Holocene History of the East Antarctic Ice Sheet Environmental Magnetic Record from Mac. Robertson Land

Kenneth R. Kacperowski Jr.
Abstract

A 24-m jumbo-piston core containing a 14 ka Holocene-Late Pleistocene sedimentary record was collected from Mac.Robertson Land during United States Antarctic Program cruise NBP01-01. This study uses environmental magnetism to trace the Holocene history of the East Antarctic Ice Sheet (EAIS), which is generally thought of as stable since it is grounded above sea level. Magnetic analyses are used to identify periods of terrigenous sedimentation from the EAIS, for example ice rafted debris layers or meltwater pulses. Terrigenous material contains magnetic minerals, which we expect to stand out from the dominantly biosiliceous sediment deposited along the East Antarctic Margin. Magnetic susceptibility is weak down core (approximately $-2$ to $5 \times 10^{-8}$ m$^3$/kg) until approximately 10.0 ka where there is a steep increase in intensity, most likely due to sandy mud laminae and underlying diamict. Analysis of X-rays down core show periods of high and low frequency of laminations down core, until a varved layer stratigraphically above the diamict (5-16 laminations per 5 cm). Gravel grains are sparsely present down core until the diamict, where they are ubiquitous. These changes in lithology represent deglaciation following the last glacial maximum and the transition to an open marine environment at 14 ka. Based on ARM and SIRM data, the upper core (0.8-7.7 ka) exhibits a low abundance of magnetic material until approximately 7.7 ka, where there is an increase in fine magnetic material above the background until 10.4 ka. Magnetic assemblages for the upper core (approximately 0.8-9.0 ka) likely consist of magnetite and titanomagnetite based upon SIRM/$\chi$ and S-ratio values of 0.1 – 12 kA/m and 0.9-0.95, respectively (Evans and Heller, 1999). The assemblage changes in the lower core (prior to 9.0 ka) with isolated peaks in SIRM/$\chi$ suggesting magnetic iron sulfites (greigite and pyrrhotite) and also the presence of high coercivity minerals such as hematite and goethite. From 0.8 to 4.4 ka, ARM exhibits a zone of regularly spaced peaks occurring on a ~ 300-700 year scale, which is a cycle typically seen in the Antarctic Peninsula during the late Holocene (e.g., Domack et al., 2001 and references therein) and interpreted as a productivity signal. A magnetic-rich interval above the background ARM and SIRM levels is seen between 7.7-10.4 ka. This period partially coincides with an early Holocene warm period seen on the Antarctic Peninsula (Leventer et al. 2002). The magnetic enrichment could have been caused by increased terrigenous input through meltwater as the nearby ice sheet retreated. This has been observed on the Antarctic Peninsula and possibly other locations such as the Canadian Arctic, Chukchi-Alaskan margin, and other western Antarctic regions (Brachfeld et al., 2009; Leventer et al., 2002; Roberts et al., 2004; Vare et al., 2009). This promotes the idea that the Holocene Thermal Maximum is likely a global occurrence, rather than a local event.
Holocene history of the East Antarctic Ice Sheet: Environmental magnetic record from Mac. Robertson Land

by

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1. Introduction

The inner Nielsen Basin was studied during United States Antarctic Program cruise 01-01 of the *Research Vessel Ice Breaker Nathaniel B. Palmer* (hereafter NBP01-01). Core sites were chosen based on previous surveys of the Mac.Robertson continental shelf led by P.T. Harris and P.E. O'Brien from the Australian Geological Survey Organization (AGSO) in conjunction with the Australian Antarctic Division (see Harris and O'Brien, 1996, 1998). AGSO took several short sediment cores on the continental shelf, including the Nielsen Basin (Fig.1), which recovered approximately 2-3 m of sediment, the maximum length possible on their research vessel. The *RVIB Palmer* utilizes a 25 m jumbo piston core, which can recover longer records than the previous researchers were able to achieve. Two companion cores were taken in the immediate area, designated Jumbo Piston Core 39 and Jumbo Piston Core 40 (JPC39 and JPC40). JPC39 targeted thicker sediment cover over a strong reflector seen in the seismic data. Due to an increased sedimentation rate in the area of JPC39, the core sacrificed collecting the entire Holocene record for higher resolution down core. JPC40 was taken where the sedimentation rate was lower, enabling this core to collect the entire Holocene record, although exhibiting a lower resolution. NBP01-01 JPC40 was ultimately chosen for this study due to the compromise of substantial sediment cover at the core site (~ 24 m) and the presence of the entire Holocene record.

Paleomagnetic analyses have proven to be a cost-effective, rapid reconnaissance method to analyze sediment cores and determine sequences of events in glacial marine environments (e.g. Stoner and Andrews, 1999; Sagnotti et al., 2001; Brachfeld et al., 2002; Rousse et al., 2006). Magnetic minerals can be used as tracers of terrigenous input
in the sediment core. Terrigenous material in the core would be due to ice rafting or meltwater pulses, resulting in deposition of terrigenous material into the adjacent oceanic basin (Fig. 2). The study area is an open ocean environment characterized predominantly by silica-rich diatomaceous sediment. Iron oxides, derived from continental bedrock, are ferromagnetic, whereas the biogenic silica background is diamagnetic. Thus magnetizations found down core could be attributed to terrigenous input.

Specific suites of parameters are utilized to target concentration, grain size, and mineralogy down core. Magnetic susceptibility ($\chi$) has been analyzed to give an overall magnetic signature of the core. This parameter assesses the degree of magnetization of a material in the presence of a weak applied magnetic field. This technique analyzes all material found in each sample, not only ferromagnetic, but also paramagnetic and diamagnetic material. Magnetic susceptibility is a rapid method of determining the concentration of iron-bearing minerals down core. Anhysteretic Remanent Magnetization (ARM) is used to analyze the fine-grained magnetic particles (< 10 microns assuming magnetite is the carrier, Dunlop and Özdemir, 1997) in the sediment core. High abundance of fine grained magnetic minerals could result from meltwater pulses that carry terrigenous sediment in suspension. Saturation Isothermal Remanent Magnetization (SIRM) applies an extremely high field (1 Tesla) to the samples in an attempt to magnetize all of the particles present, regardless of grain size. In addition, the high applied field can partially magnetize high-coercivity minerals such as hematite and goethite, which are not significant contributors to $\chi$ and ARM.
Figure 1a. Location and morphology of the Mac.Robertson shelf and Nielsen Basin (red box) (Sedwick et al., 2001).
Figure 1b. Swath bathymetry map and location of core JPC40 in Nielsen Basin (Leventer et al., 2006).
Figure 2. A schematic of sedimentation pathways and processes associated with Mac.Robertson Shelf (Harris and O'Brien, 1996).
SIRM is an indicator of the concentration of magnetic material. Unlike susceptibility, which activates diamagnetic, paramagnetic and ferromagnetic minerals, SIRM targets only the ferromagnetic minerals in the core that carry a remanence. SIRM is also used to normalize ARM and susceptibility to determine magnetic grain size variations down core. Dominant grains sizes can suggest sediment pathway processes (i.e. ice rafted debris and meltwater pulses), and the energy of transportation mechanisms (i.e. current strength).

Magnetic mineral composition can be used to trace where the particles originated and identify zones of iron-sulfur diagenesis. SIRM/\( \chi \) can show the presence of magnetic iron sulfides such as greigite and pyrrhotite if the ratio exceeds values of 20 kA/m (Maher and Thompson, 1999). The presence of these minerals would indicate an environment where magnetite is dissolving, and iron combines with sulfate in seawater forming iron sulfides as a result. S-ratio analysis can be utilized to distinguish between low-coercivity minerals such (titano)magnetite and high coercivity minerals such as hematite and goethite. This is done by comparing ratios of SIRM acquired in different applied field strengths.

2. Study Area

2.1 Mac. Robertson Land Shelf

Mac.Robertson shelf lies adjacent to Mac.Robertson Land, located on the continental margin of East Antarctica (Fig. 1a). The shelf has a mean depth of 150 m and extends approximately 400 km until the shelf break, which occurs at 350 m depth (Harris and O'Brien, 1996). The shelf break is typically significantly deeper in the Antarctic region due to ice-loading and isostatic depression of the continent (Johnson et al, 1982;
Anderson and Molnia 1989). Harris and O’Brien (1996) described 4 physiographic provinces on the Mac. Robertson shelf: (1) high relief ridge and valley topography with deep inner shelf basins; (2) U-shaped depositional valleys; (3) smooth, planed bank tops in 100-200 m water depth; and (4) iceberg-gouged and current reworked outer shelf and upper slope (Fig. 3). The inner shelf consists of generally flat and featureless banks separated by deep basins, typically up to 1200 m deep in some areas (Harris and O’Brien, 1996). The basins are steep-sided and U-shaped, thought to have been formed by faulting on the shelf compounded by glacial erosion (O’Brien et al 1994). It is also possible that these basins could have been eroded by glaciers and/or ice streams when the EAIS was much larger in the past.

Nielsen Basin is the focus of this study and is located on the Mac.Robertson shelf between Iceberg Alley to the west and Burton Basin to the east (Fig. 1). Nielsen Basin shows evidence of iceberg grounding and plough marks. These features were evident mostly on the outer shelf and upper slope region. Ice rafted debris (IRD) was observed to be fairly ubiquitous throughout the entire system (Harris and O’Brien, 1996). Surficial sediments in Nielsen Basin have been described by Quilty (1985) as dominantly mud with a diverse siliceous skeletal component (Fig. 3). Quilty (1985) also described surficial sediments from the upper slope and low relief bank tops to be coarse biogenic and terrigenous sediment. The abundance of biogenic siliceous material is due to high productivity in an open marine environment during the Holocene after the last glacial maximum and ice sheet retreat. There are presently no major glaciers terminating on the Mac.Robertson Shelf (Jezek, 2008). There are two glaciers that are adjacent to
MacRobertson Land and shelf, most notably, Lambert Glacier/Amery Ice Shelf to the east and a small unnamed glacier to the west (Fig. 4).
Figure 3. Plot of large-scale features and surface sediment grain size distribution on the Mac.Robertson Shelf (Harris and O’Brien, 1998).
Figure 4. Ice flow velocity model of the Antarctic Ice Sheet depicting the location of major fast-moving glaciers, ice shelves, and ice streams (Jezek, 2008). The boxes highlight the two glaciers adjacent to Mac.Robertson Land (Orange = Lambert Glacier/Amery Ice Shelf, Yellow = unnamed glacier).
2.2 Oceanography

Mac. Robertson shelf is subjected to the Atlantic Coastal Current system, which flows west along the shelf (down-flow from Prydz Bay and Amery Ice Shelf) (Smith et al., 1984) exhibiting hourly averaged near-bed velocities of 0.49 m/s and maxima reaching 1.96 m/s (Harris and O'Brien, 1998). This current has been classified as proficient in transporting and eroding gravel-sized particles (Harris and O'Brien, 1998). Harris and O'Brien (1998) measured the currents directly by deploying two Anderaa self-recording RCM-7 current meters on the outer shelf at 375 m and 630 m depth (Harris and O'Brien, 1998). This set up recorded on an hourly and 20-minute interval approximately 1-m above the seabed (Harris and O'Brien, 1998). Harris and O'Brien (1996) noted that surface sediment on the outer shelf and upper slope contained more coarse sand and gravel then inner shelf deposits, most likely due to fine particles being winnowed away by strong bottom currents. They also noted the presence of large three-dimensional dune structures observed in side-scan SONAR images, which are all indicative of strong bottom currents. The minimal impact of swell waves and tidal currents on shelf sediments due to yearly sea ice cover denotes another dynamic behind these findings (Harris and O'Brien, 1996). Due to the formation of sea ice in the region during winter, cold saline water sinks, transporting sediments to the shelf bottom as well.
3. Methods

3.1 Field Methods

JPC40 was collected in 750 m water depth in Nielsen Basin and recovered 24 m of sediment. The core was brought aboard the ship, where it underwent several field analyses. Volume-normalized magnetic susceptibility (k) was measured on the whole core (before it was split) using a Bartington MS2C 13-cm diameter sensor. The core was sent to the Antarctic Marine Geology Research Facility (AMGRF) at Florida State University for further analysis and curation. The core was split at AMGRF and the faces scraped in order to remove any smear layer and present the clean core surface. The core was then color photographed in 30-50 cm increments to capture the immediate state of the core profile. The core was then described fully, noting characteristics such as grain size, color, sedimentary structures, fossils, and other significant features. The split core was analyzed at AMGRF using a Geotek multisensor track. Magnetic susceptibility and bulk density were measured at 2-cm intervals.

3.2 X-Ray Analysis

A u-channel sub sample, an oriented 2 x 2 x 150 cm plastic tube, was collected down the center of the split core face. These u-channels were X-rayed at AMGRF to obtain a radiographic image of the core. This is conducted to capture laminations, gravel grains, and other stratigraphic features that may be difficult to identify from the core face only. In some instances sediment color may fade over time, making it difficult to identify laminations if the core is not analyzed immediately. Gravel grains and shells below the surface of the core face are only visible in x-rays. The radiographic images were analyzed
on a light table to obtain the frequency of laminations and gravel-sized grains per 5 cm intervals down through the core. However, due to the narrow area of the u-channel, gravel grains and IRD frequency may be underrepresented down core. The core log and photographs were also used in conjunction with the radiographic images to improve the accuracy in identifying laminations. The frequencies were logged and plotted versus depth.

3.3 Rock-magnetic samples

Sub samples were collected every 2.5 cm from the core and freeze-dried in a Labconco freeze dryer. Approximately 100-300 milligrams of freeze-dried sediment was packed into gelatin capsules for magnetic analyses. The gelatin capsules were inserted into plastic straws, which are convenient non-magnetic sample holders.

3.4 Magnetic Susceptibility

Volume-normalized magnetic susceptibility was initially analyzed aboard the ship and again at AMGRF using a Bartington MS2C 13-cm diameter loop. The results gathered in both cases were problematic due to the plastic core liner exhibiting a stronger signal than the majority of the sediment. Due to these circumstances, mass-normalized magnetic susceptibility ($\chi$) was conducted on freeze-dried sediment at Montclair State University to assess the degree of magnetization induced by an applied magnetic field on the sample. An AGICO KLY4 Kappabridge Unit was used to gather the data. The Kappabridge was calibrated using an Etalon standard provided by the manufacturer. An empty sample holder value was obtained by measuring an empty gelatin capsule and
plastic straw before each session. The empty sample holder value was subtracted from the sample measurements. Each individual, pre-weighted, sub sample was placed into the well of the Kappabridge where it was analyzed 3 times consecutively, yielding 3 measurements for increased accuracy in results. In some cases, a 4th measurement was required when the initial 3 measurements varied more than expected. These three measurements were averaged, and the average value was normalized by mass in kilograms to yield mass-normalized bulk susceptibility in units of m³/kg.

3.5 Anhysteretic Remanent Magnetization

Anhysteretic Remanent Magnetization (ARM), Saturation Isothermal Remanent Magnetization Acquisition (SIRM Acquisition), and S-ratios analyses were all conducted at the Paleomagnetics Laboratory at Lamont-Doherty Earth Observatory, Columbia University, New York. ARM is generated by subjecting the samples to a steady weak field (on the order of the Earth’s magnetic field), while simultaneously applying an AC field (“alternating field”) that uniformly decays over time. The decaying AC field, on its own, is a demagnetization process that slowly erases the remanent magnetization of the particles present. However, since single domain particles (~0.05-1.0µm for magnetite) have a high coercivity, the demagnetization has less effect on single domain particles, and a greater effect on multidomain particles. The combined AC and steady field preferentially magnetizes the single domain and pseudo single domain particles parallel to the steady field. The Earth’s magnetic field was measured to be approximately 47.35 µT with an inclination of 71.12° and declination close to North in the lab where ARM was imparted. The GE Inductrol Voltage Regulator Control was first set to 4.6 A on the
variac, which sets the peak alternating field to approximately 100 mT. Bundles of samples were centered inside the field coil and the GE Inductrol Voltage Regulator Control was set to "lower" to initiate the field decay. The GE Inductrol Voltage Regulator Control alarm sounds when the field has decayed to zero and the samples were removed from the coil and the variac was returned to the zero position. The machine was switched to the "raised" position and then recharged. Once recharged, the procedure was repeated for all samples down core.

After all the samples were imparted with ARM, they were measured on a 2G Cryogenic Superconducting Rock Magnetometer (SRM). The superconducting rock magnetometer is used in conjunction with a computer interface run on a local computer system. For all measurements, an empty holder measurement was conducted before each session to determine the remnant magnetization intensity of the holder itself. This value is subtracted from each sample’s remnant magnetization intensity. The samples were placed individually on to the instrument’s sliding holder arm, which inserts and removes the samples before and after measurements are taken. The sliding holder arm ensures the samples are centered inside the magnetometer’s superconducting quantum interference devices (SQUIDs). Each sample is labeled by its depth in centimeters on the computer interface before any measurements. Once this is completed, a background reading is first initiated via keystroke on the computer interface. Following the initial background reading, the sample is inserted and centered slowly into the SRM on the sliding holder arm. The SQUID read-outs were monitored after insertion of the sample until they stabilized in each of the axial directions (x, y, and z). A keystroke initiates the computer interface to conduct 4 sequential readings, each measuring the intensity of magnetization
in the x, y, and z directions for each sample. Once completed, the sliding holder arm is slowly removed from inside the SRM. At this point, the computer interface will prompt the user to complete a post-measurement background reading. Once this is completed the computer interface will internally average the 4 readings, subtract the background, and convert the average x, y, and z parameters to intensity of remanence (J), inclination (I), and declination (D) for each sample run. These are compiled in a table, which can be retrieved from the computer’s hard drive as a text file. Since ARM and SIRM are laboratory-induced remanences, the I and D directional values are not used. ARM Susceptibility ($\chi_{\text{ARM}}$) is calculated by dividing ARM intensity ($\text{Am}^2/\text{kg}$) by the strength of the bias field (37.68 A/m). The ratio $\chi_{\text{ARM}}/\chi$ is calculated by dividing ARM susceptibility by bulk susceptibility measured on the Kappabridge. A higher ratio results from finer single domain particles that exhibit high ARM remanence, but low magnetic susceptibility. A lower ratio would mean coarser pseudo-single and multi-domain particles that exhibit low ARM but a high susceptibility. Thus, this ratio is good for identifying magnetic particle size changes down core.

3.6 Isothermal Remanent Magnetization Acquisition

Stepwise increasing DC fields generated in an impulse magnetizer were used to generate an isothermal remanent magnetization (IRM) acquisition curve. This is done to determine the appropriate field intensity need to impart Saturation Isothermal Remanent Magnetization (SIRM). The saturation field is also indicative of which minerals may be present in the sample. Pilot samples were subjected to stepwise increasing treatments of 0, 30, 60, 80, 160, 220, 320, 490, 650, 820, 980, 1140, 1220, and 1290 mT, until the
sample's remnant magnetization intensity no longer increases, i.e. "saturates." On a plot, this would be seen as remnant magnetization intensity plateauing once saturation is reached. This was conducted using an ASC Model IM-10-30 Impulse Magnetizer, which charges and deploys a strong but short-lived magnetic field whose intensity and direction (+/-) is determined by the user. Pilot samples were selected approximately every 150 cm down core yielding 12 samples for the procedure. The 12 samples were magnetized for the lowest treatment (0 volts = 0.0T) and then their remnant magnetization intensities are measured in the cryogenic magnetometer. This is repeated for each of the treatment levels until the maximum treatment (400 volts = 1290 mT = 1.29 T). The data were then plotted as remnant magnetization intensity versus applied field to determine the point of saturation and select the appropriate field intensity to impart SIRM.

3.7 Saturation Isothermal Remanent Magnetization

Once IRM acquisition curves were determined, SIRM was imparted to the samples down core. Small bundles of samples were placed in the impulse magnetizer, where the necessary voltage was set via a dial. In this case, 350V was programmed to deliver a 1.1 T field to magnetize the sediment. Once the read-out on the impulse magnetizer reached the desired charge, the field was deployed by pressing the "trigger" button. Imparting the 350V charge was repeated twice on every bundle to ensure full magnetization was achieved. The samples were then inserted individually into the cryogenic magnetometer to measure their remnant magnetization.
3.8 S-Ratio

S-ratios is defined here as:

\[ S = \left| \frac{M_{R(300\,\text{mT})}}{M_{R(1\,\text{T})}} \right| \]

where \( M_{R(300\,\text{mT})} \) is the remanent magnetization acquired in a -300 mT applied field and \( M_{R(1\,\text{T})} \) is the remanent magnetization acquired in a 1 T applied field. S-ratios down core are obtained by looking at the relationship between the intensities of IRM of each sample after first applying a 1 T field, measuring the remanence \( (M_{R(1T)}) \), then applying a -300 mT back field to induce a remanence in the opposite direction \( (M_{R(-300\,\text{mT})}) \). Magnetite and titanomagnetite generally saturate below 300 mT, while hematite and goethite saturate above 2 T (Dunlop and Özdemir, 1997). S-ratio values near or at 1 indicate the presence of magnetite and titanomagnetite, for which the -300 mT back field completely remagnetizes the sample, and \( M_{R(-300\,\text{mT})} \) is equal to \( M_{R(1T)} \). A value < 1 signifies that some magnetic particles that are magnetized in the positive (+) direction are not affected by the -300 mT field, i.e., only the lower coercivity magnetic particles become magnetized in the opposite direction. Minerals like hematite and goethite have high coercivity and remain magnetized in the positive direction. Therefore, the remanent magnetization in the two directions are unequal, driving the ratio to be < 1. Therefore, S-ratio values, such as 0.80-0.85, indicate that presence of high coercivity minerals. To obtain the intensities for each sample in a -300 mT field, the coil in the impulse magnetizer must be removed, rotated 180°, and reinserted before the field can be applied to the samples. By doing this the applied field will now be applied 180° in the opposite direction. Following each of these two procedures, the samples were then inserted
individually into the cryogenic magnetometer and analyzed for their remnant magnetization intensities and recorded.

The order of the analyses described in sections 3.1 – 3.8 is significant due to the intensity of the magnetic field applied to the samples. SIRM and S-ratio analyses utilize a much stronger field. Remanences acquired at these field levels cannot be easily erased, if at all. It is because of this logic that the analyses were conducted in the order described here, from weakest to strongest applied field.

3.9 Time series Analysis

All resulting data from each parameter was analyzed for periodicity using ION Research Systems Interactive Wavelet Computing Program (http://ion.researchsystems.com/IONScript/wavelet/). The data on the age scale was first resampled into evenly spaced data points using Analyseries software (Paillard, 1996) for input into the ION Interactive Wavelet Program. A cone of influence is applied to the resulting power spectrum and then is subsequently analyzed for periodicity based on intensity of power. The wavelet data results can be seen in the Results section and in the Appendix.

4. Chronology

JPC40 was sampled for bulk sediment $^{14}$C dating. Twelve (12) sediment sub-samples were taken approximately every 200 cm down core. During subsampling, sediment removed from the core was thoroughly examined for calcite shells, such as bivalves or foraminifera. If no calcite macrofossils were found, bulk sediment was
transferred into a 4-oz glass jar. The samples were dried overnight at 60°C to remove any water in the sediment. The conversion of bulk sediment samples to graphite targets for radiocarbon dating was done by Dr. Robert Dunbar and colleagues at Stanford University. The samples are then treated in a 1 N HCl bath at 90°C to dissolve any calcite microfossils or detrital calcite (Rowe et al, 2003). This is to ensure only the organic carbon from the sample was contributing to the age derived. The decalcified samples were then rinsed with deionized water until neutral (Rowe et al, 2003). This fraction is then reacted with 1 N NaOH to separate out the humin and humate fractions. The humin fraction is subjected to the acid bath a second time, while humate fraction was precipitated with concentrate HCl. All fractions were subsequently rinsed 3 times, vacuum-dried, combusted, and graphitized (Rowe et al, 2003). The prepared radiocarbon samples were then sent to Lawrence Livermore National Laboratory (LLNL) to be radiocarbon dated using an accelerator mass spectrometer (AMS).

An age-depth equation was formulated from 9 dated sub samples plotted down core (Table 1 & Fig. 5). Three of the uncorrected ages at depths 2006 cm, 2051 cm, and 2211 cm (15177 yrs, 20564 yrs, and 30379 respectively) were rejected and not included in the model based on the carbon/nitrogen ratio data (Costa et al., 2008), which suggested probable influence of terrestrial carbon in these samples. The C/N ratio values range from 4.0-8.9 down core until the silty-clay/diamict where the values increase to 9.0-13.5. Multiple best-fit curves were applied to these data points including linear, exponential, and 2nd-order polynomial. A 3rd-order polynomial was ultimately selected due to its fit and R² value of 0.9995. The age-depth model from the plot is the following:

\[ y = 2E-07x^3 + 0.0016x^2 + 1.9683x + 472.65 \]
where $y$ is the age in years and $x$ is the depth in centimeters. The age model was subsequently applied down core until the mud-diamict contact, which occurs at approximately 2170 cm depth.
Table 1. $^{14}$C Data table for JPC40. $^{14}$C dates by depth, with raw and calibrated ages.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Raw $^{14}$C Age</th>
<th>Calibrated Age (yrs)$^1$</th>
<th>Carbon Source$^2$</th>
</tr>
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<tbody>
<tr>
<td>101</td>
<td>2400 ± 35</td>
<td>687 ± 126</td>
<td>AIOM</td>
</tr>
<tr>
<td>201</td>
<td>2690 ± 35</td>
<td>938 ± 142</td>
<td>AIOM</td>
</tr>
<tr>
<td>401</td>
<td>3225 ± 40</td>
<td>1493 ± 155</td>
<td>AIOM</td>
</tr>
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<td>AIOM</td>
</tr>
<tr>
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<td>4480 ± 35</td>
<td>2993 ± 169</td>
<td>AIOM</td>
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<tr>
<td>1001</td>
<td>5555 ± 35</td>
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<td>2006</td>
<td>14545 ± 40</td>
<td>15177 ± 237*</td>
<td>AIOM</td>
</tr>
<tr>
<td>2051</td>
<td>19120 ± 60</td>
<td>20564 ± 188*</td>
<td>AIOM</td>
</tr>
<tr>
<td>2119</td>
<td>13895 ± 40</td>
<td>14119 ± 230</td>
<td>AIOM</td>
</tr>
<tr>
<td>2211</td>
<td>25230 ± 110</td>
<td>30379 ± 275*</td>
<td>AIOM</td>
</tr>
</tbody>
</table>

$^1$Calibrated using CALIB data set, using $\Delta R=1400$ yrs and a built in $R=400$ yrs. Total local reservoir age is 1800 yrs.

$^2$Carbon source is defined as Acid Insoluble Organic Matter (AIOM).

$^3$This date was calibrated using the marine calibration curve of Fairbanks et al., 2005.

$^*$These ages have been rejected and not included in the final age model.

**Age Model-Core JPC40**

![Image of age model graph](image-url)

Figure 5. Age model for JPC40. Equation of the trend line shown with $R^2$ value. Points designated as an "X" are ages that were not included in the model.
5. Results:

5.1 Core Description

The uppermost 150 cm of the core was highly disturbed, water saturated diatom ooze, which was not sub-sampled. The upper 1000 cm of the core is generally very thinly bedded to laminated diatom ooze with homogeneous zones occurring at 430-460 cm, 490-530 cm, and 650-680 cm (Fig. 6). From 1000 cm to 1800 cm, a similar pattern is exhibited as seen in the upper core except with thin to thickly laminated diatom ooze. There are also numerous homogeneous zones seen in the mid section of the core at 1040-1110 cm, 1140-1195 cm, 1510-1530 cm, 1560-1610 cm, 1640-1660 cm. From approximately 1800-2100 cm there is a large homogeneous zone consisting of sandy diatom ooze followed by a varved zone until 2170 cm depth (Leventer et al., 2006). A silty-clay zone from 2170-2360 cm is located above the medium-thinly bedded diamict layer (2360-2390 cm), rich with sand and gravel (Fig. 7).

5.2 X-ray Analysis

The frequencies (thickness) of laminations varies down core, but ultimately depicts an undulatory pattern of high frequencies (thin laminations) followed by periods of lower frequencies (thick laminations) (Fig. 8). The significant high frequency zones are seen between 482.5-1087.5 cm, 1417.5-1862.5 cm, 2102.5-2202.5 cm, and 2312.5-2345 cm. The deepest two zones (2102.5-2202.5 cm and 2312.5-2345 cm) are the most densely laminated portions of the core with frequencies reaching an upwards of 16 laminations per 5 cm of core. The frequency of gravel grains down core is sparse. Gravel grains are generally absent through the first 1000 cm of the core, only seen at 542.5 cm
and 1007.5 cm depth (Fig. 8). Gravel grain abundance could be skewed by the use of a u-channel for X-ray analysis. In the lower 1000 cm of the core, single gravel grains were slightly more prevalent, appearing 14 times. It is not until the base of the core (2100-2300 cm depth) that there is a prevalence of gravel grain clusters. The base of the core (2170-2390 cm) exhibited coarse, sandy particles, which is not apparent at any other point in the core profile.
General Core Lithology of JPC-40

Figure 6. Lithostratigraphy for JPC-40.

Figure 7. Core photographs showing examples of massive texture (left), laminated texture (middle), and diamict (right).
Lamination and Gravel Grain Frequency

Figure 8. Frequency plot of laminations (grey) and gravel grains (blue) down core by depth. The black line is for the purpose of enhancing the visual trend from the histogram plot.
5.3 Concentration-Dependent Magnetic Parameters

Magnetic susceptibility is generally very weak for the majority of the upper core (Fig. 9 & 10). The intensity ranges from \(-8.33 \times 10^{-9}\) m\(^3\)/kg to \(5.00 \times 10^{-8}\) m\(^3\)/kg down core with regularly occurring high and low peaks seen approximately every 50-80 cm. At approximately 1650 cm depth, there is a significant increase down core in magnetic susceptibility until the base of the core. This significant increase at the base of the core appears to correlate with the significant changes in the lithostratigraphy. ARM values down core generally range from \(2.50 \times 10^{-7}\) Am\(^2\)/kg to \(1.50 \times 10^{-5}\) Am\(^2\)/kg with some isolated spikes of higher peak intensities down core. SIRM values range from \(1.00 \times 10^{-4}\) Am\(^2\)/kg to \(5.00 \times 10^{-4}\) Am\(^2\)/kg for most of the core until approximately 2000 cm depth where it increases up to \(1.30 \times 10^{-3}\) Am\(^2\)/kg. ARM and SIRM exhibit a similar stable trend in the upper part of the core with similar features occurring at approximately 500, 750, 1500, 1750, 2025, and 2175 cm down core. Between approximately 1500-1800 cm depth, ARM and SIRM both exhibit a similar “raised” interval that has a more intense signature than the background. Magnetic susceptibility and SIRM show a similar high section at the base of the core, whereas ARM does not.

5.4 Magnetic Grain-Size Dependent Parameters

The ARM/SIRM ratio ranges down core from approximately 0 to 0.05, except at approximately 1500 cm depth where there is a significant peak reaching 0.12 (Fig. 11 & 12). There are also less pronounced peaks seen down core at 390, 727.5, 817.5, 927.5, 995, 1180, 1315, 1585, 1785, and 2175 cm depth. The ARM/SIRM ratio appears to have no correlation to changes in sediment texture except at 1522.5 and 2175 cm depth. At
1522.5 cm depth (~7.9 ka), a large peak coincides with a laminated zone reaching values of 0.077. A single spike at 2175 cm depth (14.4 ka) occurs in the silty-clay zone with a maximum value of 4.21. $\chi_{\text{ARM}}/\chi$ ratio analysis values range down core from approximately 0 to 25.0, with isolated, single-point negative and highly positive values, likely controlled by extremely weak and diamagnetic $\chi$ values at those depths. From approximately 1850 cm depth to the base of the core, the values become more stable and steady with minimal oscillations or significant peaks. $\chi_{\text{ARM}}/\chi$ ratio also exhibited significant peaks that correlated with those of the ARM/SIRM ratio analysis described previously.

5.5 Mineralogy-Dependent Parameters

IRM acquisition curves show saturation for the upper 1500 cm of the core occurring at approximately 0.2 T, consistent with low-coercivity minerals such as magnetite, titanomagnetite, pyrrhotite and greigite (Evans and Heller, 2003) (Fig. 13). In the upper 1500 cm SIRM/$\chi$ values range from 1000-12000 A/m, which is typical for magnetite and titanomagnetite (Evans and Heller, 2003) (Fig 14 &15). Larger peaks outside this range are observed from 1500-1835 cm and could be attributed to the presence of magnetic iron sulfides, such as greigite or pyrrhotite. There are multiple spikes down core, most notably between depths of 725-1000 cm and 1500-1850 cm. Below 1850 cm, the values become very stable with the absence of any significant peaks to the base of the core. There appears to be no correlation between SIRM/$\chi$ and lithology. S-ratio analysis exhibits values between 0.9-0.95 down core until approximately 1700 cm, which is consistent with magnetite and titanomagnetite. The values are, on average,
decreasing slightly down core until approximately 1800 cm depth. A large peak with a value of 0.90 occurs at 2200 cm depth. Pilot samples from the base of the core exhibited saturation as high as 0.6-0.8 T, where S-ratios drop to 0.7 - 0.85, suggesting the presence of high-coercivity minerals such as hematite and goethite.
Figure 9. Raw data of concentration-dependent parameters plotted by depth.
Figure 10. Smoothed data (3-point) of concentration-dependent magnetic parameters plotted by depth.
Figure 11. Raw data of grain size-dependent magnetic parameters plotted by depth.
Figure 12. Smoothed data (3-point) of grain size-dependent magnetic parameters plotted by depth.
Figure 13. Stepwise IRM acquisition curves. Shaded area represents range in which samples saturated.
Figure 14. Raw data of mineralogy-dependent magnetic parameters plotted by depth.
Figure 15. Smoothed (3-point) data of mineralogy-dependent magnetic parameters plotted by depth.
6. Discussion

The base of the core consists of diamict from approximately 2360-2390 cm depth. The diamict transitions to a silty-clay zone (2170-2360 cm) and then into a diatomaceous mud and ooze above 2170 cm (Fig. 6). We interpret this sequence to represent proximity to the ice sheet grounding line (diamict), followed by either a transient ice shelf and/or heavy sea ice cover (silty-clay), and finally a transition to an open ocean environment (diatomaceous mud and ooze). The diamict at the base of the core is not stiff or compacted, so it is believed that the core site was near to but seaward of the grounding line. The transition to an open marine environment (~2170 cm) is dated at approximately 14.3 ka (all dates presented in calibrated years before present). The timing of deglaciation elsewhere along the East Antarctic Margin (EAM) is reported as 10.5-11.5 cal ka (Leventer et al. 2006). Leventer et al., 2006, argued that this late timing of deglaciation excluded the East Antarctic Ice Sheet as the source of a meltwater pulse (MWP1-A) that is observed in the sea level record at 14.2 ka (Fairbanks et al., 1989). Newly available \(^{14}\text{C}\) dates for JPC40 (Costa et al., 2008; this study) suggest that deglaciation of the Mac. Robertson shelf may have contributed to MWP1-A (Figure 17).

Within the Holocene section, the bulk of the core is laminated diatomaceous ooze, with intermittent homogeneous zones. The Holocene section generally has low concentrations of fine-grained magnetic material, which is most likely magnetite/titanomagnetite based on IRM acquisition and S-ratio analysis (Fig. 18, 19). The diatom ooze depicts the onset of a shift (~9.5 ka at ~1700 cm) in mineralogy to a
hematite/goethite assemblage and possibility magnetic iron sulfide rich zones as well (greigite/pyrrohtite). This is based on S-ratio and SIRM/χ analysis down core.

The major shift in S-ratio depicts a change from magnetite/titanomagnetite to hematite/goethite, which could be a function of the source area of the magnetic material (Fig. 20, 21). The appearance of hematite/goethite in the lower core correlates with the presence of sand grains in the lithology. Hematite/goethite assemblages could be magnetic material derived locally from Mac.Robertson Land, while magnetite/titanomagnetite seen in diatomaceous ooze zone of the core could be attributed to sediment reworking on the shelf or the coastal currents bringing in material from further afield. The shift could also represent diagenesis taking place within the basin, with magnetite preferentially dissolved at the base of the core (Fig. 21).

The frequency (thickness) of laminations varies down core, but ultimately show an undulatory pattern of high frequencies (thin laminations) followed by periods of lower frequencies (thick laminations) (Fig. 22). Homogeneous zones were seen at approximately 6.6-6.7, 7.0-7.2, 9.3-9.4, 9.7-10.0, 10.2-10.3, and 11.3-13.1 ka. These zones are likely the result of bioturbation from burrowing organisms that mix surficial sediments. These intervals may represent times of higher than normal oxygen levels in the bottom water, possibly resulting from increased mixing of the water column. A thinly laminated zone is present just above the diamict, with frequencies reaching 16 laminations per 5 cm interval. Leventer et al., 2006, observe this pattern in several EAM cores, and interpret these features as marine varves. Gravel grains are rare to absent throughout most of the diatomaceous mud. A lack of gravel grains throughout the upper core suggests that ice-rafted debris was minimal during the Holocene.
There is no correlation between sediment texture (i.e., laminated vs. homogeneous) and magnetic parameters in the Holocene section. Weak values of magnetic susceptibility in the upper core are attributed to the high biogenic silica content (35-50% biogenic silica, Costa et al., 2008), which also promotes the idea that this area was an open ocean environment with high productivity of diatoms for a period of approximately 10,500 years.

ARM exhibits a zone from 0.8-4.4 ka with regularly-spaced peaks occurring every 300-700 years (Fig. 16). This periodicity was calculated using the age model to approximate the spacing between major peaks in this zone. Wavelet analysis was also conducted to identify non-stationary periodicities. The wavelet analysis was conducted on ARM, SIRM, $\chi$, ARM/SIRM, SIRM/$\chi$, and $\chi_{ARM}/\chi$ (see Appendix). The latter four parameters yield little evidence of periodicity. ARM and SIRM showed the most evidence, although spectral power was weak (Fig. 23, 24). This is probably due to the age model being based on a small number of dates (9), which may be insufficient to constrain a highly varying sedimentation rate in the study area. ARM showed reoccurring century-scale (250-500 yr) periodicities. These occur at approximately 2.0-3.0 ka, 8.0-11.0 ka, and 12.0-13.0 ka. There is also spectral power in the 1000-2000 yr band that is present from approximately 5.0-11.0 ka. Similar periodicities are seen in SIRM, although a weaker power is exhibited. The periodicities observed in JPC40 are similar to century- and millennial-scale productivity cycles seen in late Holocene records along the Antarctic Peninsula (Fig. 25) (Leventer et al., 1996; Domack et al., 2001; Brachfeld et al., 2002). Along the Antarctic Peninsula, century-scale fluctuations were interpreted as a productivity cycle. During warm and less windy periods, sea ice melting created a low
density fresh water layer at the ocean’s surface, allowing phytoplankton to remain in the
photic zone (Leventer et al., 1996). A correlation between % biogenic silica (BSi) and
magnetic parameters such as ARM and SIRM could show that productivity may play a
role in the Nielsen Basin. However, this is most likely not the case in this study area due
to a lack of correlation between % BSi and both ARM and SIRM parameters
(unpublished BSi data from Costa et al., 2008, are not shown here). This was concluded
by calculating the Pearson’s correlation coefficient (R-value) between BSi and each
magnetic parameter. For both parameters the R-value was near zero (ARM: -0.00753,
SIRM: -0.00915) meaning lack of significant correlation with % BSi. In JPC40, these
periodic inputs of finer-grained material could be attributed to periodic meltwater plumes
from the EAIS during warm periods. Alternately, changes in current activity could lead to
resuspension of adjacent shelf sediments that are deposited into the deep basis.

A magnetic-rich interval above the background ARM and SIRM levels is seen
between 7.7-10.4 ka (Fig. 16). This magnetic-rich zone is seen in both ARM and SIRM,
but is not present in the ARM/SIRM profile until 7.8 ka (Fig. 17). The ARM/SIRM ratio
ranges from approximately 0 to 0.05, except at approximately 7.4-8.0 ka where there is a
significant peak in value reaching 0.12. This peak in the data shows that there is a
significant increase in finer particles at this horizon. This magnetic-rich period partially
coincides with an early Holocene warm period (Holocene Thermal Maximum = HTM)
seen on the Antarctic Peninsula from 6.7-9.0 ka (Fig. 25) (Leventer et al. 2002). Along
the Antarctic margin, an increase in temperature could result in increased glacial
meltwater carrying terrigenous material into the ocean. This early and mid Holocene
warm period has also been evident in Western Antarctica and in other studies from
different regions of the Earth such as Canadian Arctic, Chukchi-Alaskan margin, and other western Antarctic regions (Brachfeld et al., 2009; Roberts et al., 2004; Vare et al., 2009). Roberts et al. studied the paleosalinity history of Beall Lake, Windmill Islands (Eastern Antarctica). They noted changes in diatom assemblages between 8.0-4.8 ka, which was ultimately attributed to a warm period that caused glacial meltwater to enter the lake, reducing salinity in the lake, allowing freshwater diatoms to thrive.

Early Holocene warmth is also recorded in the Arctic. Brachfeld et al. 2009, used environmental magnetism on a sedimentary record from the eastern Chukchi Sea (Chukchi-Alaskan margin) and noticed the occurrence of the HTM from 9.5-8.7 ka. This was manifested as pulses of fine-grained magnetite entering the Chukchi Sea. This was thought to be derived from glacial recession and thawing of the permafrost, thus increasing the amount of discharge into rivers and other waterways, eventually entering the Arctic Ocean. Vare et al. 2009, studied sea ice records from the Barrow Strait in the Canadian Arctic Archipelago (CAA). They used IP$_{25}$ as a biomarker which is a mono-unsaturated highly branched isoprenoid (HBI) that is biosynthesised specifically by sea ice diatoms. Vare et al. 2009 has shown this compound to be stable in sediments below Arctic sea ice. They noted the HTM to occur from 8.9-4.9 ka. This was manifested as temporal changes in spring sea ice. They concluded that the onset and termination of the HTM was likely later in the CAA due to the slow retreat of the Laurentide Ice Sheet. These studies strengthen the theory that this warm period is not a local event but likely a large-scale global warm period.
Figure 16. Smoothed (3-point) ARM and SIRM data. Boxes highlight zones 0.8-4.4 ka and 7.7-10.4 ka discussed in the text.

Figure 17. This plot shows the rate of sea level change (upper horizontal-axis) vs. time (vertical-axis). The two major peaks are Meltwater pulses 1A and 1B. The timing of Meltwater pulse-1A is highlighted (red circle). Summer insolation at 60° N is plotted on the lower horizontal-axis and is shown as the orange/red line (Ruddiman, 1999, Figure 14-4, with original data from Fairbanks, 1989).
Figure 18. Smoothed (3-point) data of grain size-dependent magnetic parameters plotted by age.
Figure 19. Smoothed (3-point) data of mineralogy-dependent magnetic parameters plotted by age.
Figure 20. Plot of ARM/SIRM and S-ratio by age noting significant features in the data.
Figure 21. Plot showing the shift in mineralogy from magnetite/titanomagnetite in the upper core to hematite/goethite in the lower core.
Figure 22. Laminations and Gravel Grain frequency plotted vs. age. The black line is for the purpose of enhancing the visual trend from the histogram plot.
Figure 23. Wavelet analysis plot for ARM 3-point smoothed data.
Figure 24. Wavelet analysis plot for SIRM 3-point smoothed data.
Figure 25. Comparison of magnetic susceptibility data from the Palmer Deep, Antarctic Peninsula (Domack et al. 2001) and ARM data from Nielsen Basin (this study). The blue boxes highlight the areas of interest. The orange arrow denotes the timing of the HTM according to Leventer et al. 2002.
7. Conclusions

A late Holocene-Pleistocene core from the Mac. Robertson shelf, East Antarctica, has been analyzed using sedimentologic and magnetic methods. The majority of the Holocene sediment is laminated diatomaceous mud and ooze. The formulation of a new age model based on $^{14}$C dates (Costa et al, 2008) places the transition to an open marine environment at approximately 14.3 ka. Therefore, deglaciation of the Mac.Robertson Shelf could have potentially contributed to MWP-1A. Homogenous zones are seen down core with the Holocene section, possibly a result of stronger mixing of the water column and higher dissolved oxygen in the bottom waters. This would have allowed burrowing organisms to bioturbate the sediment.

Magnetic parameters do not correlate with sediment fabrics, although the S-ratio plot may exhibit correlation but is inhibited by its lower resolution sample interval. However, several distinct intervals are seen in the down core magnetic profiles. There is a shift in the magnetic mineral assemblage at 9 ka. S-ratios indicate that hematite/goethite is present from 9.0-14.6 ka and absent from 0.8-9.0 ka. This shift from hematite/goethite could be due to locally derived magnetic material from Mac.Robertson Land, while magnetite/titanomagnetite could have been introduced to the basin from an outside source, or through sediment resuspension on the shelf and/or coastal current activity.

ARM shows 300-700 yr oscillations between 0.8-4.4 ka. This pattern is similar to century-scale productivity cycles seen in the late Holocene along the Antarctic Peninsula (Leventer et al., 1996; Domack et al., 2001; Brachfeld et al., 2002). A magnetic-rich interval manifested as enhanced ARM and SIRM between 7.7-10.4 ka partially corresponds with an early Holocene warm period observed from 6.7-9.0 ka on the
Antarctic Peninsula and other locations such as the Canadian Arctic, Chukchi-Alaskan margin, and other western Antarctic regions (Brachfeld et al., 2009; 2009; Leventer et al., 2002; Roberts et al., 2004; Vare et al., 2009). This promotes the idea that the HTM is likely a global occurrence, rather than a local event. These conclusions can be considered evidence that the ice-sheet-ocean interactions along the East Antarctic margin have contributed to and responded to major continental and global events. This evidence suggests that the EAIS has the potential to contribute to similar events that may occur in the future as a result of local and large scale climate change.
References


ION Interactive Wavelet, http://ion.researchsystems.com/IONScript/wavelet/


Appendix

Wavelet Analysis Plots
Wavelet analysis plot for ARM raw data.

a. ARM Am2/kg Raw

b. Wavelet Power Spectrum

c. Global Wavelet

Morlet 6.00
Real (solid) Imaginary (dash)

http://paos.colorado.edu/research/wavelets/
A.2 Wavelet analysis plot for SIRM raw data.

- **a. SIRM Raw**

- **b. Wavelet Power Spectrum**

- **c. Global Wavelet**

http://paos.colorado.edu/research/wavelets/
A-3. Wavelet analysis plot for Magnetic Susceptibility raw data.
A-4. Wavelet analysis plot for Magnetic Susceptibility 3-point smoothed data.
A-5. Wavelet analysis plot for ARM/SIRM raw data.
A-6. Wavelet analysis plot for ARM/SIRM 3-point smoothed data.