Changes in Antarctic Bottom Water Formation During Interglacial Periods

Samuel Glasscock
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CHANGES IN ANTARCTIC BOTTOM WATER FORMATION DURING INTERGLACIAL PERIODS

by

Samuel Kirk Glasscock

A Thesis
Submitted to the Graduate School, the College of Arts and Sciences and the School of Ocean Science and Engineering at The University of Southern Mississippi in Partial Fulfillment of the Requirements for the Degree of Master of Science

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ABSTRACT

In the modern Southern Ocean and during the last interglacial period, Marine Isotope Stage 5, there are observations that point to reduced Antarctic Bottom Water (AABW) formation. These reductions are believed to be driven by an increase in the strength of the Southern Ocean density stratification due to surface water freshening. Any reduction in AABW formation has important implications for global climate as AABW plays a vital role in the cycling of carbon in the world’s ocean. The primary question this study seeks to answer is do these AABW reductions occur during any of the other interglacials of the past 500 thousand years? To study AABW changes in the paleoceanographic we look at changes in the redox record. Newly formed AABW is oxygen-rich, so any reduction should lead to a decrease in oxygen concentrations in the deep Southern Ocean. The trace element uranium is useful for studying these redox changes as it is enriched in marine sediments under low-oxygen conditions. When accounting for other factors, such as paleoproductivity, that can also decrease the oxygen concentrations in sedimentary porewater, it is possible to identify changes in AABW using authigenic uranium. The survey conducted by this study found a possible AABW reduction during late Marine Isotope Stage 11 (~395 ka). However, this reduction does not appear to be triggered by a decrease in surface water density. Instead, a change in the strength or position of the southern hemisphere westerly winds is hypothesized to be the driving mechanism behind this reduction.
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A sincere thank you goes out to all the faculty and staff at the Division of Marine Science for all their support during my time at USM.

Lastly, I would like to the National Science Foundation for funding this research.
DEDICATION

This thesis is dedicated to the 2019 Stanley Cup Champions, the St. Louis Blues.

“We went Blues!” – Brett Hull
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<td>Antarctic Bottom Water</td>
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<td>aU</td>
<td>Authigenic Uranium</td>
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CHAPTER I – INTRODUCTION

As the earth continues to experience warm climate conditions, it is becoming more important that we understand how the earth will respond to ever-increasing sea levels and global temperatures. In order to understand what the future may hold, we turn to past warm climate, or interglacial, periods for study. Of particular interest is Marine Isotope Stage 11 (MIS 11), one of the longest interglacial periods (374-424ka) with orbital parameters similar to the present interglacial, the Holocene (~11ka to present) (Droxler et al. 2003). Additionally, MIS 11 exhibited atmospheric CO$_2$ levels that were similar to preindustrial values, and temperatures at high latitudes that were 4°C to 9°C warmer than Holocene conditions (Dutton et al. 2015). These CO$_2$ and temperature conditions led to sea levels that were, at a minimum, close to present values, with some estimates of MIS 11 sea level being 6 to 13 m or as much as 20 m higher than present (Raymo and Mitrovica 2012; Dutton et al. 2015). These high values would suggest significant collapse of both the Greenland and the West Antarctic Ice Sheets (Raymo and Mitrovica 2012).

The purpose of this study is to determine how deep water formation in the Southern Ocean fared during MIS 11. The formation of deep water masses in the Southern Ocean is of critical importance for global climate because this is the primary location where CO$_2$ sequestered in deep water is returned to the atmosphere via upwelling (Sigman et al. 2010). The cause of this sea to air CO$_2$ release is incomplete nutrient utilization in Southern Ocean surface waters (Sigman et al. 2010). In this location, deep water is upwelled and subsequently, downwelled before all of the nutrients can be used by phytoplankton resulting in a net flux of CO$_2$ from the ocean to the atmosphere.
(Sigman et al. 2010). However, over the past 2.5 million years, the time during which the earth has experienced glacial-interglacial cycling, this has not always been the case (Sigman et al. 2010). Changes in nutrient utilization and the interaction between upwelling water and the atmosphere in the Southern Ocean are two of the primary causes of the large amplitude atmospheric CO$_2$ variations between glacial and interglacial periods (Sigman et al. 2010).

While there is a relatively good understanding of how deep water formation changed over glacial-interglacial timescales (as reviewed in Chapter 2), we do not have as clear of a picture as to how Antarctic Bottom Water (AABW) formation varied within interglacial periods. Understanding this water mass is becoming ever more important due to new observations and modeling studies suggesting that anthropogenic global warming is already leading to reductions in deep water formation around Antarctica, and will continue to do so in the future (de Lavergne et al. 2014; Williams et al. 2016).

Anthropogenic global warming impacts AABW formation due to increased freshwater input from melting glaciers which stratifies the water column, preventing the upwelling of deep water around Antarctica (de Lavergne et al. 2016). This led to the disappearance of a polynya, an ice-free area of deep water formation, in the Weddell Sea in 1976, and has prevented its return since then (de Lavergne et al. 2016). Freshwater input into the ocean surface around Antarctica also decreases the salinity, and therefore the density, of dense shelf waters that form in polynyas (Williams et al. 2016). These dense shelf waters contribute to AABW and surface freshening can suppress their formation (Williams et al. 2016). The future likely holds enhanced mass loss from
Antarctica’s ice shelves which will likely work to reduce the production of AABW in the future (Williams et al. 2016).

In addition to the modern reduction in AABW formation, there is strong evidence for a reduction in AABW during the last interglacial period Marine Isotope stage 5 (MIS 5), ~80-130ka. Using authigenic uranium (aU), a portion of the sedimentary U inventory sensitive to redox conditions, Hayes et al. (2014) recorded low porewater oxygen conditions in the sediment core ODP 1094 at ~127,000 years ago. Low oxygen in porewaters could be caused by reduced circulation supplying the oxygen or by increased removal by respiration of organic matter in the sediments. The 127 ka event occurred during a period of relatively low export and delivery of organic carbon to the sediments. In addition, these low oxygen conditions occurred during a period of sea level rise (Hayes et al. 2014). The prevailing hypothesis for the peak in aU at this time is a decrease in the bottom water oxygen concentrations, due to a decrease in AABW formation caused by surface freshening induced stratification (Hayes et al. 2014).

The goal of this study is to use the redox-sensitive trace element, uranium, and its isotopes, to investigate the bottom water oxygen conditions of the Southern Ocean during MIS 11 as another important analogue for future climate. Using authigenic uranium, along with paleoproductivity, sea level, and other proxies, it will be determined if and why AABW formation was altered during MIS 11. Due to the similarities of MIS 11 to the present interglacial period, any variation in AABW formation during MIS 11 may be able to give us some insight into how AABW formation will change in the future and the effects it may have on our climate.

3
Chapter II – BACKGROUND

2.1 The Southern Ocean and the Meridional Overturning Circulation

As mentioned previously, deep water formation in the Southern Ocean plays a critical role in the global climate system (Sigman et al. 2010). To understand exactly how Southern Ocean circulation affects global climate, we must first understand the Southern Ocean’s role in the framework of the Ocean’s meridional overturning circulation (MOC). The simple conceptual model of MOC in the Atlantic Ocean has the ocean circulation divided into two cells rotating in opposite directions (Marshall and Speer 2012; Rintoul 2018). Water in the upper cell is supplied from several sources: the North Atlantic, where surface waters experience buoyancy loss and sink forming North Atlantic Deep Water (NADW), diapycnal mixing at depth, and from upwelling and buoyancy gain in the Southern Ocean, which creates Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW) (Marshall and Speer 2012). The upwelling in the Southern Ocean is driven by the strong westerly winds that blow over the water’s surface (Marshall and Speer 2012). Denser NADW that enters the Southern Ocean is upwelled closer to the continent, where the majority of it experiences buoyancy loss and once again enters the deep ocean as various bottom water classes including AABW, and a smaller portion as Circumpolar Deep Water (CDW) (Marshall and Speer 2012). This upwelling is one of the controls on the extent of the cryosphere, as upwelling water is relatively warm which leads to the melting of both ice on the continental shelf and further out in the ocean.
Figure 2.1 Water Mass Distributions in the Atlantic Ocean

Simplified conceptual model of the Atlantic meridional overturning circulation, with labeled water masses (modified from Marshall and Speer 2012). Note the higher oxygen concentrations in AABW due to its recent contact with the atmosphere near Antarctica.

These overturning cells are important for the transport of heat, oxygen, and carbon dioxide throughout the global ocean (Rintoul 2018). The formation of deep water from surface waters at the poles transports oxygen to the ocean’s interior. Upwelling of water in the Southern Ocean returns nutrients from the deep ocean to the surface, working to balance the transport of nutrients and carbon via export production. In the modern Southern Ocean, the formation of mode and intermediate waters is seen as a sink for both anthropogenic heat and carbon dioxide, with the carbon dioxide balance being influenced both by the outgassing of natural carbon dioxide during AABW formation and the uptake of anthropogenic carbon during mode water formation. Recently, there has been a reduction in this carbon sink, which has been hypothesized to be a result of increased outgassing of natural carbon dioxide due to stronger westerly winds (Rintoul 2018). However, an alternative mechanism could be related to a noted decrease in areas of deep water formation (de Lavergne et al. 2014).
2.2 Past Changes in Ocean Circulation

This style of MOC has not been constant throughout history. Long-lasting changes to the overturning circulation in the Southern Ocean have caused drastic climatic changes over glacial-interglacial cycles, and more rapid changes to MOC during glacial periods and deglaciations have also left their imprint on the marine record. It is important to understand these changes as they may give us some insight as to what an interglacial AABW reduction might look like in the marine record. During glacial periods, atmospheric carbon dioxide levels are reduced, meaning more carbon dioxide must be stored in the deep ocean, as this is the only reservoir that equilibrates with the atmosphere on the necessary timescales (Sigman et al. 2010). Because the Southern Ocean is the primary location where carbon dioxide from the deep ocean is released to the atmosphere, some mechanism must work to reduce this effect (Sigman et al. 2013).

Southern Ocean surface waters could have been more stratified during glacial periods, thus reducing the ability of deep water to upwell, causing more carbon dioxide to be stored in the deep ocean (François et al. 1997). Enhanced glacial stratification may have been related to a northward shift in the position of the Southern Hemisphere westerly winds (Toggweiler et al. 2006). This would have resulted in a low salinity lid building up around the Antarctic continent, decreasing the ability of the saltier deep water to reach the surface (Sigman et al. 2010). Indeed, the density of the deep water during the last glacial maximum has been determined to be colder and saltier, determined by measuring the chlorinity and $\delta^{18}$O of sedimentary porewater from the Last Glacial Maximum (LGM) (Adkins et al. 2002). Additionally, significant freshening of the high latitude ocean in both the northern and southern hemisphere coincides with the onset to
the Earth’s glacial cycles (~2.5 Ma), suggesting that greater stratification at the poles allowed the deep ocean to sequester more carbon dioxide (Sigman et al. 2004). A similar phenomenon has been noted in the Southern Ocean during the transition from 41,000 to 100,000-year glacial cycles that occurred during the Mid-Pleistocene transition (~1.25 Ma - ~700ka). With an increase in the strength of the Southern Ocean halocline, the release of carbon dioxide from the deep ocean is hindered (Hasenfratz et al. 2019).

Increased Southern Ocean sea ice coverage could have also limited the degree of air-sea exchange between deep water and the atmosphere (Stephens and Keeling 2000; Sigman et al. 2010). This process does not require any increase in productivity, thus allowing carbon dioxide to accumulate in the deep ocean without a change in biological pump efficiency (Stephens and Keeling 2000).

Nutrient utilization in the Southern Ocean likely also increased, thus reducing the carbon dioxide outgassing effect (François et al. 1997, Sigman et al. 2010). Nitrogen isotopic evidence from the Southern Ocean points to greater nutrient utilization during this time (François et al. 1997, Wang et al. 2017). Increased nutrient utilization leads to elevated δ^{15}N values due to the kinetic mass fractionation of nitrogen during photosynthesis (François et al. 1997). During the LGM, south of the modern polar front, bulk sediments, as well as fossil organic material in deep sea corals, record higher δ^{15}N than their Holocene counterparts, suggesting increased nutrient utilization (François et al. 1997; Wang et al. 2017).

Past changes in total biological productivity (as opposed to nutrient utilization) are somewhat more complicated. Latitudinal shifts in the zones of highest primary productivity occurred during the last glacial period, with reduced production in the
Antarctic Zone, and increased production in the Subantarctic Zone (Kumar et al. 1995; Chase et al. 2001; Jaccard et al. 2013). For the Antarctic, this is likely due to a combination of reduced wind-driven upwelling and an increase in the density stratification, both of which cause a reduction in deep water formation (Jaccard et al. 2013). Particle flux proxies, such as $^{231}$Pa/$^{230}$Th, which reflects the affinity of $^{231}$Pa for opal particles, show greater Subantarctic export production compared to the modern ocean during the LGM (Kumar et al. 1995). Suboxic porewaters were also recorded, using $\Delta U$, in a well-ventilated core site from the Discovery Seamount during the LGM, suggesting that the flux of organic matter to the sediment was greater in this Subantarctic location (Chase et al. 2001). The primary cause of the increase in Subantarctic productivity is believed to be an increase in aeolian Fe supply to this zone (Martinez-Garcia et al. 2014). This productivity increase coupled with reduced Antarctic overturning and deep water formation would have worked in concert to lower atmospheric carbon dioxide, by allowing for greater nutrient utilization with enhanced carbon sequestration (Jaccard et al. 2013). An additional effect would be the lowering of deep ocean oxygen concentrations as organic matter is respired at depth (Sigman et al. 2010). Therefore, it is likely that a combination of the stratification, sea ice, and productivity mechanisms worked to reduce atmospheric carbon dioxide during glacial periods.

Whatever the mechanism, the deep water in the glacial ocean was more isolated from the atmosphere than it is today. The effects of this deep water isolation are also noted to have decreased during deglaciation. Because deep water was more isolated, it was also much older, with respect to contact with the atmosphere, during the LGM than
during the Holocene (Skinner et al. 2010). For the LGM, this effect can be measured using radiocarbon (Skinner et al. 2010). By comparing the radiocarbon age of the deep water, using benthic foraminifera, to the surface ocean, using planktic foraminifera, and the atmosphere, an age offset can be calculated (Skinner et al. 2010). This age offset was found to be much greater during the LGM, with a benthic-planktic offset of ~1600 years and a benthic-atmosphere offset of ~2000-3750 years, compared to a modern offset of ~550 years and ~1250 years, respectively (Skinner et al. 2010). Furthermore, these age offsets were found to decrease across the deglacial interval, indicating a better ventilated deep ocean, and the $\Delta^{14}C$ of the atmosphere was found to have been diluted by outgassing of aged carbon dioxide (Skinner et al. 2010, Anderson et al. 2009). This resumption of deep ocean ventilation was also recorded by a deglacial increase in the flux of biogenic opal in the Antarctic zone (Anderson et al. 2009). Biogenic opal is a useful upwelling proxy because, in the zone of maximum opal production, surface seawater Si concentrations are totally depleted during Southern Hemisphere summers (Anderson et al. 2009). This means that the supply of Si to the surface via upwelling controls the amount of opal that can be produced (Anderson et al. 2009). This opal flux shows an increase across the deglacial interval with a decreasing flux into the Holocene (Anderson et al. 2009). The increase in opal flux and decrease in deep water age are both associated with an increase in atmospheric carbon dioxide, consistent with a resumption in Southern Ocean upwelling (Skinner et al. 2010, Anderson et al. 2009).

Within this larger deglacial picture are more abrupt changes in Atlantic MOC. Evidence for these changes comes from $^{231}$Pa/$^{230}$Th in the North Atlantic (McManus et al. 2004). Here $^{231}$Pa/$^{230}$Th can be used as a circulation proxy because both Pa and Th are
produced via radioactive decay of U (McManus et al. 2004). While both have a tendency to stick to sinking particles, Th is removed from the water column much more quickly, with Pa having a residence time of a couple of hundred years (McManus et al. 2004). Under typical conditions with strong circulation, the $^{231}\text{Pa}/^{230}\text{Th}$ ratio is relatively low in the North Atlantic, but when circulation shuts down this ratio will increase (McManus et al. 2004). Two of these North Atlantic MOC reductions were recorded during the last deglaciation (McManus et al. 2004). The first occurs during Heinrich Event 1 (H1; $\sim$17.5ka), a massive iceberg discharge event that would have led to a freshening of the surface ocean in areas of NADW formation, and the second during the Younger Dryas ($\sim$12.7ka) (McManus et al. 2004). The H1 shutdown led to a $^{231}\text{Pa}/^{230}\text{Th}$ of $\sim$0.093, indicating a near total shutdown in the North Atlantic MOC, while the Younger Drays $^{231}\text{Pa}/^{230}\text{Th}$ was only $\sim$0.065 indicating a partial shutdown (McManus et al. 2004). These events are separated by warmer Northern Hemisphere conditions and rejuvenated North Atlantic MOC during the Bølling-Allerød interval (McManus et al. 2006). When we look at Southern Hemisphere circulation during this interval, we see opposite effects (Anderson et al. 2009). During H1 and the Younger Dryas, the Southern hemisphere experienced greater overturning, and warmer conditions, separated by a colder AABW reduction during the Antarctic Cold Reversal (concurrent with the Bølling-Allerød period) (Anderson et al. 2009). This antiphase relationship highlights the bipolar seesaw, in which reduced deep water formation in one hemisphere is compensated by enhanced deep water formation in the other hemisphere (Broecker 1998).

2.3 Agulhas Leakage
Another process which has the ability to affect overturning circulation in the Atlantic is the transport of warm, salty Indian Ocean water westward around the tip of Africa, known as the Agulhas leakage. Agulhas leakage is a net input of salt into the Atlantic which can enhance NADW formation in two ways (Weijer et al. 2002). Salinization of water increases its density allowing for greater susceptibility for inclusion into deep water (Weijer et al. 2002). Additionally, Agulhas leakage modifies the overall meridional pressure gradient in the Atlantic (Weijer et al. 2002). This is because the heat and salt input of Agulhas leakage in the South Atlantic largely cancel one another out in terms of buoyancy change, but as this water is advected into the North Atlantic, it loses its heat and retains its salt, increasing the meridional pressure gradient (Weijer et al. 2002; Beal et al. 2011).

Large variations in Agulhas leakage have been observed over glacial-interglacial timescales, and are believed to be controlled by the position of the subtropical front (STF) (Peeters et al. 2004; Bard and Rickaby 2009; Beal et al. 2011; Caley et al. 2012). At present, this front is at about 45˚S, well below the tip of the African continent (37˚S), allowing for the Agulhas current to shed rings into the South Atlantic (Beal et al. 2011). However, during glacial periods this front shifted up to 7˚ northward, greatly reducing the ability of the Indian and Atlantic Oceans to exchange waters (Bard and Rickaby 2009, Beal et al. 2011, Caley et al. 2012). One method for accounting for the amount of Agulhas leakage over time is to use Agulhas leakage fauna (ALF) (Peeters et al. 2004). The Agulhas current sheds water into the Atlantic in the form of Agulhas rings that carry with them a characteristic planktonic faunal assemblage (Peeters et al. 2004). Therefore, by tracking the faunal assemblage of the fossil record in the South Atlantic, it is possible
to qualitatively assess the amount of Agulhas leakage over time (Peeters et al. 2004). ALF is typically low during glacial periods and peaks just prior to glacial terminations and the resumption in MOC, suggesting that this process does indeed play an important role in global climatic cycles and ocean circulation (Peeters et al. 2004, Beal et al. 2011). Expanded sea ice cover and changes in the wind patterns may have also worked to reduce Agulhas leakage during glacial times (Peeters et al. 2004). As conditions shift from glacial to interglacial, the STF migrates south and the gateway for Agulhas leakage is opened (Bard and Rickaby 2009, Caley et al. 2012).

The position of the STF is not the only control on Agulhas leakage, and the strength of Southern Hemisphere wind systems also plays a role (Sebille et al. 2009; Beal et al. 2011). Model simulations have shown that when the Agulhas current is stronger, the retroflection, or the point at which the current turns back into the Indian Ocean, occurs farther northward, and this reduces the overall leakage (Sebille et al. 2009; Beal et al. 2011). Under weaker conditions, the retroflection occurs further south and has less inertia, causing more of the flow to divert into the Atlantic (Sebille et al. 2009; Beal et al. 2011). Thus, while multiple possible forcing mechanisms must be taken into account to interpret past changes in Agulhas leakage, observed changes in Agulhas leakage can help support the interpretation other proxy-observations of past circulation changes.

2.4 Using Uranium as a Redox Record

In the marine environment, aU is useful as it responds to changes in porewater oxygen concentrations (Henderson & Anderson 2003). Uranium is typically found in seawater as soluble U$^{6+}$, is reduced to U$^{4+}$ under suboxic conditions and becomes insoluble (Henderson and Anderson 2003). This authigenic portion of the uranium is an
operationally defined portion of the sedimentary U inventory. This process is microbially mediated and aU has been noted to co-precipitate and adsorb onto iron oxides (Morford et al. 2009) Porewater oxygen concentrations can change for two reasons, either deep water oxygen concentrations decrease or benthic respiration increases as a result of increased organic matter delivery to the sediment (Henderson and Anderson 2003). AABW becomes charged with oxygen during its formation at the surface; therefore, any reduction in AABW formation should result in a decrease in deep water oxygen concentration, and thus porewater oxygen concentrations, leading to an increase in the aU concentration of the sediment (Hayes et al. 2014, Henderson and Anderson 2003).

However, there are some possible complications with aU as a proxy for redox conditions. One major complication is that aU can be remobilized in the sediments following a return to oxidative conditions (Severmann and Thomson 1998). In this study, we monitor the effects of remobilization using $\delta^{234}$U. In this case, $\delta^{234}$U is defined as the per mil deviation of the $^{234}$U/$^{238}$U activity ratio from secular equilibrium when $^{234}$U and $^{238}$U activities are equal ($\delta^{234}$U = $\left[\frac{^{234}U}{^{238}U}\right] - 1$)*1000). The $\delta^{234}$U of marine sediments can be related to uranium remobilization because radiogenic $^{234}$U that is produced in situ is more mobile than $^{238}$U (Henderson & Anderson 2003). This mobility results from the process of $\alpha$-recoil during the $\alpha$-decay of $^{238}$U into $^{234}$U. During this process, $^{234}$U is recoiled ~550 angstroms and is either totally ejected from the mineral or damages its structure, making $^{234}$U more mobile (Henderson and Anderson 2003, Severman and Thomson 1998). Because of the alpha recoil effect, freshwater inputs to the ocean all tend to have positive $\delta^{234}$U values (Tissot et al. 2018). This, in turn, gives seawater a $\delta^{234}$U of ~145‰. Changes to the $\delta^{234}$U of seawater over geologic time may reflect
changes in the rate of continental weathering (Tissot et al. 2018, Henderson 2002). Just as $^{234}\text{U}$ is preferentially weathered from the continents, alpha-recoiled $^{234}\text{U}$ in marine sediment may preferentially remobilize during changes in oxidation state. Preferential migration of $\delta^{234}\text{U}$ has been observed in sediments such as Mediterranean sapropels, sedimentary layers that are rich in organic material and are therefore reducing (Severman and Thomson 1998). In these sapropels, $\delta^{234}\text{U}$ values are greater than expected at the base of aU peaks, possibly pointing to preferential reduction in these locations (Severman and Thomson 1998). Tracking $\delta^{234}\text{U}$ along with aU will help track the extent of U remobilization during past events and give us more confidence in the aU record.

Therefore, from these observations, we can begin to envision what an interglacial AABW reduction might look like. We would expect increased storage of carbon dioxide in the ocean, a reduction in atmospheric carbon dioxide, a reduction in deep ocean oxygen concentrations, and an increase in North Atlantic deep water formation (Sigman et al. 2010, Jaccard et al. 2013, Broecker 1998). In fact, during the MIS 5 AABW reduction (~127ka) these predictions are largely validated (Hayes et al. 2014). Deep ocean oxygen concentrations were reduced as recorded by an increase in aU concentrations not related to increases in export production, atmospheric carbon dioxide concentrations were reduced suggesting greater storage in the deep ocean, and there is an apparent antiphase behavior with deep water formation in the North Atlantic (Galaasen et al. 2014; Hayes et al. 2014). This scenario provides a basis for identifying past interglacial bottom water reductions.
CHAPTER III – MATERIALS AND METHODS

3.1 Materials

The core used to identify the MIS 5 AABW reduction, ODP 1094 (Figure 3.1), was used for this study. The core has demonstrated sensitivity to changes in ocean circulation, with elevated aU concentrations during glacial periods and the noted MIS 5 aU peak (Hayes et al. 2014). ODP 1094 is in a good position to record changes in AABW (53°S, 5°E; 2807 m water depth). Well within the core of CDW, and proximal to the Weddell basin and associated locations of AABW formation, ODP 1094 sits in the ice-free zone of the Antarctic Zone of the Southern Ocean (Shipboard Scientific Party 1999).

Predominantly composed of siliceous sediments, ODP 1094 provides a high-resolution deep sea core, with an average sedimentation rate of ~17 cm/kyr (Shipboard Scientific Party 1999). We analyzed sediment samples from ODP core 1094 within the composite depth range of 28.44 m to 76.29 m corresponding to ages of ~156 kya to ~504 kya. Analyzing material from all marine isotope stages between MIS 11 and MIS 5 help put into context how unique events during MIS11 might have been. Samples are taken at ~10 cm intervals allowing us to construct a record with an age resolution on the scale of hundreds of years.

We also investigated another core in the South Atlantic Antarctic Zone to investigate any corroborating events found in ODP 1094. The core RC13-259 was taken from south of the Antarctic Polar Front, with sediments in ages between MIS 5 and MIS 12 (Charles et al. 1991). Additionally, the sedimentation rate at RC13-259 is much lower than ODP 1094. Even during MIS 11, the period during which RC13-259 experienced the highest sediment accumulation rates, RC13-259 was only accumulating with an average
sedimentation rate of ~4.65 cm/kyr. This slower sediment accumulation leads to a lower age resolution for this core, and also makes the aU in the sediments more susceptible to burn down.

3.2 Methods

Samples from ODP core 1094 were analyzed for both aU concentration and U isotopic composition. Samples corresponding to MIS 11 are from the depths 58.16 m to 69.58 m. Samples have been taken at ~10 cm intervals allowing us to construct an aU record with a resolution on the order of hundreds of years or better as the average sedimentation rate is 17 cm/kyr. This allows us to resolve events on sub-millennial timescales. aU concentrations of the sediments samples were processed by published methods (Hayes et al. 2017) and analyzed on the Thermo Fisher Element XR Inductively Coupled Plasma Mass Spectrometer (ICP-MS) in the Shiller Trace Element Lab at the University of Southern Mississippi.

Subsamples of ~0.1 g of sediment were taken and a known amount of $^{236}$U spike (Eckert & Ziegler Isotope Products) was added to account for any U lost during sample processing. Standards were prepared using a $^{236}$U-spiked uranium standard (CRM112-A). Samples were totally dissolved using nitric, hydrochloric and hydrofluoric acids and hydrogen peroxide. Once totally dissolved, 100 µL of iron chloride solution (~1 M) was added, and iron oxyhydroxide was precipitated by raising the pH of the solution above 8 using NH$_4$OH. This precipitate retains any U present in the samples, while most of the other major elements are left in the overlying solution which can be discarded. The iron precipitate was then dissolved again in hydrochloric acid and the U was isolated with column chromatography using anion exchange resin (AG1-X8). Samples were loaded in
8 M nitric acid and U was eluted with 0.12 M hydrochloric acid. Once the U has been isolated, the samples dried down and taken up in 1.5 mL 2% optima grade HNO₃ solution for ICP-MS analysis.

Using the ICP-MS the amount of $^{234}\text{U}$, $^{235}\text{U}$, and $^{236}\text{U}$ in the samples was measured. Because the most abundant U isotope $^{238}\text{U}$ was not measured, U concentrations were calculated using the seawater $^{238}\text{U}/^{235}\text{U}$ ratio of 137.824 (ratio obtained from Weyer et al. 2008) according to the following equation:

$$U \text{ concentration } (\mu g/g) = \frac{^{235}\text{U}_{\text{Sample}} - ^{235}\text{U}_{\text{Blank}}}{g \text{ sample}} \times \frac{^{238}\text{U}}{^{235}\text{U}}_{\text{SW}} \times 1000$$

For this study, 83 duplicate and 3 triplicate samples were analyzed. The average standard deviation of the U concentrations of these replicates was ~5%. To assess external reproducibility the gravimetrically calibrated solution CRM 112a was used, with a concentration of 1 ppb. Measurements of this solution gave us a precision of ~3%.

In order to determine the aU content of a sediment sample, a correction for the detrital portion of the U was made. Typically, this is done by measuring the amount of $^{232}\text{Th}$, a Th isotope only present in detrital material, and making a correction based on the $^{238}\text{U}/^{232}\text{Th}$ ratio of the detritus (Henderson & Anderson 2003). Measurements of $^{232}\text{Th}$ were made for a subset of the samples analyzed for U in ODP 1094 for this study, and for all samples analyzed from RC13-259. To measure $^{232}\text{Th}$, a known amount of $^{229}\text{Th}$ spike (Eckert and Ziegler Isotope Products) was added to the sample. Additionally, during the column chromatography phase the thorium was eluted first using a solution 12 M trace metal grade HCl (Hayes et al. 2017). With a known concentration of $^{232}\text{Th}$, the detrital U concentration can be calculated using the following equation:
\[ \alpha U = U_{\text{Bulk}} - (^{232}\text{Th} \times 0.2429 \times \frac{U}{T h_{\text{ar}}} \times 1.3297) \]

Where 0.2429 and 1.3297 are constants to convert \(^{232}\text{Th}\) from concentration in \(\mu g/g\) to activity in dpm/g, and U from activity in dpm/g to concentration in \(\mu g/g\), respectively, and \(U/Th_{\text{ar}}\) is an estimation of the U/Th activity ratio of detrital material. A range of estimations are available for \(U/Th_{\text{ar}}\) between 0.4 and 0.7, however a U/Th activity ratio of 0.5 was used in these calculations for this correction to give a conservative estimate of aU using this method. Interglacial portions of the cores analyzed for this study would not be affected much even with a larger estimate of the U/Th activity ratio, as there was very little detrital material in these sections.

Figure 3.1 Map of Core Locations

Map displaying the locations of the two cores analyzed for this study.
Ice-rafted detritus (IRD) is the dominant detrital material in ODP 1094 (Kanfoush et al. 2002). In order to make an estimate of the detrital U in ODP 1094 samples for which \(^{232}\text{Th}\) was not measured, I have made a correlation between the IRD and \(^{232}\text{Th}\) for MIS 5, collected by Hayes et al. 2014 (Figure 3.2). The linear regression equation from this correlation was used to estimate the \(^{232}\text{Th}\) for the samples used in this study. The range of calculated detrital uranium concentrations was 0.0637 µg/g to 0.932 µg/g with an average value of 0.143 µg/g, representing 0.8% to 61% of the total measured uranium, with the higher percentages being representative of glacial-like conditions or a few infrequent points during interglacials. This means that the measured uranium is dominated by authigenic uranium and our estimates of aU are relatively insensitive to the detrital correction, particularly during the interglacial time periods this study analyzed.

![IRD vs \(^{232}\text{Th}\) Activity](image)

**Figure 3.2 IRD vs \(^{232}\text{Th}\) Activity**

Plot displaying the relationship between \(^{232}\text{Th}\) and IRD for MIS5. This relationship was used to approximate the \(^{232}\text{Th}\) activity for the entirety of ODP 1094 using the Kanfoush 2002 IRD record. With \(^{232}\text{Th}\) values calculated detrital U can be estimated.
Using the $^{235}U/^{234}U$ ratio the $\delta^{234}U$ of the samples was calculated using the equation:

$$\delta^{234}U = \left( \frac{1}{^{235}U_{Sample} \times \frac{^{238}U}{^{235}U} SW} \right) \left( \frac{^{238}U_{t_1/2} / ^{234}U_{t_1/2}}{^{234}U_{Sample}} - 1 \right) \times 1000$$

The standard deviation of the $\delta^{234}U$ for the duplicate and triplicate samples in this study was $\sim 5\%$.

However, $\delta^{234}U$ values measured this way are representative of the bulk U. Detrital U can have $\delta^{234}U$ values in the range of $-150 \pm 100\%$ (Bourne et al 2012, DePaolo et al. 2006). Because the calculated values are a combination of both detrital and
authigenic $\delta^{234}$U, it is useful to have a model to compare the results to. A prediction of the $\delta^{234}$U was calculated using a two component mixing model with an age correction:

$$\delta^{234}U_i = 147\%_0 \frac{U_{auth}}{U_{total}} + -150\%_0 \frac{U_{detrital}}{U_{Total}}$$

$$\delta^{234}U_f = \delta^{234}U_i e^{(-\lambda t)}$$

Additional paleoceanographic proxies were used to determine possible causes in the changes in the redox conditions of the sediment. Accounting for changes in productivity was particularly important, as increased benthic respiration can also drive down porewater oxygen concentrations. A high-resolution paleoproductivity proxy exists for ODP-1094 (Jaccard et al. 2013). This productivity proxy is based on the Ba/Fe ratio of the sediment. Barium and organic carbon fluxes have been shown to have a strong correlation (Dymond et al. 1992). By normalizing the Ba to Fe (which is mostly of detrital origin) to account for detrital Ba, this allows for a qualitative assessment of the export production for ODP-1094 (Jaccard et al. 2013). Comparing the aU record to this paleoproductivity proxy allows for the identification of peaks in aU due primarily to AABW circulation reductions rather than productivity changes. For instance, periods of high aU concentrations with relatively low Ba/Fe values would mean that reductive conditions in the sediment could not have been caused by high organic carbon fluxes, therefore making them likely candidates for AABW reductions.

Possible AABW reductions require a mechanism to trigger them. In the case of the MIS 5 AABW reduction, it appears that surface freshening may be a possible trigger (Hayes et al. 2014). This is the case because this reduction occurs during a period of sea level rise driven by glacial meltwater (Hayes et al. 2014). This meltwater would have
strengthened the density stratification in the Southern Ocean leading to reduced AABW formation (Hayes et al. 2014). Surface freshening has also been hypothesized to be the cause of modern reductions in AABW formation (de Lavergne et al. 2014). To determine if AABW reductions during previous interglacial periods were related to increased surface freshening driven by glacial meltwater inputs, the aU reconstruction was compared to sea level reconstructions. The high resolution, multi-proxy sea level reconstruction produced by Spratt & Lisecki 2016 was used for this purpose. However, the Spratt & Lisecki 2016 reconstruction is a global curve, and may not accurately reflect the conditions of the Southern Ocean surface at the time of the reduction.

Other factors that affect AABW formation include increases in NADW formation. An antiphase relationship has been observed in the temperature records of the northern and southern hemispheres during the last deglaciation, suggesting that increases in deep ocean ventilation in one region are accompanied by decreases in the other region (Broecker 1998). Agulhas leakage is believed to play a large role in this process, as the salty water it supplies is advected to the North Atlantic, leading to denser surface water that can be more readily incorporated into NADW (Weijer et al. 2002; Weijer 2002; Knorr and Lohmann 2003). Changes in Agulhas leakage are controlled in large part by the position and strength of Southern Hemisphere westerly winds, with more southerly westerlies contributing more Agulhas leakage (Peeters et al. 2004; Beal et al. 2011; Caley et al. 2012). Meaning that these factors can also have an influence on deep ocean water mass distributions. These factors can be assessed using proxies such as the δ¹³C of benthic foraminifera to assess NADW formation, and Agulhas leakage fauna to assess polar frontal positions.
Multiple studies have utilized the core ODP 1094 (e.g. Jaccard et al. 2013, Hayes et al. 2014, Hasenfratz et al. 2019). As a result, some discrepancies are present in the age-depth relationship between these studies. For this study all the data analyzed from ODP 1094 was aligned with the Hatzenfraz et al. 2019 age-depth model. This was done using the Interp1 function in MATLAB.

For cores other than ODP 1094, a different method was used to correct older age models. In these instances the benthic δ¹⁸O records of the cores were compared to the LR04 global benthic stack, a global compilation of 57 benthic δ¹⁸O records. Major events were then correlated, e.g. the highest values during a glacial period, the lowest values during an interglacial period, the point at which sea level began to fall in the later stages of an interglacial period, and a linear interpolation was used to align the cores to the LR04 benthic stack according to the following equation:

\[ y = y_1 - (x - x_1) \frac{y_2 - y_1}{x_2 - x_1} \]

Where y is the age being calculated, x is its corresponding depth, \( y_1 \) and \( y_2 \) are ages from the LR04 benthic stack assigned to specific depths of the record in question, and \( x_1 \) and \( x_2 \) are the depths with the assigned ages. The reference for the oxygen isotope record of ODP-1094 is Hasenfratz et al. (2019) and that for RC13-259 is Charles et al. (1991). Hasenfratz et al. (2019) presents the first benthic δ¹⁸O record of ODP 1094 and tunes this to the LR04 benthic stack. For this study the core GeoB-3603-2 was also tuned to the LR04 benthic stack when necessary. For RC13-259 no benthic data was available so the planktonic record was tuned to the planktonic record of ODP 1094 when necessary.
Figure 3.4 Age Correlation Example

This figure illustrates the process of adjusting an age model using the LR04 benthic δ¹⁸O stack (Black Line) (Lisiecki and Raymo 2005). Above the reference stack is the original benthic δ¹⁸O record for the core GeoB-3603-2 from Peeters et al. (2004) (Red Line), with the numbers indicating tie points. The adjusted age model (Blue Line) represents the new age model with ages calculated via linear interpolation.

In order to assess the uncertainty in this method the program Undatable was utilized (Lougheed and Obrochta 2019). This allows for an estimation of uncertainty in both the accuracy of the graphical correlation of our tie points combined with a statistical approach for estimating values between tie points. An average uncertainty for the age models used in this study is ~2.4 ka. It is worth noting however that this uncertainty could be reduced simply by assigning more tie points, and that due to relationship of the
benthic $\delta^{18}O$ to sea level the timing of some events must be closely correlated regardless of the numerical ages assigned to them.

To provide more in-depth analysis, the carbon stable isotope record $\delta^{13}C$ from core ODP 1089 was used (Hodell et al. 2003a). This can be a useful in assessing the influence of AABW vs NADW as the two have distinct isotopic values, with NADW being higher due to greater nutrient utilization at its formation sites (Hodell et al. 2003b). So, greater NADW influence is represented by higher $\delta^{13}C$ values, and greater AABW influence is represented by lowed $\delta^{13}C$ values.
CHAPTER IV – RESULTS

4.3 Overview

![Figure 4.1](image)

**Figure 4.1 Overview of the aU record for the past 4 Interglacial Periods**

This figure illustrates the relationship between aU and δ²³⁴U (Hayes et al. 2014 and this study), paleoproductivity (Jaccard et al. 2013), global Sea Level (Spratt and Lisiecki 2016), and atmospheric CO₂ (Lüthi et al. 2008). The orange highlighted areas indicate productivity induced aU enrichments, the yellow highlighted areas show glacial aU enrichments, and the purple highlighted areas show possible circulation induced aU enrichments. Data plotted in blue on the aU and δ²³⁴U records is from Hayes et al. (2014), data plotted in black on these records is from this study.

Plotted above (Figure 4.1) is the aU record developed by this study (~156-500 ka) as well as the record developed by Hayes et al. (2014) (~1.22-158 ka) along with paleoproductivity, global sea level, and atmospheric CO₂ records. This allows for large scale observations of how the aU record at ODP 1094 is related to ocean circulation,
global climate, and Southern Ocean productivity. The ODP 1094 aU displays elevated values during glacial periods (yellow highlighted areas), sea level is low and deep ocean circulation is sluggish leading to increased deep ocean carbon storage, and depleted deep ocean oxygen concentrations. The other major feature of the aU record is the aU peaks associated with periods of high productivity typically during degradations (orange highlighted areas). Increased organic matter delivery to the sediment results in increased respiration in the sediment, and an associated reduction in porewater oxygen concentrations, thus leading to the precipitation of aU. For MIS 7 and 9 these aU peaks are associated with periods characterized by rising sea levels and maximum atmospheric CO₂ values. For MIS 5 the aU record also shows an initial productivity peak, along with a second peak associated with an AABW reduction though difficult to visualize at this scale (Hayes et al 2014). The δ²³⁴U record displays some interesting features, with values elevated with respect to the mixing model expectation during deglacial events, and values lower than expected following them. We interpret this behavior to best reflective of some preferential migration of the ²³⁴U into the sections of core associated with the deglacial aU peaks. However, this migration must only be a certain percentage of the total uranium because the authigenic uranium is never reduced to zero, as would be expected in a typical “burn down” scenario where aU is totally re-oxidized and not present in the sediment. A complete investigation of the δ²³⁴U behavior is beyond the scope of this study. We assume that the measured aU record is not substantially affected by remobilization for the purposes of interpreting past oxygenation events.

We now turn to a closer view of the new observations of each of the past interglacial periods analyzed.
This figure illustrates the relationship between aU, paleoproductivity (Ba/Fe) (Jaccard et al. 2013), global sea level (Spratt and Lisiecki 2016), and atmospheric CO$_2$ (Lüthi et al. 2008) During MIS 7. The orange highlighted area shows a productivity induced aU enrichment, and the purple highlighted area shows a glacial-like enrichment.
The aU record for MIS 7 displays some typical features of interglacial periods, the preceding glacial period shows elevated aU values typical of the sluggish deep ocean circulation during these times that decrease into the interglacial period. In the early stages of MIS 7, as wind driven upwelling delivered nutrients to the surface of the Antarctic Zone of the Southern Ocean productivity increased dramatically (Jaccard et al. 2013). This led to an increase in organic matter delivery to the sediment, and subsequent suboxic conditions in the sedimentary porewater, reflected in elevated aU values during this time (orange highlighted area). However, MIS 7 is peculiar in that it experienced a sea level low during the middle portion of the interglacial, with sea levels and atmospheric CO₂ concentrations more similar to glacial values. During this period, there was also a slight increase in aU (purple highlighted area). This is likely due to sluggish glacial-like conditions during the middle portion of MIS 7, where AABW was more isolated from the atmosphere leading to a greater storage of CO₂ in the deep ocean.
4.5 MIS 9

This figure illustrates the relationship between aU, paleoproductivity (Ba/Fe) (Jaccard et al. 2013), global sea level (Spratt and Lisiecki 2016), and atmospheric CO$_2$ (Lüthi et al. 2008) during MIS 9. The orange highlighted area shows a productivity induced aU enrichment.
The MIS 9 record shows no conclusive evidence of a reduction in AABW. During the time leading into MIS 9, slow deep ocean circulation led to more carbon stored in the deep ocean, thus lower atmospheric CO$_2$ concentrations and lower deep ocean oxygen concentrations, and elevated aU values. Leading into the interglacial increased wind driven upwelling in the Southern Ocean allowed for greater exchange between the deep ocean and the atmosphere. This in turn delivered nutrients to the surface of the Antarctic Zone of the Southern Ocean leading to increased productivity and export production driving the aU increase in the early stages of MIS 9 (orange highlighted area). No other circulation features of note are present in the aU record here, values remained consistently low suggesting no deep water suboxia due to vigorous AABW formation.
This figure illustrates the relationship between aU, paleoproductivity (Ba/Fe) (Jaccard et al. 2013), global sea level (Spratt and Lisiecki 2016), and atmospheric CO$_2$ (Lüthi et al. 2008) during MIS 11. Note the anomalous aU peak at ~395 kry (purple highlighted area). The purple highlighted area shows a possible AABW reduction while the orange highlighted area shows a productivity induced enrichment.
The MIS 11 record displays some features not typical of the other interglacials analyzed for this study. First, the productivity history does not display the same trend as MIS 7 or 9, with large increases leading into the interglacial. Instead, MIS 11 displays more moderate Ba/Fe values for a longer period of time. Second, the period of peak productivity does not occur near the onset of when sea level and atmospheric CO$_2$ are rising, but instead occurs at their maximum values (orange highlighted area). The third feature is a second peak in aU after the period of peak productivity (purple highlighted area ~395ka).
CHAPTER V–DISCUSSION

The ~395ka aU event requires an explanation other than productivity-driven low oxygen. Even though the relationship between aU and productivity is non-linear, it is hard to imagine that the 395 ka event would have higher aU concentrations simply due to export production during a time when paleoproductivity is decreasing. Therefore, the oxygen concentration in the deep water surrounding ODP 1094 must have been lower during this time. Because of this, the aU peak recorded in the later stages of MIS 11 is a likely candidate for an AABW reduction. Further evidence for this hypothesis comes from the CO₂ record. This aU peak occurs during a period of decreasing atmospheric CO₂ concentrations, reflecting increase carbon storage in the deep ocean. Additionally, immediately after the aU peak there is a slight increase in atmospheric CO₂, perhaps due to increased ocean-atmosphere exchange. However, these features occur during a more general decrease in atmospheric CO₂.

Figure 5.1  
MIS 11 Potential
This figure compares the potential depth offset ($z_u$) to the distance between the MIS 11 aU peaks in ODP 1094.

In addition to the $\delta^{234}$U results presented earlier, another analysis was performed to assess whether the aU peaks in late MIS11 could be the result of U diffusion. This is done by calculating the potential depth offset between the sediment water interface and depth of authigenic uranium enrichment using the following equation (Francois et al. 1993):

$$z_u = \frac{D \times U_{sw}}{S \times \rho \times aU}$$

Where $D$ is the diffusivity for UO$_2^{2+}$, $U_{sw}$ is the U concentration of seawater, $S$ is the sedimentation rate, $\rho$ is the dry bulk density, and $aU$ is concentration of authigenic uranium in the sediment. This depth offset, $z_u$, represents the distance over which dissolved U is diffusing between the sediment-water interface and the maximum depth at which authigenic uranium is precipitated in the sediments. As this study also deals with changes in the bottom water oxygen concentration which would affect this depth it is important to note that suboxic bottom water conditions should only lower $z_u$ if they could also be accounted for in this equation. Our assumption is that we could not resolve uranium enrichment events that occur over a distance shorter than $z_u$. Our two aU peaks during MIS11 are 1.86 m apart, whereas the calculated $z_u$ values for this section of the core are ~15-25 cm. Therefore, these two peaks should be representative of the conditions in the bottom water close to the time when they were deposited and are not overprinted by uranium remobilization.
Figure 5.2 *MIS 11 and RC13-259 Comparison*

This figure compares the aU and isotopic record from RC13-259 to the aU record from ODP 1094.

When considering another core from the area, RC13-259, it is more difficult to definitively say that an AABW reduction occurred. The highest aU in RC13-259 occur at the same time as the productivity induced aU peak in ODP 1094, with a similar isotope effect, and there is no second peak. This could possibly mean that RC13-259 did not experience suboxic bottom water conditions. Additionally, RC13-259 has a sedimentation rate much lower than that of ODP 1094 so there is greater potential for some depth/age offset.
This figure illustrates the different responses of the aU records of the cores analyzed by this study during MIS 9 and MIS 11. Additionally noted on the graph is the depth down core of the aU enrichments in both cores for MIS 11 (Jaccard et al. 2013).

Depth offset may be somewhat problematic for RC13-259 as the entire MIS 11 aU enrichment occurs over ~30 cm of core, and while depth offset \((z_u)\) could not be calculated for RC13-259 due to lack of information on sediment density, it seems reasonable that the depth offset could be close to this value. However, further analysis of the aU record of RC13-259 may point to an AABW reduction during MIS 11.
productivity record during MIS 9 shows a much larger productivity spike than at any point during MIS 11. In contrast, the RC13-259 aU record shows much larger aU enrichments during MIS 11, the opposite of the ODP 1094 record. This could possibly mean that there were more reducing conditions due to decreased bottom water oxygen concentrations in MIS11 than in MIS9 and because of the lower sedimentation rate of RC13-259, the shorter term 395 ka event could not be resolved. However, it is also possible that productivity was just simply not as high during MIS 9 locally at site RC13-259. The Ba/Fe paleoproductivity record is from ODP 1094 and the sedimentation rate was lower during MIS 9 than MIS 11 at site RC13-259. To resolve this discrepancy, further study of high sedimentation rate cores covering MIS11 in other areas of the Southern Ocean are warranted.

Taking for granted the observations in ODP1094, if the MIS 11 aU peak is the result of a reduction in AABW, what could have been the trigger? In the modern Southern Ocean and during MIS 5, the trigger appears to be increased density stratification due to surface freshening (de Lavergne et al. 2014; Hayes et al. 2014). However, the MIS 11 event (purple highlighted area in Figure 4.4) occurs during a period of falling global sea level. Because the sea level record used is a global curve it may not be truly representative of conditions in the Southern Ocean at this time, so the surface freshening hypothesis cannot be totally excluded, but another mechanism to trigger a shutdown would help support the AABW reduction hypothesis.
Another peculiar feature of MIS 11 is its record of Agulhas Leakage Fauna (ALF). The ALF record is typically used as a proxy for the position of the subtropical front (STF), with more ALF meaning a more southerly front (Peeters et al. 2004). As shown in the figure below, ALF tends to be highest at the onset of interglacial periods, where the STF migrates allowing for the introduction of warm, salty water from the Indian ocean to the Atlantic Ocean, which is believed to play a critical role in jump starting AMOC at the onset of interglacial periods (Peeters et al. 2004).
Figure 5.5 ALF Record for the Four Previous Interglacial Periods

This figure shows the ALF record for the past 500 kyr (Peeters et al. 2004) with their comparison to global sea level (Spratt and Lisiecki 2016). This illustrates the different response of the MIS 11 ALF compared to the other interglacial periods.

The record of ALF during MIS 11 is somewhat peculiar. While there is a drastic increase in ALF at the onset of MIS 11 peak ALF values are not seen until late in the interglacial stage, and actually occur during a period when sea level is falling. It seems unlikely that the STF would be farthest south during this time, so another mechanism must be at work. It has been suggested that this increase in ALF could be the result of an increase in the strength of the Agulhas Current driven by stronger trade winds (Dickson et al. 2010). However, some modelling evidence suggests that a stronger Agulhas Current would lead to less Agulhas leakage (Sebille et al. 2009). Additionally, if the wind systems in the southern hemisphere were stronger we would expect more upwelling in the Southern Ocean and associated deep water formation. This also seems unlikely for the 2nd MIS 11 aU peak as productivity, which is in large part representative of upwelling in
this region of the Southern Ocean, is decreasing, and deep water oxygen concentrations must have been low to precipitate the aU.

Figure 5.6 Typical glacial and Deglacial Properties compared to MIS 11
This figure compares the observations seen at the end of MIS 11 to typical glacial and deglacial observations. Where low Agulhas leakage, upwelling, and deepwater oxygen are noted during glacial periods. High Agulhas leakage, upwelling, and deepwater oxygen are noted during deglacial events. The ~395 ka event is odd in that it has high Agulhas leakage but low productivity and deepwater oxygen.

Figure 5.6 illustrates the peculiarities of late MIS 11, where the Agulhas leakage does not fit with the observations of either glacial or deglacial periods. These factors combined point to a weaker westerly wind system following peak productivity in the Southern Ocean, allowing for enhanced Aghulas leakage, low productivity, and low oxygen concentrations. This could be due to a decrease in the strength of the westerly winds, or they could have shifted to a more northerly position decreasing the energy input over the Southern Ocean (Durgadoo et al. 2013). This decrease in energy input would result in less upwelling and associated deep water formation, thus lowering deep water
oxygen concentrations. There is some modelling evidence to support the idea that an equatorward shift in the westerlies could result in increased ALF (Durgadoo et al. 2013). However, the consensus tends to be that stronger southward shifted westerlies leads to more Agulhas leakage (Peeters et al. 2004; Biastoch et al. 2009; Durgadoo et al. 2013). The observations of this study point to some other mechanism being at work at the end of MIS 11 that seem to support the idea that a weaker current or northward shifted winds could result in increased Agulhas leakage. Paleoproduction proxies have been used to infer information about wind driven upwelling in the Antarctic Zone of the Southern Ocean (Anderson et al. 2009). Therefore, the corresponding decrease in Ba/Fe seen at during the ~395 ka event could possibly be a result of weakened winds during this time. However, this is representative of upwelling and not necessarily wind strength. Surface freshening could also act to prevent upwelling with no change in wind strength or position. Additionally, a decrease in wind strength or northward shift may allow a freshwater lid to persist around Antarctica similar what is hypothesized to occur when winds are shifted northward during glacial periods (Sigman et al. 2010).
Figure 5.7 Possible Scenarios for the ~395 ka Event

Figure A represents the typical situation for wind driven upwelling and associated AABW formation in the Southern Ocean, Figure B represents what a situation with weaker westerly winds would result in. With decreased upwelling leading to decreased AABW formation and low oxygen concentrations in the bottom water. Figure C represents what a northward shifted wind scenario would look like with reduced AABW formation and low oxygen concentrations in the bottom water. Both situations B and C could potentially explain the ~395 ka event with high Agulhas leakage and low AABW.

Because Agulhas leakage supplies salty water to the Atlantic, increased Agulhas leakage should feedback into the MOC by way of increased NADW formation (Knorr and Lohmann 2003). NADW carries with it a high $\delta^{13}C$ signature, due to its low
preformed nutrient content, so we also might expect to see increased NADW input into the Southern Ocean during this time, which also has a lower oxygen concentration than AABW in this region, and could be a factor in the elevated aU concentrations near the end of MIS 11.

Figure 5.8 MIS 11 ALF and $\delta^{13}C$

This figure shows the relationship between ALF at GeoB-3603-2 (Peeters et al. 2004) and $\delta^{13}C$ 1089 (Hodell et al. 2003a) a proxy for NADW influence in the Southern Ocean.

In the case of the $\delta^{13}C$ data from ODP 1089 and the ALF data from GeoB-3603-2, there seems to be a period where $\delta^{13}C$ is high, although not necessarily increasing during most of the period with high ALF. While this does not necessarily confirm the hypothesis, it does somewhat fit with the prediction (Hodell et al. 2003a; Peeters et al. 2004).
CHAPTER VI– CONCLUSIONS

During the later stages of the interglacial period MIS 11, an abrupt low bottom water oxygen event occurred that suggests a reduction in AABW from ~395 ka to ~398 ka. This is the second such interglacial AABW reduction found in the aU record of the deep sea marine sediment core ODP 1094, with the other reduction (~127 ka) occurring during MIS 5e (Hayes et al. 2014). While both events are represented by low bottom water oxygen conditions though they appear to have different driving mechanisms. The MIS 5e event being caused by a decrease in Southern Ocean surface density, and the MIS 11 event being driven by a change in the intensity or location of the southern hemisphere westerly winds. As the earth continues to experience warming temperatures and rising sea levels a continued decrease in Southern Ocean surface density is expected (de Lavergne et al. 2014), and the southern hemisphere westerlies are expected to move poleward and intensify (Russell et al. 2006). In fact, this increase in the intensity of the westerlies is hypothesized by some modelling evidence to overcome the increase in density stratification (Russell et al. 2006). However, many of the trends we see related to the modern Southern Hemisphere’s climate are believed to be related to ozone loss (Thompson and Solomon 2002). So these past events may not be exactly analogous to the modern climate. This still raises future research questions if we are to fully understand earth’s past climate:

Is the MIS 11 AABW reduction truly driven by some change in the southern hemisphere westerlies, and if so what kind?

What could cause the westerlies to move or weaken so abruptly?
Was this same mechanism at work during MIS 5, working in concert with Southern Ocean surface freshening?

It is important that we understand the mechanisms at work during these past interglacial periods, and particularly MIS 11 as it has orbital parameters similar to the Holocene and higher-than-present sea levels, as it may give some insight into future climatic conditions.
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