglacial” values of CO₂\text{atm}. Records of summer monsoon strength on the Chinese Loess Plateau indicate persistently strong monsoons from MIS 15 through 13. A missed glacial period during MIS 14 permitted a build-up of land biomass that subsequently led to the oceanic δ¹³C enrichment during MIS 13.

A gradient of North and South Atlantic δ¹³C stacks indicates a sharp increase in similarity across Termination V. Phase relationships between the two stacks exhibit a lead of the South Atlantic by 2-7 kyr and is evidence for a reduction of AABW formation in the post-MBT. We therefore suggest that the MBT represents a
reorganization of the Atlantic water masses controlled by either a reduced influence of AABW or an invigoration of NADW influence in the South Atlantic.

In summary, we have further documented the increases in interglacial sea-surface and Antarctic temperatures, CO$_2$, and CH$_4$ during MIS 11 that have been the basis for characterizing the MBT. However, our analyses document a number of changes in other components of the climate system that began as early as MIS 14 and that may suggest a more complex sequence of events was involved in the MBT (Fig. 4.16), although the exact mechanisms involved and their relationship to the MBT remain unclear. In particular, we note an increase in Asian summer monsoon strength that began during MIS 15 and persisted through MIS 14 and into MIS 13. The strong monsoon strength during MIS 14 is associated with a weak glacial, which would have been conducive a build-up of Northern Hemisphere land biomass. A continued strong Asian summer monsoon during MIS 13 associated with greater precipitation would have further sequestered land biomass and provided a reservoir for light carbon, resulting in the oceans becoming unusually enriched in $\partial^{13}$C. MIS 12 was associated with the return of large ice sheets, collapse of the Asian summer monsoon, the first increase in amplitude of Southern Hemisphere dust, and a drop in the latitudinal gradient of Atlantic $\partial^{13}$C suggesting a reorganization of the water masses in the basin.
Figure 4.16 – Schematic representation of the sequence of events leading to the Mid-Brunhes Transition. Corresponding marine isotope stages are located on the left side of each row. Boxes in a row indicate synchronous events.
4.7 References


Curry, W.B., Oppo, D.W., 2005. Glacial water mass geometry and the distribution of δ13C of ΣCO2 in the western Atlantic Ocean. Paleoceanography 20, n/a-n/a.


Chapter 5

Conclusions

This dissertation is a compilation of three studies that address climate variability on millennial-to-orbital timescales. Questions in paleoclimatology and glacial geology have been addressed through geochemical and geostatistical techniques.

5.1 Chapter summaries

In Chapter 2, $^{10}$Be data from two cirque moraines in the Alohart basin of the MacGillycuddy’s Reeks in southwest Ireland yield ages of 24.5±1.4 ka and 20.4±1.2 ka. The oldest age unequivocally demonstrates that either the Irish Ice Sheet (IIS) or the Kerry-Cork Ice Cap (KCIC) did not cover the MacGillycuddy’s Reeks since 24.5±1.4 ka. Reconstructions of the IIS that show a contiguous northern-sourced ice sheet extended onto the southwestern continental shelf during the Last Glacial Maximum are thus too large, whereas reconstructions of the KCIC need to be reduced. These results are consistent with the hypothesis that the KCIC diverted around the MacGillycuddy’s Reeks to form piedmont lobes in the northern lowlands. Ice coverage of the MacGillycuddy’s Reeks prior to 24.5±1.4 ka remains a possibility though substantial thinning would be necessary to allow cirque moraine deposition by the age of the oldest Alohart moraine.

In Chapter 3, a new $^{10}$Be deglacial chronology derived from eight cirque basins across Ireland demonstrate persistent deglaciation over the interval of 24.5±1.4 ka to 10.8±0.7 ka. Ireland is highly sensitive to changes in Atlantic Meridional Overturning Circulation (AMOC) and is demonstrated by correlation of ages in the deglacial chronology. However, many of the new deglacial ages occur during a period of reduced AMOC and cooler North Atlantic temperatures that would normally lead
to positive glacial surface mass balance. Analysis of Greenland ice core variance demonstrates millennial-scale variability of the North Atlantic climate. A glacier model yields kilometer-scale fluctuations of glacier length due to melt-temperature and precipitation variability. We suggest that our chronology is a response to high-frequency natural variability with periods of enhanced variability on millennial-scale timescales.

In Chapter 4, statistical analyses of multiple climate proxies were conducted to address the hypothesized increase in climate cycle amplitude across the Mid-Brunhes Transition (MBT). Empirical orthogonal function tests on records of sea-surface temperatures (SST), benthic carbon isotopes ($\delta^{13}C$), and dust accumulation resulted in objective characterization of climate variability over the last 800-kyr. Our results demonstrate a significant change in variance across the MBT (~430 ka) in proxies such as SST, dust, CO$_2$, CH$_4$, and benthic oxygen isotopes. Our results identified a carbon isotope excursion during Marine Isotope Stage (MIS) 13. Comparison across climate proxies demonstrates a sequence of events starting with an increase of Asian summer monsoons during MIS 15. Low ice volume during MIS 14 suggests reduced glacial conditions allowing for a build-up of land biomass that ultimately led to an enrichment of the global oceans in $\delta^{13}C$. The subsequent glacial period exhibited larger ice volume, collapse of the Asian summer monsoons, stepwise increase in glacial Southern Hemisphere dust concentrations, and a reorganization of the Atlantic water masses. The following glacial termination led to the highest interglacial SST, and CO$_2$ and CH$_4$ atmospheric concentrations that have been identified as the MBT. We now suggest that this sequence of events defines the true transition as opposed to the singular event 430 ka.
## Table A1. Moraine surface-exposure sample details and $^{10}$Be data and ages

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<th>Longitude (DD)</th>
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<th>Length (cm)</th>
<th>Height (cm)</th>
<th>Width (cm)</th>
<th>Sample thickness (cm)</th>
<th>Shielding correction</th>
<th>Quartz weight (g)</th>
<th>Carrier added (g)</th>
<th>$[^{10}\text{Be}]$ / $[^{9}\text{Be}]$ (10$^{-14}$ atoms g$^{-1}$)</th>
<th>AMS standard</th>
<th>$^{10}$Be age and internal uncertainty (ka)</th>
<th>Mean Age$^c$ (ka)</th>
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$^a$Age calculations use standard atmosphere, density of 2.65 g cm$^{-3}$, and zero erosion.

$^b$1-sigma AMS uncertainty.

$^c$Moraine mean age and uncertainty is calculated as the straight mean and standard error with the production rate uncertainty added in quadrature.

$^d$Age excluded using Chauvenet's criterion.
**Table A2.** Procedural blank $^{10}\text{Be}$ data.

| Blank No. | Sample ID    | Carrier Added (g)$^a$ | $^{10}\text{Be}/^{9}\text{Be} \pm 1\sigma$ ($10^{-15}$) | $[^{10}\text{Be}] \pm 1\sigma$ ($10^3$ atoms)$^b$ | AMS Std$^c$ |
|-----------|--------------|------------------------|--------------------------------------------------------|-----------------------------------------------------|
| 1         | 201400642    | 0.7490                 | 1.41 ± 0.71                                            | 25.2 ± 12.7                                          | 07KNSTD     |
| 2         | 201401897    | 0.8446                 | 0.52 ± 0.18                                            | 10.4 ± 3.7                                           | 07KNSTD     |
| 3         | 201500452    | 0.8380                 | 1.73 ± 0.68                                            | 34.7 ± 13.7                                          | 07KNSTD     |

$^a$Carrier $^9\text{Be}$ concentration is 358 ppm.
$^b$Total $^{10}\text{Be}$ (in atoms) in each procedural blank.
$^c$AMS standard which ratios and concentrations are measured. $^{10}\text{Be}/^{9}\text{Be}$ ratio is $2.85 \times 10^{-12}$. 
Table A3. Comparison of $^{10}$Be production rates.

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<th>SCOT(^c) Age (yrs) ± 1(\sigma)</th>
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\(^a\)Northeast North American production rate (Balco et al., 2009) using the Lal (1991) and Stone (2000) time dependent scaling scheme (3.87 ± 0.19 atoms g\(^{-1}\) yr\(^{-1}\)).

\(^b\)1-sigma analytical and production rate uncertainty.

\(^c\)Baffin Bay/Arctic $^{10}$Be production rate (Young et al., 2013) using the Lal (1991) and Stone (2000) time dependent scaling scheme (3.93 ± 0.15 atoms g\(^{-1}\) yr\(^{-1}\)).

\(^d\)Scotland $^{10}$Be production rate (Small and Fabel, 2015) using the Lal (1991) and Stone (2000) time dependent scaling scheme (4.26 ± 0.21 atoms g\(^{-1}\) yr\(^{-1}\)).
Appendix B – Persistent millennial-scale cirque-glacier fluctuations in Ireland between 24,000 and 10,000 years ago
Figure B1 – Sample and moraine ages from Ireland cirque basins. Each circle represents a single sample. Error bars represent 1-sigma analytical uncertainty. Black lines represent the reported age of each moraine and gray bars the uncertainty following the reporting methodology outline in the text. Sample ID numbers are on the x-axis.
Figure B2 – Cirque basin characteristic analyses. Comparison of each moraine age against geographic and morphologic characteristics. a, Cirque latitude (°N). b, Cirque glacier total area (km²). c, Orientation of the cirque (°). d, Width:Length ratio of the glacier. e, Elevation at the base of the headwall (m). f, Slope gradient of the basin floor.
Figure B3 – Spectral analysis of the Greenland ice core variance. Plots of the dominant frequencies found in the variance of the Greenland ice cores (GISP2, NGRIP, GRIP) for the interval of 22.99 to 15.01 ka. Yellow lines highlight certain frequencies for reference. Black vertical lines indicate the 90% confidence interval for each plot.
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Table S1: Moraine surface-exposure sample details and ²⁷⁰Be data and ages
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<td>2</td>
<td>0.960</td>
<td>40.8790</td>
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<td>Outer moraine</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>SCM-13-11</td>
<td>53.5331</td>
<td>-9.78078</td>
<td>140</td>
<td>310</td>
<td>90</td>
<td>140</td>
<td>2</td>
<td>0.973</td>
<td>42.8806</td>
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<tr>
<td>SCM-13-12</td>
<td>53.5331</td>
<td>-9.780712</td>
<td>141</td>
<td>150</td>
<td>60</td>
<td>100</td>
<td>2</td>
<td>0.973</td>
<td>38.9436</td>
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<tr>
<td>SCM-13-13</td>
<td>53.5366</td>
<td>-9.78565</td>
<td>124</td>
<td>160</td>
<td>40</td>
<td>110</td>
<td>2</td>
<td>0.958</td>
<td>40.3191</td>
</tr>
<tr>
<td>SCM-13-14</td>
<td>53.5343</td>
<td>-9.78526</td>
<td>115</td>
<td>130</td>
<td>90</td>
<td>90</td>
<td>2</td>
<td>0.975</td>
<td>40.2410</td>
</tr>
<tr>
<td>SCM-13-15</td>
<td>53.5347</td>
<td>-9.78353</td>
<td>107</td>
<td>180</td>
<td>110</td>
<td>154</td>
<td>2</td>
<td>0.900</td>
<td>40.2054</td>
</tr>
</tbody>
</table>

*a*Age calculations use standard atmosphere, density of 2.65 g cm\(^{-3}\), and zero erosion.

*b*1-sigma AMS uncertainty.

*Moraine mean age and uncertainty is calculated as the straight mean and standard error with the production rate uncertainty added in quadrature.

*d*Age excluded using Chauvenet’s criterion.

*e*Moraine mean age and uncertainty is calculated as the error-weighted mean and uncertainty with the production rate uncertainty added in quadrature.

*f*Age excluded due to morphotectonic discordance.
Table B2. Procedural blank $^{10}$Be data.

<table>
<thead>
<tr>
<th>Blank No.</th>
<th>Sample ID</th>
<th>Carrier Added (g)$^a$</th>
<th>$^{10}$Be/$^9$Be ± 1σ (10$^{-15}$)</th>
<th>$[^{10}$Be] ± 1σ (10$^3$ atoms)$^b$</th>
<th>AMS Std$^c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>201400642</td>
<td>0.7490</td>
<td>1.41 ± 0.71</td>
<td>25.2 ± 12.7</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>2</td>
<td>201401897</td>
<td>0.8446</td>
<td>0.52 ± 0.18</td>
<td>10.4 ± 3.7</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>3</td>
<td>201500452</td>
<td>0.8380</td>
<td>1.73 ± 0.68</td>
<td>34.7 ± 13.7</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>4</td>
<td>201500529</td>
<td>0.8440</td>
<td>0.86 ± 0.49</td>
<td>17.3 ± 9.9</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>5</td>
<td>201501106</td>
<td>0.8610</td>
<td>2.45 ± 0.63</td>
<td>50.5 ± 34.5</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>6</td>
<td>201501098</td>
<td>0.8760</td>
<td>1.45 ± 0.63</td>
<td>30.4 ± 13.2</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>7</td>
<td>201501608</td>
<td>0.8615</td>
<td>0.79 ± 0.32</td>
<td>16.3 ± 6.7</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>8</td>
<td>201501943</td>
<td>0.8753</td>
<td>0.77 ± 0.26</td>
<td>16.1 ± 5.4</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>9</td>
<td>201501944</td>
<td>0.8752</td>
<td>0.76 ± 0.26</td>
<td>15.8 ± 5.4</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>10</td>
<td>201503085</td>
<td>0.8748</td>
<td>0.40 ± 0.46</td>
<td>8.3 ± 9.6</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>11</td>
<td>201503095</td>
<td>0.8775</td>
<td>0.74 ± 0.40</td>
<td>15.6 ± 8.3</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>12</td>
<td>201600931</td>
<td>0.8710</td>
<td>1.26 ± 0.45</td>
<td>26.3 ± 9.3</td>
<td>07KNSTD</td>
</tr>
<tr>
<td>13</td>
<td>201600938</td>
<td>0.8814</td>
<td>0.80 ± 0.30</td>
<td>16.9 ± 6.4</td>
<td>07KNSTD</td>
</tr>
</tbody>
</table>

$^a$Carrier $^9$Be concentration is 358 ppm.

$^b$Total $^{10}$Be (in atoms) in each procedural blank.

$^c$AMS standard which ratios and concentrations are measured. $^{10}$Be/$^9$Be ratio is 2.85x10$^{-12}$. 
Alohart Sample: ALH-13-1

$^{10}\text{Be}$ age: 20530 ± 645

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0126
Longitude: W 9.6784
Altitude: 428 m
Boulder Height: 0.70 m
Lithology: Quartz-lithic wacke
Shielding: 0.988

Note:
Alohert Sample: ALH-13-2

$^{10}Be$ age: $22986 \pm 660$

Location: Alohert, MacGillycuddys Reeks, County Kerry

Latitude: N 52.0126

Longitude: W 9.6784

Altitude: 428 m

Boulder Height: 1.00 m

Lithology: Quartz-lithic wacke

Shielding: 0.989

Note:
Alohart Sample: ALH-13-3

$^{10}$Be age: 19987 ± 611

Location: Alohart, MacGillycuddys Reeks, County Kerry

Latitude: N 52.0120

Longitude: W 9.6777

Altitude: 421 m

Boulder Height: 0.90 m

Lithology: Quartz-lithic wacke

Shielding: 0.988

Note:
Alohart Sample: ALH-13-4

$^{10}$Be age: 55131 ± 1351

Location: Alohart, MacGillycuddys Reeks, County Kerry

Latitude: N 52.0137
Longitude: W 9.6790

Altitude: 390 m
Boulder Height: 0.80 m
Lithology: Quartz-lithic wacke

Shielding: 0.995

Note:
Alohart Sample: ALH-13-5

$^{10}$Be age: 25674 ± 778

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0177
Longitude: W 9.6702
Altitude: 354 m
Boulder Height: 0.45 m
Lithology: Quartz-lithic wacke
Shielding: 0.995

Note:
Aloh art Sample: ALH-13-6
$^{10}$Be age:  23664 ± 751
Location:  Aloh art, MacGillycuddys Reeks, County Kerry
Latitude:  N
Longitude:  W
Altitude:  354 m
Boulder Height:  1.10 m
Lithology:  Quartz-lithic wacke
Shielding:  0.997
Note:
Alohart Sample: ALH-13-7

$^{10}\text{Be}$ age: 26189 ± 1100

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0188
Longitude: W 9.6715
Altitude: 332 m
Boulder Height: 0.50 m
Lithology: Quartz-lithic wacke with quartz veins
Shielding: 0.997
Note:
Alohart Sample: ALH-13-9

$^{10}\text{Be}$ age: $21881 \pm 939$

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0183
Longitude: W 9.6708
Altitude: 340 m
Boulder Height: 0.50 m
Lithology: Quartz-lithic wacke
Shielding: 0.998

Note:
Alohart Sample: ALH-13-10

$^{10}$Be age: 24349 ± 832

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0213
Longitude: W 9.6793
Altitude: 261 m
Boulder Height: 1.00 m
Lithology: Quartz-lithic wacke
Shielding: 0.997

Note:
**Alohart Sample: ALH-13-11**

$^{10}$Be age: 25596 ± 1116

Location: Alohart, MacGillycuddys Reeks, County Kerry

Latitude: N 52.0197

Longitude: W 9.6788

Altitude: 294 m

Boulder Height: 1.00 m

Lithology: Quartz-lithic wacke

Shielding: 0.996

Note:
Alohart Sample: ALH-13-12

\(^{10}\text{Be}\) age: 27725 ± 784

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0170
Longitude: W 9.6801
Altitude: 344 m
Boulder Height: 1.00 m
Lithology: Quartz-lithic wacke
Shielding: 0.997
Note:
Alohart Sample: ALH-13-13

$^{10}$Be age: 24484 ± 997

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0166
Longitude: W 9.6799
Altitude: 359 m
Boulder Height: 1.30 m
Lithology: Quartz-lithic wacke
Shielding: 0.995

Note:
Alohart Sample: ALH-13-15

$^{10}$Be age: 21077 ± 847
Locations: Alohant, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0220
Longitude: W 9.6794
Altitude: 236 m
Boulder Height: 0.80 m
Lithology: Quartz-lithic wacke with quartz veins
Shielding: 0.998
Note:
Alohart Sample: ALH-13-16

$^{10}\text{Be}$ age: 20878 ± 823

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N 52.0183
Longitude: W 9.6754
Altitude: 297 m
Boulder Height: 0.80 m
Lithology: Quartz-lithic wacke
Shielding: 0.993

Note:
Alohart Sample: ALH-13-17

$^{10}$Be age: 18783 ± 1115

Location: Alohart, MacGillycuddys Reeks, County Kerry

Latitude: N 52.0182
Longitude: W 9.6746
Altitude: 305 m
Boulder Height: 1.40 m
Lithology: Quartz-lithic wacke

Shielding: 0.995

Note:
Alohart Sample: ALH-13-18

$^{10}$Be age: $19.2 \pm 0.8$ ka

Location: Alohart, MacGillycuddys Reeks, County Kerry
Latitude: N
Longitude: W
Altitude: 306 m
Boulder Height: 0.70 m
Lithology: Quartz-lithic wacke
Shielding: 0.995

Note:
**Corranabinna Lough Sample: CRB-13-1**

$^{10}$Be age: 19935 ± 814 yrs

Location: Corranabinna Lough, Nephin Beg, County Mayo

Latitude: N 53.9771

Longitude: W -9.68077

Altitude: 322 m

Boulder Height: 0.70 m

Lithology: Quartz

Shielding: 0.995

Note:
**Corranabinna Lough Sample: CRB-13-2**

$^{10}$Be age: 21449 ± 1134 yrs  
Location: Corranabinna Lough, Nephin Beg, County Mayo  
Latitude: N 53.9757  
Longitude: W -9.68052  
Altitude: 335 m  
Boulder Height: 1.40 m  
Lithology: Quartz  
Shielding: 0.995  
Note:
Corranabinna Lough Sample: CRB-13-3

$^{10}$Be age: 20885 ± 862 yrs

Location: Corranabinna Lough, Nephin Beg, County Mayo

Latitude: N 53.9756
Longitude: W -9.68045

Altitude: 333 m
Boulder Height: 1.00 m
Lithology: Quartz

Shielding: 0.995

Note:
Corranabinna Lough Sample: CRB-13-4

$^{10}\text{Be}$ age: 17474 ± 699 yrs
Location: Corranabimna Lough, Nephin Beg, County Mayo
Latitude: N 53.9731
Longitude: W -9.67947
Altitude: 331 m
Boulder Height: 1.30 m
Lithology: Micaceous quartzite
Shielding: 0.995
Note:
Corranabinna Lough Sample: CRB-13-5

$^{10}$Be age: 18501 ± 712 yrs
Location: Corranabinna Lough, Nephin Beg, County Mayo
Latitude: N 53.9724
Longitude: W -9.67888
Altitude: 339 m
Boulder Height: 0.50 m
Lithology: Micaceous quartzite
Shielding: 0.977
Note:
Corranabinna Lough Sample: CRB-13-6

$^{10}$Be age:  19090 ± 763 yrs
Location:  Corranabinna Lough, Nephin Beg, County Mayo
Latitude:  N 53.9718
Longitude:  W -9.67948
Altitude:  332 m
Boulder Height:  0.80 m
Lithology:  Quartz
Shielding:  0.981
Note:
Corranabinna Lough Sample: CRB-13-7

\(^{10}\text{Be} \) age: 19484 ± 637 yrs
Location: Corranabinna Lough, Nephin Beg, County Mayo
Latitude: N 53.9719
Longitude: W -9.67978
Altitude: 332 m
Boulder Height: 1.00 m
Lithology: Micaceous quartzite
Shielding: 0.984
Note:
Corranabinna Lough Sample: CRB-14-1

$^{10}$Be age: 15366 ± 562 yrs
Location: Corranabinna Lough, Nephin Beg, County Mayo
Latitude: N 53.9683
Longitude: W -9.68093
Altitude: 376 m
Boulder Height: 0.60 m
Lithology: Quartz
Shielding: 0.981
Note:
Corranabinna Lough Sample: CRB-14-2

$^{10}$Be age: 17205 ± 655 yrs
Location: Corranabinna Lough, Nephin Beg, County Mayo
Latitude: N 53.9678
Longitude: W -9.68019
Altitude: 398 m
Boulder Height: 1.80 m
Lithology: Quartz
Shielding: 0.977
Note:
Corranabinna Lough Sample: CRB-14-3

$^{10}$Be age: $14151 \pm 620$ yrs
Location: Corranabinna Lough, Nephin Beg, County Mayo
Latitude: N 53.9683
Longitude: W -9.68098
Altitude: 376 m
Boulder Height: 1.50 m
Lithology: Micaceous quartzite
Shielding: 0.982
Note:
Corranabinna Lough Sample: CRB-14-4

$^{10}$Be age: $14901 \pm 599$ yrs

Location: Corranabinna Lough, Nephin Beg, County Mayo

Latitude: N 53.9685

Longitude: W -9.68159

Altitude: 372 m

Boulder Height: 1.40 m

Lithology: Micaceous quartzite

Shielding: 0.983

Note:
Carrawaystick Sample: CRS-14-1

$^{10}$Be age: 11415 ± 514

Location: Carrawaystick, Wicklow Mountains, County Wicklow

Latitude: N 52.9602

Longitude: W 6.42674

Altitude: 491 m

Boulder Height: 0.60 m

Lithology: Pegmatitic granite

Shielding: 0.996

Note:
Carrawaystick Sample: CRS-14-2

$^{10}$Be age: $9373 \pm 510$

Location: Carrawaystick, Wicklow Mountains, County Wicklow

Latitude: N 52.9609
Longitude: W 6.42742
Altitude: 588 m
Boulder Height: 1.50 m
Lithology: Pegmatitic granite
Shielding: 0.997

Note:
Carrawaystick Sample: CRS-14-3

$^{10}$Be age:  11852 ± 599
Location:  Carrawaystick, Wicklow Mountains, County Wicklow
Latitude:  N 52.9615
Longitude:  W 6.42912
Altitude:  596 m
Boulder Height:  1.00 m
Lithology:  Pegmatitic granite
Shielding:  0.995
Note:
Carrawaystick Sample: CRS-14-4

$^{10}$Be age: $9670 \pm 411$

Location: Carrawaystick, Wicklow Mountains, County Wicklow

Latitude: N 52.9621
Longitude: W 6.43001
Altitude: 604 m
Boulder Height: 0.80 m
Lithology: Pegmatitic granite
Shielding: 0.994

Note:
Carrawaystick Sample: CRS-14-5

\(^{10}\text{Be}\) age: 139086 ± 3700

Location: Carrawaystick, Wicklow Mountains, County Wicklow
Latitude: N 52.9620
Longitude: W 6.42980
Altitude: 607 m
Boulder Height: 0.70 m
Lithology: Quartz
Shielding: 0.994

Note:
Carrawaystick Sample: CRS-14-6

$^{10}$Be age: 11627 ± 430
Location: Carrawaystick, Wicklow Mountains, County Wicklow
Latitude: N 52.9606
Longitude: W 6.42689
Altitude: 586 m
Boulder Height: 0.75 m
Lithology: Pegmatitic granite
Shielding: 0.997
Note:
Glascairns Hill Sample: GCH-13-1

$^{10}$Be age: 18691 ± 526

Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7785
Longitude: W 8.0174
Altitude: 293 m
Boulder Height: 1.20 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.998

Note:
Glascairns Hill Sample: GCH-13-2

$^{10}\text{Be}$ age: 14079 ± 486
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7785
Longitude: W 8.0174
Altitude: 293 m
Boulder Height: 1.20 m
Lithology: Quartz-bearing granite
Shielding: 0.997

Note:
Glascairns Hill Sample: GCH-13-3

$^{10}$Be age: 16585 ± 493
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7783
Longitude: W 8.0245
Altitude: 307 m
Boulder Height: 1.60 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.984
Note:
Glascairns Hill Sample: GCH-13-4

$^{10}$Be age: $16260 \pm 577$

Location: Glascairns Hill, Blue Stacks, County Donegal

Latitude: N 54.7784
Longitude: W 8.0240

Altitude: 251? m

Boulder Height: 2.00 m

Lithology: Orthoclase, quartz, biotite granite

Shielding: 0.991

Note:
Glascairns Hill Sample: GCH-13-5

\(^{10}\text{Be}\) age: 16497 ± 480
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7785
Longitude: W 8.0228
Altitude: 254 m
Boulder Height: 1.10 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.995
Note:
Glashairns Hill Sample: GCH-13-6

$^{10}$Be age: 17728 ± 580
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7787
Longitude: W 8.0217
Altitude: 165? m
Boulder Height: 1.70 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.995
Note:
Glascairns Hill Sample: GCH-13-7
$^{10}$Be age: 13751 ± 440
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7786
Longitude: W 8.0194
Altitude: 314 m
Boulder Height: 1.50 m
Lithology: Quartz-bearing granite
Shielding: 0.998
Note:
Glascairns Hill Sample: GCH-13-8

$^{10}$Be age: 15150 ± 609
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7783
Longitude: W 8.0217
Altitude: 309 m
Boulder Height: 1.10 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.993

Note:
Glascairns Hill Sample: GCH-13-9

$^{10}$Be age: 75752 ± 1477

Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7782
Longitude: W 8.0213
Altitude: 308 m
Boulder Height: 0.60 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.993

Note:
Glascairns Hill Sample: GCH-13-10

\(^{10}\)Be age: 18158 ± 505

Location: Glascairns Hill, Blue Stacks, County Donegal

Latitude: N 54.7783
Longitude: W 8.0208
Altitude: 310 m
Boulder Height: 0.70 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.994

Note:
Glascairns Hill Sample: GCH-13-11

$^{10}$Be age: 15355 ± 465

Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7780
Longitude: W 8.0190
Altitude: 303 m
Boulder Height: 1.50 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.994

Note:
Glascairns Hill Sample: GCH-13-12

$^{10}$Be age: 17917 ± 584

Location: Glascairns Hill, Blue Stacks, County Donegal

Latitude: N 54.7778
Longitude: W 8.0180
Altitude: 298 m
Boulder Height: 0.90 m

Lithology: Orthoclase, quartz, biotite granite

Shielding: 0.997

Note:
Glascairns Hill Sample: GCH-13-13

$^{10}$Be age: 12117 ± 462
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7769
Longitude: W 8.0247
Altitude: 320 m
Boulder Height: 1.40 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.981

Note:
Glascairns Hill Sample: GCH-13-14

\(^{10}\text{Be}\) age: 13160 ± 481

Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7769
Longitude: W 8.0232
Altitude: 308 m
Boulder Height: 1.50 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.989

Note:
Glascairns Hill Sample: GCH-13-15

$^{10}\text{Be}$ age: 20,572 ± 562

Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7768
Longitude: W 8.0223
Altitude: 299 m
Boulder Height: 1.00 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.991

Note:
Glascairns Hill Sample: GCH-13-16

$^{10}$Be age: 11415 ± 426
Location: Glascairns Hill, Blue Stacks, County Donegal
Latitude: N 54.7766
Longitude: W 8.0211
Altitude: 294 m
Boulder Height: 1.10 m
Lithology: Orthoclase, quartz, biotite granite
Shielding: 0.993

Note:
Bunnafreva Lough Sample: LC-13-1

$^{10}$Be age: 23237 ± 746 yrs
Location: Bunnafreva Lough, Achill Island, County Mayo
Latitude: N 53.9977
Longitude: W 10.18057
Altitude: 293 m
Boulder Height: 2.00 m
Lithology: Quartz vein
Shielding: 0.995
Note:
**Bunnafreva Lough Sample: LC-13-2**

$^{10}$Be age: 18354 ± 557 yrs  
Location: Bunnafreva Lough, Achill Island, County Mayo  
Latitude: N 53.9976  
Longitude: W 10.18016  
Altitude: 338 m  
Boulder Height: 1.00 m  
Lithology: Quartz vein  
Shielding: 0.995  
Note:
Bunnafreva Lough Sample: LC-13-3

$^{10}$Be age: 20020 ± 889 yrs
Location: Bunnafreva Lough, Achill Island, County Mayo
Latitude: N 53.9976
Longitude: W 10.17993
Altitude: 339 m
Boulder Height: 0.40 m
Lithology: Quartz vein
Shielding: 0.995
Note:
Bunnafreva Lough Sample: LC-13-5

$^{10}$Be age: 19207 ± 752 yrs

Location: Bunnafreva Lough, Achill Island, County Mayo

Latitude: N 53.9974

Longitude: W 10.17873

Altitude: 339 m

Boulder Height: 0.40 m

Lithology: Quartz

Shielding: 0.993

Note:
Bunnafreva Lough Sample: LC-13-6

$^{10}$Be age: 18712 ± 808 yrs

Location: Bunnafreva Lough, Achill Island, County Mayo
Latitude: N 53.9971
Longitude: W 10.17860
Altitude: 340 m
Boulder Height: 0.30 m
Lithology: Quartz vein
Shielding: 0.988

Note:
Bunnafreva Lough Sample: LC-13-7
{}^{10}\text{Be} \text{ age: } 20715 \pm 1026 \text{ yrs}
Location: Bunnafreva Lough, Achill Island, County Mayo
Latitude: N 53.9983
Longitude: W 10.17905
Altitude: 319 m
Boulder Height: 1.10 m
Lithology: Quartz
Shielding: 0.988
Note:
**Bunafreva Lough Sample: LC-13-9**

$^{10}$Be age:

Location: Bunafreva Lough, Achill Island, County Mayo

Latitude: N 53.9974

Longitude: W 10.17885

Altitude: 342 m

Boulder Height: 0.80 m

Lithology: Quartzite sandstone

Shielding: 0.988

Note:
Bunnafreva Lough Sample: LC-13-10

$\text{^{10}Be}$ age: $20994 \pm 778$ yrs

Location: Bunnafreva Lough, Achill Island, County Mayo
Latitute: N 53.9982
Longitude: W 10.18003
Altitude: 328 m
Boulder Height: 1.30 m
Lithology: Quartz
Shielding: 0.988

Note:
Nahanagan Sample: NHG-14-1

$^{10}$Be age: 16817 ± 638 yrs
Location: Lough Nahanahan, Wicklow Mountains, County Wicklow
Latitude: N 53.0334
Longitude: W 6.38540
Altitude: 447 m
Boulder Height: 3.00 m
Lithology: Granite
Shielding: 0.998
Note:
Nahanagan Sample: NHG-14-2

$^{10}$Be age: $9504 \pm 384$ yrs

Location: Lough Nahanagan, Wicklow Mountains, County Wicklow

Latitude: N 53.0345

Longitude: W 6.38851

Altitude: 396 m

Boulder Height: 0.80 m

Lithology: Granite

Shielding: 0.996

Note:
Nahanagan Sample: NHG-14-3

$^{10}$Be age:  20697 ± 857 yrs

Location:  Lough Nahanagan, Wicklow Mountains, County Wicklow

Latitude:  N 53.0340
Longitude:  W 6.38660
Altitude:  368 m
Boulder Height:  0.70 m
Lithology:  Granite
Shielding:  0.996

Note:
Sruhauncullinmore Sample: SCM-13-1

$^{10}$Be age: 11110 ± 524 yrs
Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo
Latitude: N 53.6512
Longitude: W 9.78993
Altitude: 208 m
Boulder Height: 1.10 m
Lithology: Quartz-lithic wacke
Shielding: 0.951
Note:
Sruhauncullinmore Sample: SCM-13-2

$^{10}$Be age: 10270 ± 429 yrs

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6512
Longitude: W 9.78993

Altitude: 208 m

Boulder Height: 2.00 m

Lithology: Quartz-lithic wacke

Shielding: 0.951

Note:
Sruhauncullinmore Sample: SCM-13-3

$^{10}$Be age:

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6518
Longitude: W 9.7898
Altitude: 174 m
Boulder Height: 2.50 m
Lithology: Quartz-lithic wacke
Shielding:
Note:
Sruhauncullinmore Sample: SCM-13-4

$^{10}$Be age:
Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo
Latitude: N 53.6519
Longitude: W 9.7900
Altitude: 117 m
Boulder Height: 1.50 m
Lithology: Quartz-lithic wacke
Shielding:
Note:
Sruhauncullinmore Sample: SCM-13-5

$^{10}\text{Be}$ age:

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo
Latitude: N 53.6525
Longitude: W 9.7899
Altitude: 172 m
Boulder Height: 1.00 m
Lithology: Quartz-lithic wacke
Shielding:

Note:
Sruhauncullinmore Sample: SCM-13-6
$^{10}$Be age: 12222 ± 621 yrs
Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo
Latitude: N 53.6523
Longitude: W 9.78907
Altitude: 179 m
Boulder Height: 1.10 m
Lithology: Quartz-lithic wacke
Shielding: 0.960
Note:
Sruhauncullinmore Sample: SCM-13-11

$^{10}$Be age: $14368 \pm 538$ yrs

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6531
Longitude: W 9.78708
Altitude: 140 m
Boulder Height: 0.90 m
Lithology: Quartz-lithic wacke
Shielding: 0.973

Note:
Sruhauncullinmore Sample: SCM-13-12

$^{10}$Be age: 00000 ± 000 yrs

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6531
Longitude: W 9.78712
Altitude: 141 m
Boulder Height: 0.60 m
Lithology: Quartz-lithic wacke
Shielding: 0.973

Note:
Sruhauncullinmore Sample: SCM-13-13

$^{10}$Be age: 10840 ± 552 yrs

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6536
Longitude: W 9.78665
Altitude: 124 m
Boulder Height: 0.40 m
Lithology: Quartz-lithic wacke
Shielding: 0.958

Note:
Sruhauncullinmore Sample: SCM-13-14

$^{10}\text{Be}$ age: 13169 ± 537 yrs

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6543
Longitude: W 9.78526
Altitude: 115 m
Boulder Height: 0.90 m
Lithology: Quartz-lithic wacke
Shielding: 0.975

Note:
Sruhauncullinmore Sample: SCM-13-15

$^{10}$Be age: 12957 ± 644 yrs

Location: Sruhauncullinmore, Mweel Rea Mountains, County Mayo

Latitude: N 53.6547
Longitude: W 9.78535
Altitude: 107 m
Boulder Height: 1.10 m
Lithology: Quartz-lithic wacke
Shielding: 0.980

Note:
### Appendix C – Climate evolution across the Mid-Brunhes Transition

#### Table C1 - Data compilation

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### Table C2 - Variance tests

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Bibliography


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