Orbital Forcing of West and East Antarctic Ice Sheets  
3.8-3.0 Ma: A Climate Analogue in Weddell Sea  
ODP Site 697  

by  
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To

Suzanne

my mom my dad

Zack E&ES CA

the 409 crew, Boat House, the Art Lib crew

Cindy, Marty, Eduardo, Elena, Graham

Jamie, Tavo, Jason, Team Antarctica

Keck, COE, IODP, Claire

Espwesso

Wes

My warmest thanks for your support, care and guidance

Suzanne,
your mentorship has engendered in me a life-long passion for studying
the complex interconnectedness of our planet
you showed me that I too could be a
scientist
if I really put my mind to it
and make a tangible contribution
to addressing global problems
I’m passionate about
I’m so grateful you took me under your wing
cheers to more pie
chocolate covered espresso beans
and Torpedos to come
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Abstract

As climate change intensifies and the world begins to feel its impacts, scientists are looking to past climatically similar periods to understand how our planet will respond. During the Pliocene Epoch (5.33-2.58 millions of years ago), the West Antarctic Ice Sheet (WAIS) nearly de-glaciated under peak CO$_2$ concentrations of ~400 ppm (Fedorov et al., 2013) and global average temperatures 2-3°C warmer than pre-industrial values (Yamane et al., 2015). The sensitivity of the WAIS to CO$_2$ and ocean temperature fluctuations on orbital timescales is well established (DeConto & Pollard, 2003; Naish et al., 2009). A more controversial question is how sensitive the East Antarctic Ice Sheet (EAIS) is to similar forcings. To address this question, we analyze sedimentological changes in Antarctic Weddell Sea ODP Site 697 marine cores deposited 3.8-3.0 Ma for cyclicities in ice sheet dynamics and bottom water fluctuations. Spectral analysis of fine fraction mineral assemblages identifies smectite, illite, chert, chalcedony, SiO$_2$, garnet and feldspar. Continuous detrital smectite deposition points to high Antarctic Bottom Water (AABW) circulation through the Jane Basin 3.8 to 3.0 Ma. Varimax-rotated principal component analysis and wavelet analysis of elemental count ratios suggest Jane Basin sediment deposition transitioned from a dominant eccentricity forcing with obliquity inputs to a dominant precession forcing modulated by eccentricity at ~3.3 Ma. Derived linear sedimentation rates slow from ~6.3 cm/kyr between 3.78 and 3.21 Ma to 4.74 cm/kyr between 3.14 and 3.03 Ma. These findings indicate that EAIS and WAIS behavior as moderated by an eccentricity and obliquity forcing during Pliocene peak warming and a precession forcing during cooler conditions after ~3.3 Ma.
Chapter 1. INTRODUCTION

Climate change presents one of the most significant social and environmental challenges of our time. Sea level rise from anthropogenic climate change has already begun affecting coastal communities and island nations around the world (IPCC, 2014). These changes stress existing inequalities and disproportionately impact those with the least resources, presenting a substantial human rights dilemma (Barnett & Adger, 2007). Concern for these social impacts as well as environmental changes has prompted further research on the present and past functioning of the Earth climate system.

Satellite observations reveal that global sea level rise can primarily be attributed to melting of the Greenland and West Antarctic Ice Sheets (WAIS) under warming temperatures (Pritchard, 2012; Rignot et al., 2013). Studies of past cryospheric responses to modern climate conditions support 22-25 meter sea level rise from near total deglaciation of the Greenland and West Antarctic Ice Sheet (WAIS) and melting of low-lying margins of the East Antarctic Ice Sheet (EAIS) (Miller et al., 2012). Further study of the sensitivity of these ice sheets, especially the more massive WAIS and EAIS, to incremental increases in CO₂ and warm temperatures can inform more realistic predictions of future sea level rise and aid adaptation and mitigation strategies for at-risk populations.

1.1 Pliocene Epoch Climatic Conditions

The Pliocene Epoch (5.33-2.58 Ma) serves as a climate analogue to present day conditions. The Pliocene is a geologic epoch that occurred at the end of the Neogene Period in the middle of the Cenozoic Era (International Chronostratigraphic
Chart, 2015). During this epoch, the intensity of sunlight incident on Earth, global geography and atmospheric carbon dioxide levels were similar to present day conditions. Atmospheric carbon dioxide concentrations were predicted to be ~400 ppm (Raymo et al., 1996; Haywood et al., 2009; Fedorov et al., 2013), a level which was recently surpassed in 2013 (Blunden, 2014). Global mean annual temperatures during the Pliocene were 2-3°C warmer than today’s (Dowsett, 2007; Miller et al., 2012). Southern Hemisphere high latitude temperatures were even higher at 3–3.5°C above present values (Raymo et al., 1996; Fedorov et al., 2013). Peak Pliocene sea level rise estimates ranged from 5m to 40m higher than present, with ~25m most commonly cited (Dowsett et al., 1999; Raymo et al., 2009; McKay et al., 2012).

The Pliocene can be characterized by two contrasting climatic regimes; the first of increased warmth in the early Pliocene and the second of cooling in the late Pliocene (Haywood et al., 2004). Early Pliocene (~4.2-4.0 Ma) global mean annual temperatures are predicted to have been ~4°C above preindustrial values (Pagani et al., 2009). These two climatic periods are separated by a short burst of high warmth, known as the Pliocene Climatic Optimum (PCO). The exact date of this peak warm period is unclear (Table 1.1). Using the LRO4 benthic δ¹⁸O stack, the PCO is thought to be 3.3-2.9 Ma (Raymo et al., 2009; PLIOMAX; Dutton et al., 2015). Studies utilizing spectral analysis and sea surface temperature proxies find the PCO lasted from 4.3-3.5 Ma (Patterson et al, 2014), 4.4-4 Ma (Fedorov et al., 2013) or 5-4 Ma (Pagani et al., 2009).
<table>
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Table 1.1 Dates of the Pliocene Climatic Optimum as determined by six studies.
Figure 1.1 Global oxygen and carbon isotope records of from 65 Ma to present. Note history of Antarctic ice sheet formation (Zachos et al., 2001).

1.2 Orbital Forcings during the Pliocene Epoch

Climatic changes during the Pliocene Epoch were driven by periodicities in the Earth’s orbit, which determine the level and distribution of insolation received by the Earth (Ruddiman, 2008). Known as Milankovitch cycles, these periodic variations in the distance between the Earth and the Sun control average temperature; and in turn, atmospheric carbon dioxide concentrations, ice sheet behavior, deep ocean circulation, sea level, and sea surface temperature (Fedorov et al., 2006).

The Milankovitch cycles include eccentricity, obliquity and precession, each displaying a distinct periodicity. Eccentricity, the shape of the Earth’s orbit around the Sun (0-5% ellipticity), runs on a cycle of 100kyrs with manifestations at 400kyrs as well. It determines the total distance between the Earth and the Sun, thus driving small variations in the total energy received by the Earth. The 100kyr eccentricity signal is not obvious at high latitudes because these fluxes in insolation are evenly distributed around the globe. The eccentricity signal causes modulation in the precession signal at periods of 100kyr and ~400kyr (Ruddiman, 2008).

Precession, the wobbling of the Earth as it spins on axis (from pointing to the star Polaris to the star Vega) has a periodicity of 23kyr and a weaker cycle every 19kyr. Periodicities in the precession cycle result in greater seasonal contrast, which should impart a major influence on high-latitude summer insolation intensity (Patterson et al., 2014). However, at high-latitudes, the precession cycle is largely absent from geologic records of ice volume and sea level (Hays & Shackleton, 1976; Patterson et al., 2014). This has historically been attributed to the larger importance of
obliquity-paced annual insolation in influencing polar temperatures compared to precession-modulated seasonal insolation (Naish et al., 2009). While the influence of precession on orbital pacing of Late Pleistocene cycles is well established (Lisiecki & Raymo, 2005; Raymo & Huybers, 2008), the absence of a precession cycle in Antarctic climate indicators before this time hampers analysis of Pliocene precession forcings (Patterson et al., 2014).

Obliquity, oscillations in the degree of Earth’s axial tilt (21.5 -24.5°), occurs on a periodicity of 41 kyr and changes the severity of Earth’s seasonal variation. Obliquity is commonly cited as the dominant forcing of WAIS glacial behavior during the Pliocene Epoch (Naish et al., 2009). Changes in obliquity have the largest effect on seasonal insolation at high latitudes, amplifying seasonal differences during times of increased tilt or suppressing them during times of decreased tilt (Ruddiman, 2008). During periods of smaller axial tilt, the insolation distribution anomaly decreases between winter and summer, but increases between equatorial and polar latitudes. As a result, warmer winters increase meridional moisture transport and cooler summers minimize seasonal melt back, thus increasing ice sheet surface mass balance (Raymo & Nisancioglu, 2003; Hansen et al., 2015).

During the Pliocene, changes in Earth’s obliquity, in addition to changes in the precession of the equinox, made large contributions to high-latitude Southern Hemisphere climate records (Zachos et al., 2001; Grützner et al., 2005; Fedorov et al., 2006; Naish et al., 2009). At ~4.5 Ma the LRO4 benthic d18O stack shows a growing global coherency with precession, which becomes more significant between ~4.1-2.8 Ma (Lisiecki & Raymo, 2005). Hansen et al. (2015) also shows a shift to a dominant
precession response of Wilkes Land EAIS IRD at ~4.6 Ma. Other studies of Wilkes Land EAIS marine sediment cores (Patterson et al., 2014) show a dominant periodicity of 40kyrs between 4.3 and 3.5 Ma and transition to precession thereafter. Ross Sea ANDRILL sediment cores show obliquity-paced WAIS retreated prior to 3.0 Ma (Naish et al., 2009). See Figure 1.3 for location. Early to mid-Pliocene obliquity-paced mean annual insolation was produced by an increase in the duration and intensity of austral summer surface warming (Patterson et al., 2014). Under these conditions, the precession signal is canceled out over the course of a seasonal cycle (Huybers & Tziperman, 2008).

Patterson et al. (2014) argue the emergence of a stronger precession 20kyr signal at ~3.3 Ma emerged from southern high-latitude cooling. This cooling likely raised the EAIS melt threshold and restricted the duration of the melt season to times of austral summer insolation maxima, which is controlled by precession (Huybers & Denton, 2008; McKay et al., 2012). After ~3.3 Ma, precession response to perennial sea ice was modulated by eccentricity, where influence of oceanic warming was enhanced during 100kyr sea ice minima (Patterson et al., 2014).

Near the end of the Pliocene Epoch at ~3 Ma, various feedbacks came into play that amplified Milankovitch forcings. These changes are signaled by the appearance of northern continental glaciers and cold surface waters in oceanic upwelling zones in low-latitudes (Fedorov et al., 2006). It is possible that at ~3 Ma, the gradual shoaling of the oceanic thermocline reached a threshold after which winds began bringing cold waters to surface in low-latitude areas. This prompted changes in ocean-atmosphere interactions, which together with ice-albedo feedbacks, amplified
Obliquity cycles in Antarctic glaciomarine records (Fedorov et al., 2006). Obliquity cycles changed character again at 1 Ma from manifestations at 40kyrs to multiples of 40kyrs, specifically 80kyrs and 120kyrs (Fedorov et al., 2006). Since then, obliquity has continued to play a dominant role in pacing glacial terminations (Beaufort et al., 2001; Liu et al., 2004).

**Figure 1.2** Milankovitch Cycles Illustrated (CO2CRC).

**Figure 1.3** Vertical cross section through present day Antarctica (LIMA, NASA).
1.3 Antarctic Geology

Lying on the South Pole, the Antarctic land mass exceeds 14 million km$^2$, making it almost twice the size of Australia. Formation of the Antarctic continent began during the Napier Orogeny in the early Archaen with accretion around a cratonic nucleus (Anderson & Andrews, 1999). The geologic makeup of Antarctica can be subdivided into five source regions. Of the five, East Antarctica possesses a larger and more diverse composition. The continent is separated into two distinct
geomorphologic regions: the West and East, each hosting its own ice sheets. East Antarctica bedrock formed during the Precambrian eon and the Paleozoic era while West Antarctica bedrock formed during the Mesozoic and Cenozoic eras. The Transantarctic Mountains (TAM) extend from the Ross to Weddell Sea and act as the defining barrier between the WAIS and EAIS (Harley, 2005).

East Antarctica first began developing large-scale ice sheets between 14.8 and 14.5 Ma after the mid-Miocene climatic optimum (Zachos et al., 2001). West Antarctica did not experience widespread glaciation until the late Miocene and early Cenozoic eras (Levitan et al., 2014). See Figure 1.1 for relation to oxygen and carbon isotopic records and context in paleoclimatic history. Today, the majority of Antarctica is covered in mile thick ice sheets (Fretwell et al., 2013). The WAIS and EAIS hydrologically behave independently of each other as a result of their unique topographies and dimensions. The marine-based WAIS (3.3 million km$^3$) is more sensitive to sea level and sea temperature change, as heating through conduction contributes an additional mechanisms for ice melt. The EAIS (26.5 million km$^3$) is land-based and more massive, making it less vulnerable to warming mechanisms (Ingölfson, 2004).

The extent to which the EAIS contributed to Pliocene sea level rise is highly disputed. Recent studies show it is conceivable that the EAIS coastline experienced significant retreat in the late Pliocene (Raymo et al., 2006). In these studies, EAIS ice margin retreat is predicted to be several hundred kilometers, which could have contributed to 3-10m of global sea level rise. Further analysis of this question could
aid our understanding of future behavior of the EAIS in low-lying areas to current warming temperatures (Cook et al., 2013).

![Figure 1.5 Antarctica Bedrock A. Distribution of Antarctic Rock Exposures. B. Gondwana Affinities of Main Tectonic Regions (Boger, 2010).]

1.4 Weddell Sea Fine Fraction Minerals

Analysis of clay (Ehrmann et al., 1992) and silt (Diekmann & Kuhn, 1998) mineral associations is a well-established tool in studies of Southern Ocean circulation and the evolution of Antarctic climate (Robert & Maillot, 1983; Junttila et al., 2005). Changes in climate and continental morphology affect the genesis of clay and silt minerals and as a result, the composition of oceanic detrital fine fraction associations (Millot, 1970; Chamley, 1997). The small size of clay particles (<2µm) makes them prone to erosion, transport and redistribution by various media, including fluvial transport, wind transport and bottom currents (Ehrmann et al., 1992). Clay and silt mineral analysis is useful for reconstructing sedimentary processes and glacial-interglacial cycles (Chamley, 1979).
Through the Neogene Period (23.03 Ma- 2.58 Ma), Antarctic clay mineral records exhibit only slight long-term changes attributed to local changes in glacial erosion and the supply of source material (Hillenbrand & Ehrmann, 2001). Weddell Sea clay particles are primarily of detrital origin; eroded from parent-rocks, ancient sediments and soils. Antarctic water circulation has had a strong influence on clay sedimentation near West Antarctica since the early Pliocene and near East Antarctica since the middle Miocene (Robert & Maillot, 1990). Central and northern Weddell Sea fine fraction minerals arrive via bottom currents and/or episodic turbidity currents and likely originate from southern Weddell Sea sources (Diekmann & Kuhn, 1999).

In past studies of Antarctic marine sediments (Robert & Maillot, 1990), the siliceous minerals smectite, kaolinite, chlorite and illite were identified in rare to abundant quantities. Smectite is a 2:1 silicate that includes a layer consisting of two inward-pointing tetrahedral sheets with a central alumina octahedral sheet. In modern soils of ice-free Antarctic regions near the Ross Sea, smectite formation is mainly driven by moisture and abundant in wet climates and preset in soils on volcanic rocks (Ehrmann et al., 1992). Early Cretaceous clay minerals from the Dronning Maud Land continental margin and Late Cretaceous clay fraction minerals on Maud Rise contain high levels of smectite (Barker & Kennett, 1988). Detrital smectites are closer to motmorillonite-beidellite in composition and contain Ca, Na, K (Iacoviello et al., 2012), Al or Fe phase presence in their matrix (Rivagli et al., 2014).

Kaolinite is a dioctahedral mineral composed of a single-layer structure of one octahedral sheet condensed with one tetrahedral sheet (Bohor & Hughes, 1971). Kaolinite forms from the weathering of feldspars under acid conditions or
hydrothermal alteration, thus formation is favorable in warm and humid conditions (Robert & Chamley, 1987). Kaolinite presence in Antarctica is assumed to be a relic of warmer times and is relatively resistant to erosion (Biscaye, 1965). Smectite and kaolinite result from chemical weathering and therefore their presence points to warmer climatic conditions of increased precipitation and decreased glaciation (Junthilia et al., 2005).

Illite and chlorite are both micaceous silicate minerals derived from erosion of parent-rock and poorly developed soils (Grim, 1962). Illite forms from the weathering of primarily feldspar silicates, the degradation of muscovite, or the alteration of other clay minerals (Deer et al., 1962). Formation of both illite and chlorite require weak chemical and strong physical weathering in which clay minerals cannot complete their evolution. Such conditions are usually associated with cold climates (Junthilia et al., 2005).

**Figure 1.6** Spectra of smectite, chlorite, kaolinite and illite created on ENVI Classic using the USGS mineral spectra library.
1.5 The Weddell Sea and Antarctic Bottom Water (AABW) Formation

The Weddell Sea receives sediments from the EAIS, WAIS and Antarctic Peninsula Ice Sheet, making it particularly interesting for studies of deep-sea sediment transport and depositional processes related to ice-sheet growth and shrinkage (Lindeque et al., 2013). Furthermore, the Weddell Sea is the principal locality of Antarctic Bottom Water (AABW) production and therefore a crucial piece of the global ocean circulation system (Carmack and Foster, 1975; Anderson & Andrews, 1999).

AABW is the densest water in the ocean. It sinks below southerly water masses in the Weddell Sea carrying CO$_2$ and O$_2$ to great depths, travels along the ocean floor and at times reaches areas north of the equator (Orsi et al., 1999). When sea ice forms in the Antarctic margin, the remaining seawater becomes increasingly saline and sinks, mixing downslope with Circumpolar Deep Water (CPDW) and Sub Antarctic Slope Water to eventually become AABW. AABW creates a seafloor boundary over which warmer waters are able to travel and upwell, providing prime conditions for bioproducitivity (Anderson, 1999). As AABW flows out of the Weddell Sea, it is strongly stirred by deep cross shelf valleys in the northwest Weddell Sea and shapes the seabed with outwash (Fischer et al. 1988; Jerosch et al., 2015). Erosional channel-levee systems filled with thick chaotic-facies deposits direct the densest variety of bottom water, Weddell Sea Bottom Water (WSBW), to the northeast (Jerosch et al., 2015; Haid & Timmerman, 2013).

AABW formation and outflow has varied with temperature fluctuations through geologic history (Martin et al., 2002). Peak Pliocene warming reduced
meridional and zonal temperature gradients, slowing AABW outflow through the meridional stream function by up to 5 m$^3$/sec (Haywood et al., 2004; Fedorov et al., 2013). Reduced AABW flow had implications for various global biological and climatic systems; including nutrient upwelling, oxygen transport, and carbon sequestration (Sigman et al., 2010).

Figure 1.7 Schematic representation of Southern Ocean circulation. Antarctic Circumpolar Current shown in blue with thin dark blue lines showing average positions of Antarctic Front. Grey shading indicates water depth less than 3,000 m. Note Weddell Sea Gyre circulation. Approximate location of Jane Basin ODP Site 697 marked (Ocean Circulation, Open University Course Team, 1989).
Figure 1.8 AABW Outflow from the Weddell Sea to the Polar Front. Flow direction indicated with arrows. Note ODP Site 697 location in Jane Basin (Shipboard Scientific Party, 1988).

1.6 The Jane Basin

The Jane Basin is located on the East Antarctic continental slope in the Antarctic Weddell Sea. AABW flows through the Jane Basin as part of the general clockwise circulation of the Weddell Gyre (Foster & Middleton, 1979). This area was selected for this study because it records a thick Pliocene sequence of sediment deposit from the EAIS and WAIS with high sedimentation rates. In addition, the Jane Basin sediment record contains information on AABW outflow and the confluence of Antarctica’s major surface currents, including the Antarctic Circumpolar Current, the East Wind Drift, and the Weddell Gyre (Pudsey, 1990).

The Jane Basin formed 30-25 Ma as a back-arc basin separating the Jane Bank island arc from the South Orkney microcontinent (Shipboard Scientific Party, 1988). The oldest sediments in the Jane Basin date back to 15.2 Ma (Lindeque et al., 2013). Sediment thickness is estimated 1580 meters and sedimentation rates are estimated to
be 10.5 cm ky in preglacial periods, 11.5 cm ky transitional periods, 9.9 cm ky glacial periods (Barker & Kennett, 1988; Lindeque et al., 2013).

Figure 1.9 Weddell Sea and Jane Basin Topography. Jane Basin location is highlighted in yellow (Lindeque et al., 2013).
Figure 1.10 Weddell Sea and Jane Basin Sediment Deposition with Age along the WS-SS Seismic Transect. Jane Basin location is highlighted in yellow (Lindeque et al., 2013).

1.7 The International Ocean Discovery Program (IODP)

The International Ocean Discovery Program (IODP), formerly known as the Ocean Drilling Program (ODP), is a scientific deep-sea drilling collaboration that collects ocean floor geological data and samples for research on Earth history and dynamics. Sediment cores for this study were recovered on March 5, 1987 on ODP Expedition 113 to the Weddell Sea, Antarctica. The goal of this Expedition was to collect information for subsequent studies on Antarctic ice sheet formation, Antarctic Bottom Water (AABW) formation and other Antarctic phenomena.

Cores from Hole 697B (61°48.626'S, 40°17.749’W) in the Jane Basin were selected for analysis for their well recorded magnetic reversals and high sedimentation rates, which provides good age resolution; especially for a site at a
pelagic depth (Pudsey, 1990). Cores 697B 13X through 17X, spanning 3.0 to 3.8 Ma, were focused on because they contain six well defined magnetic reversals and cover multiple proposed dates of mid-Pliocene peak warming (see Table 1.1).

Site 697 sits in the deepest of a three-site depth transect on the southern margin of the South Orkney microcontinent (SOM) (Shipboard Scientific Party, 1988). Under present day conditions, the site is ice-free for several months of the year and receives high levels of nutrient upwelling (Pudsey, 1990). The core was drilled for 69 hours using a piston core, at a water depth of 3480 m, yielding 32 cores with a recovery rate of 62%. Initial shipboard analysis identified primarily hemipelagic sediments, ice rafted detritus (IRD) throughout, numerous thin volcanic ash layers, and minor siliceous biogenic component (Shipboard Scientific Party, 1988). Smectite, kaolinite, illite and chlorite were identified in the clay fraction. ODP cores are labeled first with the expedition number, then site number, hole letter, core number, letter indicated the type of drilling (R for rotary drilled and X for piston cored) and the section number.

Figure 1.11 Stratigraphy of ODP Leg 113 Site 697 with time scale, depth of main horizons and extracts of WS-SS transect seismic data (Barker & Kennett, 1988; Lindeque et al., 2013).
Figure 1.12 Site 697 clay mineralogy determined using XRD (Shipboard Scientific Party, 1988).
Chapter 2. METHODS

To determine orbital forcings on depositional behavior in the Jane Basin during the mid-Pliocene (3.8-3.0 Ma), varimax-rotated principal components of elemental count ratios of Cores 697B 13X-17X were created and run through wavelet analysis. The loading size of the elemental count ratio principal components as well as identification of fine fraction minerals aids our interpretation of AABW behavior and climatic conditions during deposition. In conjunction with this study, Kaufman, et al. (2016) analyzes these cores for heavy mineral provenance and IRD, which points to iceberg abundance and ice sheet instabilities (Marcott et al., 2011). Cindy Flores and Graham Stewart performed biosilica extraction to collect more information on paleoproductivity and sea ice cover. These various methods collaborate to examine EAIS and WAIS glacial behaviors as well as AABW flow under mid-Pliocene climatic conditions.

2.1 X-Ray Fluorescence Core Scanning

X-Ray Fluorescence (XRF) spectroscopy is an analytical technique used as a means of determining major and trace element concentrations of a particular material. XRF analysis produces element counts per area through the bombardment of high-energy wavelength X-ray photons at the surficial atom of an object. If the energy of the X-ray is larger than the ionization potential of the electrons, one or more electrons in an inner K-shell are excited and ejected when struck with the X-ray. As a result, the atom becomes unstable. To re-stabilize, an electron from the atom’s outer shell fills the vacancy left in the innermost shell and in doing so, emits energy in the form of photons, which is released in the form of fluorescent light. The element-specific
wavelength is detected and quantified to determine the relative abundance of a specific element in a specific area. The elemental values that are obtained are recorded as counts per second of a given area (cps), which is a relative and not absolute measurement. The XRF method is particularly popular because it is a quick and non-destructive method of obtaining elemental compositions (Tjallingii, 2006).

Cores were scanned by a third-generation AVAA TECH® XRF Core Scanner at the IODP Texas headquarters at Texas A&M in College Station, Texas. Before scanning, the split cores were lightly scraped with common glass to remove any mold and to produce a smooth, fresh surface. The cores were then imaged with the Ocean Drilling Program Core Imaging Machine, providing a .tif file and high-resolution RGB data excel file for each core section. Dropstones, which are pebbles and small rocks, were imaged, measured and identified before being removed from the cores for XRF scanning. A 4 \( \mu \)m thick 3525 Ultralene film (SPEX Centriprep, Inc.) was placed over the cores to improve contact between the XRF reader and the core.

The cores were then placed in the XRF Core Scanner and surficial scans were taken down the center of the core at 2 cm intervals with an X-ray illumination area of 1 cm\(^2\). Scans were taken at 10kV and 30kV. The 10kV energy runs detect the major elements Al, Si, P, S, Cl, Ar, K, Ca, Ti, Cr, Mn, Fe, and Rh. The 30kV energy detects heavier trace elements including Ni, Cu, Zn, Ga, Br, Rb, Sr, Y, Zr, Nb, Mo, Pb, and Bi. On cores where the presence of barium was of interest, an additional scan at 50 kV (for Ag, Cd, Sn, Te, and Ba) was performed. Areas with an uneven surface were skipped because elemental count values can be skewed by changes in topography or a bad contact. Water content can also skew the results, however, these cores were
largely dry. Elemental counts were recorded in .spe files that were converted to .xls files using WinAxilBatch. Standards were run periodically to ensure there was no measurement drift.

**Figure 2.1** Schematic representation of XRF action on a Ti atom

a) An electron of the inner K-shell is targeted by an incoming X-ray, which ejects the electron out of its orbital. b) The ejected inner K-shell electron is replaced by an L-shell or M-shell electron, emitting a characteristic K\(_\alpha\) or K\(_\beta\) fluorescence, respectively c) The ejected L-shell or M-shell electron is replaced by an M-shell or N-shell electron, emitting a characteristic L\(_\alpha\) or L\(_\beta\) fluorescence, respectively (Tjallingi, 2006).

**Figure 2.2** Illustration of XRF core scanning apparatus (Tjallingii, 2006).

### 2.2 XRF Dataset Preparation and Magnetostratigraphy

Abnormal argon (Ar) elemental counts were used to identify artificially high or low readings due to bad contact with the XRF scanner. Ar is present in the air and
normally in very low quantities in sediment cores, therefore high Ar values indicate a point of bad contact with the instrument. Values outside of the argon elemental count second standard deviation were identify and removed from the dataset. Elemental counts were then normalized with titanium to account for terrigenous sources of sediment deposition (Jorgensen, 2012). Elemental counts will be referred to as elemental count ratios (i.e. Fe/Ti or Si/Ti) throughout this study to clarify that all values account for Ti content.

The depth measurements within each individual core were converted to a meters below sea floor measurements by adding the distance of the core top to the sea floor to the depth measurement within the core. Core catchers were removed from XRF analysis and accounted for during interpolation and meter below sea floor unit conversion. Meter below sea floor core measurements were transformed from a depth scale to a complete time scale by interpolating the ages of all depth measurements between the known ages of magnetic reversals. The magnetic time scale of Gee & Kent (2007) was used to date the magnetic reversal pattern from Hamilton & O’Brien (1990), which is in Pudsey (1990). Ages were added starting at the most recent magnetic reversal, which occurred in core 697B_2H_2 at 30 meters below sea floor (mbsf). The formula used for interpolation was: initial chronological boundary + (change in mbsf x (chron boundary f – chron boundary i)/ (mbsf f – mbsf i)). To account for water loss, each core section was stretched to its shipboard recorded interstitial length and the XRF elemental accounts were interpolated across the new range.
Table 2.1 Ages of each magnetic reversal in millions of years ago (Ma) with corresponding meters below sea floor (mbsf) measurement and core section for core 697B. Gee & Kent (2007) was used to date the magnetic reversal pattern from Hamilton & O’Brien (1990), which is in Pudsey (1990). Magnetic reversals cited in Lisiecki & Raymo (2005) also used.

<table>
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<th>Chron/Subchron</th>
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<th>1990 Age (Ma)</th>
<th>2015 Age (Ma)</th>
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</table>

2.3 Magnetic Susceptibility Analysis

Magnetic susceptibility provides a dimensionless value that reflects the total amount of magnetic material present in a given interval of sediment. Site 697B cores were run through a multi-track sensor at the IODP Texas headquarters to conduct magnetic susceptibility analyses. Measurements were taken at 0.1 cm intervals down the center of the core and recorded in an .ms (magnetic susceptibility) file. The .ms files were converted to .csv file in R Studio (see code in Appendix D) and interpolated. Meter below sea floor and age values were then calculated. High measurements that correspond to the depth of a recorded dropstone were removed.
2.4 Spectral Reflectance Data Collection and Processing

Spectral reflectance identifies the presence, but not abundance of common minerals in the sediment sample. The physical properties of the minerals present in the sediment determine the percent light reflected by the sample. Core samples collected at the IODP Texas headquarters were sorted into <63 µm, 63-150µm, 150-500 µm, and >500 µm size fractions at Wesleyan University. Thirty-six fine-fraction (<63 µm) sediment samples from Cores 697B 13X through 17X were identified for spectral analysis and mineral identification. Samples chosen contain high or low coarse fraction (>63 µm) accumulation and/or high or low biosilica accumulation.

Reflectance spectra were obtained using an ASD FieldSpec FR® spectroradiometer with a wavelength range of 350–2500nm. Readings were taken at an interval of 1.4nm between 350–1000nm and 2nm between 1000–2500nm at a spectral resolution of 3nm between 350–1000nm and 10nm between 1000–2500nm. Spectra were taken by hand-positioning a 25° field of view fiber optic sensor 10 cm above the fine fraction sediment sample under uniformed, artificial light. The <63 um fraction settled in a 1 liter plastic bucket during the grain size separation process. Water was decanted and when <3 cm of liquid remained, the samples were air dried. Silt grains settled first and as a result most samples had two distinct sides, a light silt fraction on the bottom topped with a dark, shiny clay fraction. Spectra were taken on both sides to account for heterogeneity in the sample. Ten measurements were taken per scan, converted to ASCII files using SpecPro and averaged in Excel to provide a single spectrum for each target. Spectra were normalized to a white Spectralon® panel.
2.5 Fine Fraction Mineral Identification

The averaged spectra were saved as ASCII files and inputted to XQuartz 2.7.7 ENVI 5.1 and ENVI Classic for mineral identification. The collected spectra were then individually compared to the USGS mineral spectra library. Minerals displaying similar spectral behavior were preliminarily identified. To verify the match, the major pronouncements (mineral-specific dips) of each collected spectra and each relevant library spectra were recorded and compared. Figures 2.3 and 2.4 illustrate this procedure using Sample 697B_14X_4 136-138 as an example. Figure 2.3 shows how the sample was matched to montmorillonite while Figure 2.4 shows how the sample did not match with kaolinite. Note the mismatch at the diagnostic kaolinite 2315 um absorption indicated with an arrow in Part A Figure 2.4.
Figure 2.3 Mineral Identification Procedure

A and B. The montmorillonite (smectite) spectra from the USGS library in comparison to Sample 697B_14X_4 136-138, clay side. Both spectra are plotted as a continuum removed data spectrum against wavelength in micrometers.

C. Comparison on wavelength (um) values for major absorptions.
Figure 2.4 Example of a Nonmatch in Mineral Identification Procedure. 
A and B. The kaolinite spectra from the USGS library in comparison to Sample 697B_14X_4 136-138, clay side. Both spectra are plotted as a continuum removed data spectrum against wavelength in micrometers. The absence of one of the kaolinite diagnostic dip at 2315 um (highlighted in yellow) prevents kaolinite being identified in this sample. C. This table shows the verification procedure for matching spectra behavior from a collected sample to the mineral-specific pronouncements at wavelength values.
2.6 Varimax-rotated Principal Component Analysis

Principal Component Analysis (PCA) is an exploratory statistical procedure that elucidates underlying relationships and structures in a complex dataset. Using an orthogonal transformation, it converts positively correlated variables into linearly uncorrelated variables called principal components (Jolliffe, 2002). The first principal component explains the largest amount of variance in the dataset, the second, orthogonal to the first, explains the second largest amount of variance, and so on (Cope, 2009; Hall, 2015). This procedure is commonly used in geophysics to identify traveling and standing waves in datasets (Horel, 1984).

Varimax-rotated principal component analysis (VPCA) is a common variation that allows variance in the dataset to be widely distributed across principal components (Cope, 2009). Doing so increases the number of loadings for variables on the components that follow the first component (Revelle, 2015). VPCA infers general trends and allows components to be more easily interpreted, still it is important to note that it is not a confirmatory procedure.

Analysis was conducted in RStudio, which is an open source software for R language statistical computing. The elemental count ratios used were Ag/Ti, Ba/Ti, Ca/Ti Cd/Ti, Cl/Ti, Cr/Ti, Cu/Ti, Fe/Ti, K/Ti, Mn/Ti, Mo/Ti, Ni/Ti, P/Ti, Rb/Ti, S/Ti, Sn/Ti, Sr/Ti, Zr/Ti, and Zn/Ti. Datasets were detrended to make cyclical patterns more easily identifiable (Hall, 2015). This was done by creating a second-degree polynomial trendline in each dataset, inserting the meter below sea floor depth at each interval into the equation for the trend line and then subtracting the trend line. Sample RStudio codes explaining this procedure are included in Appendix D.
2.7 Wavelet Analysis

Wavelet analysis was applied to XRF elemental count ratios to analyze the effect of different Milankovitch cycles on depositional behavior at ODP Site 697. Wavelet analysis is a statistical tool that illustrates localized variations of power within a time series dataset. Wavelet transform decomposes a time series into time-frequency space to identify dominant modes of variability and how they vary temporally (Torrence & Compo, 1998). As a result, information can be gathered on the amplitude of periodic signals in a nonstationary power series at specific frequencies and how the amplitude varies with time (Daubechies 1990; Torrence & Compo, 1998).

Wavelet analysis was performed for each individual de-trended principal component produced in VPCA. As is suggested for datasets of this nature, a red noise background spectrum was used to account for a trend of increasing power with decreasing frequency (Torrence & Compo, 1998). A Morlet base wavelet was chosen to project the wavelet spectra because the data set is reasonably localized in time and frequency (Najmi & Sadowsky, 1997; Grinsted et al., 2004).

\[
\psi(t) = \left[ \exp(-2i\pi f_0 t) - \exp\left(-2\pi^2 f_0^2 \sigma^2 \right) \right] \exp\left(-\frac{t^2}{2\sigma^2} \right)
\]

**Figure 2.5** Morlet continuous wavelet definition (Najmi & Sadowsky, 1997).
2.8 Assigning Milankovitch Cycles and Linear Sedimentation Rates

Orbital cyclicity is illustrated in the wavelet spectra as the periods that have the highest power units. Milankovitch cycles are assigned in descending order of period length, with the longest peak corresponding to the longest Milankovitch cycle. The dominate Milankovitch forcing and driver of the elemental processes in that wavelet is that which has the most power in the global wavelet. The peaks found in wavelet analysis correspond to lengths of sedimentation as peaks show periods and periods are lengths in the core (Hall, 2015).

Linear sedimentation rates (cm/kyr) were calculated using the lengths of Milankovitch cycles and period length of a cycle identified in wavelet power spectra. Trends were identified in linear sedimentation rates by experimenting with Milankovitch cycle lengths. Wavelets within each dataset were then matched with Milankovitch cycles. Estimated linear sedimentation rates were compared to values derived using known magnetic reversals and change in meter below sea floor values. Accumulation rates (g/cm$^2$/kyr) were calculated by multiplying linear sedimentation rate (cm/kyr) by Shipboard Report average bulk density of a given core (g/cm$^3$).
Chapter 3. RESULTS

3.1 Fine Fraction Mineral Identification

Most fine fraction minerals identified in Cores 697B 13X-17X are siliceous, including smectite, illite, chlorite, chert, chalcedony, quartz, garnet and feldspar.

Table 3.1 shows all identified minerals with core section number and age of deposition. Smectite, in the form of montmorillonite, is found consistently through Cores 697B 13X and 17X, which span 3.8-3.0 Ma. Illite is commonly and evenly distributed through these samples as well. The only gap in illite deposition appears to be at the end of Core 15X (3.5-3.4 Ma). In contrast with previous studies of this area (Robert & Maillot, 1990), chlorite is only identified once and kaolinite is never identified. The mixed-layer clays kaolinite-smectite, chlorite-smectite, and chlorite-vermiculite were also not found. Fine fraction mineral spectra for all samples can be found in Appendix B.

Chert and chalcedony are found consistently throughout the samples. Chalcedony, a cryptocrystalline silicate rock with fine intergrowths of quartz and moganite, was present in the top of Core 13X (deposited 3.07-3.03 Ma) as well as between Cores 697B 15X and 17X (3.77-3.34 Ma). Chert, formed of microcrystalline interlocked quartz grains or fibrous chalcedony (Folk et al., 1952), was also identified in many of these samples. Another form of quartz, aventurin, is found consistently in Core 13X and interspersed in Core 14X. Other forms of quartz could also be present in these sediment samples, but were not found because the ASD FieldSpec FR® spectro-radiometer can only identify a limited number of quartz types.
Feldspar, a common tectosilicate, is distributed evenly throughout Cores 697B 13X and 14X, but mostly absent between the base of Core 14X and the base of Core 17X. Garnets identified here were grossular, which is part of the ugrandite series, and almandine, which is part of the pyralspite series (Deer et al., 1992). Garnets are nesosilicate minerals and characteristic of metamorphic rocks, but also found as detrital grains in some sediments (Poppe et al., 2001). Dipyre, which is a scapolite that forms in the presence of small amounts of chlorine, was identified in Cores 697B 14X and 15X. Wollastonite, a calcium silicate mineral, was identified in Core 14X. The silicate mineral vermiculite was not identified. Apatite, a phosphate mineral, was identified only once. The carbonate minerals calcite and dolomite were not found.

Table 3.1 shows variance in spectra collected on the darker (clay) and lighter (silt) sides of the sample. Figure 3.1 shows SEM images of presumed clay and silt fractions, illustrating the lighter fraction contains larger and more heterogeneous grains. This finding has little bearing on identifying mineral assemblages as no identified mineral displayed a tendency to appear in either the presumed majority silt or clay fraction. Chert, chalcedony, illite, grossular and anorthite are often identified on either the clay or silt side of a sediment sample, but not both. Montmorillonite is more often identified in both the clay and silt sides of the sediment sample. This could be interpreted to mean that montmorillonite grain size is more heterogeneous or that montmorillonite is in such high abundance that it is detected on both sides of the sample. Because the instrument used for this study can only identify presence and not abundance, no quantitative assessments of fine fraction mineral assemblages can be made.
Figure 3.1 Scanning Electron Microscope (SEM) Images of selected fine fraction mineral assemblages. Presumed silt fraction on left and clay on right. Scale is 50µm.
Figure 3.2 Core 697B 13X Spectra for Clay and Silt Sides (Continuum Removed)  
Silt spectra are in lighter colors and clay spectra are in darker colors.

Figure 3.3 Core 697B 13X Spectra for Clay and Silt Sides (Data Value)  
Silt spectra are in lighter colors and clay spectra are in darker colors.
Table 3.1 697B 13X-17X (3.77-3.03 Ma) Fine Fraction Mineral Identification.

<table>
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<th>Age (Ma)</th>
<th>Sample</th>
<th>Apatite</th>
<th>Chlorite</th>
<th>Feldspar</th>
<th>Garnet</th>
<th>Illite</th>
<th>Scapolite</th>
<th>SiO2</th>
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Each category is split into a clay (light grey) and silt (white) section to clarify whether the mineral was identified on the clay or silt side of the fine sediment sample. The presence of a mineral in a particular sample is indicated with the initial of its mineral name under its mineral category as follows:

- Ca – Chlorapatite
- C – Chlorite
- A – Anorthite
- O – Oligoclase
- G – Grossular
- I - Illite
- Aq - Aventurin Quartz
- D – Dipyre
- Ch – Chert
- Cl – Chalcedony
- W – Wollastonite
- M - Montmorillonite

3.2 Elemental Count Ratios

Figure 3.4 compares downcore five-point weighted averages of Fe/Ti and Si/Ti elemental ratios to the LRO4 benthic δ¹⁸O stack, diatom percent area, magnetic susceptibility values, IRD accumulation and biosilica accumulation. Fe/Ti is included as an indicator of wind blown iron dust content, with high values pointing to conditions where iron is not a limiting nutrient and paleoproductivity is high (Dezileau et al., 2004). Si/Ti is indicative of non-terrigenous siliceous material. The LRO4 stack is included as a proxy of global temperature and glacial cover (Lisiecki & Raymo, 2005). Biosilica extraction was performed at Wesleyan in 2015-2016 and diatom percent area is from Pudsey, 1990. Coarse fraction weight percent is included as an indicator of IRD and magnetic susceptibility is included for comparison to IRD.
Figure 3.4 Comparison of Fe/Ti and Si/Ti to Paleoclimate Proxies.
Units I, II, III and IV depict distinct climate regimes. Interpretation of units can be found in Table 4.1. Gaps in data at ~3.1-3.2 Ma and ~3.65-3.7 Ma of magnetic susceptibility, Si/Ti and Fe/Ti data sets due to inability to scan poorly recovered cores. Note LRO4 Stack low values indicate global warming and high values indicate global cooling. LRO4 Source: Lisieki & Raymo (2005) and Diatom % Content Source: Pudsey (1990).
3.3 Varimax-rotated Principal Component Composition

A total of fifty varimax-rotated principal components (VPCs) were created for this study. For each of the five ODP Site 697B Cores 13X-17X, two types of VPCs were created. The first set of five VPCs was created treating all elemental count ratio data for a core section as one dataset. This grouping will be referred to as “DR” to indicate the components were detrended and rotated. The second set of five VPCs was created treating each individual core section (i.e. 14X_1 through 14X_7) as a separate data set. The VPCs for each core section where compiled afterwards into one dataset to show the five VPCs for that core. This grouping will be referred to as “DR C” to indicate that the components were detrended, rotated and compiled.

Table 3.2 shows the DR five largest elemental count ratio loadings with their loading values for Cores 697B 13-17. DR values are displayed so changes can be compared between cores and not just core sections, as DRC values would show. Component one holds the most influence, followed by component two with the second most influence, and so on. These values are displayed for the DR and DR C VPCAs of each core section. A positive loading value indicates the subject is positively correlated with principal components while a negative sign indicates a negatively correlation. Figures 3.5 through 3.7 compare DR and DR C downcore component values. The DR values for each core section are displayed sequentially. The graphs exhibit large variance in some areas, showing that these two methods yield different principal components. Neither method has more visible variance in the dataset than the other, but Table 3.3 shows that DR values have over 30% red noise in
wavelet spectra more often than DR C values. Therefore, DR values may be less reliable.

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<th>VPC</th>
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<td>Element/Ti (Loading Value)</td>
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<tr>
<td>DR 1</td>
<td>Y (0.97) Zr (.96) Mo (-0.95) Ni (0.92) Cu (-0.92)</td>
</tr>
<tr>
<td>DR 2</td>
<td>Ca (0.96) Sr (0.85) Ba (-0.83) Mn (0.89) Al (0.66)</td>
</tr>
<tr>
<td>DR 3</td>
<td>Cl (0.85) S (0.84) Ar (0.78) Br (0.71) Cr (-0.57)</td>
</tr>
<tr>
<td>DR 4</td>
<td>Bi (0.79) Rb (-0.68) Pb (-0.46) Ni (0.35) Cu (-0.35)</td>
</tr>
<tr>
<td>DR 5</td>
<td>P (-0.65) Fe (0.69) Si (0.49) S (0.34) K (0.32)</td>
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<tr>
<td><strong>14X</strong></td>
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<td>DR 1</td>
<td>Br (0.95) Ar (0.95) Ba (0.94) Sr (0.91) Rh (0.89)</td>
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<tr>
<td>DR 2</td>
<td>Ni (0.96) Ga (0.96) Cu (-0.95) Zr (0.94)</td>
</tr>
<tr>
<td>DR 3</td>
<td>Fe (0.92) S (0.87) P (-0.62) K (0.48) Ph (0.28)</td>
</tr>
<tr>
<td>DR 4</td>
<td>Ca (0.88) Al (0.82) Rb (0.41) K (0.39) P (0.39)</td>
</tr>
<tr>
<td>DR 5</td>
<td>Cr (0.73) Cl (-0.59) S (-0.33) Rh (0.34) Zn (-0.32)</td>
</tr>
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<td><strong>15X</strong></td>
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<tr>
<td>DR 3</td>
<td>S (0.90) Fe (0.90) P (-0.65) K (0.47) Bi (0.32)</td>
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<td>DR 4</td>
<td>Ca (0.87) Al (0.86) K (0.39) P (0.34) Mn (0.31)</td>
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<td>DR 5</td>
<td>Cr (0.84) Cl (-0.57) Zn (-0.35) P (0.32) S (-0.31)</td>
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<tr>
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<td>DR 2</td>
<td>Ca (0.86) Ba (-0.80) Ar (-0.79) Rh (0.66) Sr (0.59)</td>
</tr>
<tr>
<td>DR 3</td>
<td>Cl (0.88) Br (0.76) Sr (0.59) K (-0.59) Al (-0.54)</td>
</tr>
<tr>
<td>DR 4</td>
<td>S (-0.94) P (0.91) K (0.62) Al (0.52) Rh (0.33)</td>
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<td>DR 5</td>
<td>Bi (0.81) Fe (0.68) Pb (-0.41) Zr (0.37) Cr (-0.30)</td>
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<tr>
<td><strong>17X</strong></td>
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<tr>
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<td>Ni (0.98) Cu (-0.98) Ga (0.98) Y (0.97) Zr (0.97)</td>
</tr>
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<td>DR 2</td>
<td>Ca (0.93) Sr (0.86) Ba (-0.85) Ar (-0.76) Rh (0.73)</td>
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<tr>
<td>DR 3</td>
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<td>DR 5</td>
<td>Cl (0.91) Ar (0.54) Rh (-0.50) Cr (0.34) Sr (0.21)</td>
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Table 3.2 Varimax-rotated Principal Component Loadings of Elemental Count Ratios.
Figure 3.5 Downcore comparisons of Detrended Varimax-rotated Principal Components DR1 v. DR1 C and DR2 v. DR2 C against two y-axis mbsf (meters below sea floor) and age (millions of years ago). 128 mbsf is ~3.0 Ma in age and the top of Core 13X.
Figure 3.6 Downcore comparisons of Detrended Varimax-Rotated Principal Components DR3 v. DR3 C and DR4 v. DR4 C against two y-axis mbsf (meters below sea floor) and age (millions of years ago). 128 mbsf is ~3.0 Ma in age and the top of Core 13X.
Figure 3.7 Downcore comparisons of Detrended Varimax-Rotated Principal Components DR5 and DR5 C against two y-axis mbsf (meters below sea floor) and age (millions of years ago). 128 mbsf is ~3.0 Ma in age and the top of Core 13X.
3.4 Wavelet Power Spectra

The VPCs for DR and DR C data sets were individually input into an Interactive Wavelet Plot to create wavelet spectra. An example wavelet spectra is included in Figure 3.8. All spectra for Cores 697B 13X-17X DR and DR C can be found in Appendix C. In each wavelet figure, Part A shows the interpolated and detrended data for the component. Part B shows the wavelet power spectrum color-coded based on the wavelet’s power. Part C shows the variance in each wavelet, overlaid on a line that represents 30% red noise. Peaks with variance above the line can confidently be considered signs of orbital cyclicity. Peaks with variance below the line are marked in red in Table 3.3 and could be signs of orbital cyclicity, but our confidence interval is smaller (Torrence & Campo, 1998).

Table 3.3 shows the linear sedimentation rates (centimeters per thousand years) calculated for each VPC one through five for each peak (either one, two or three) in the wavelet power spectrum. For each peak, the length of the period of sedimentation (cm) as well as the length of Milankovitch cycle (kyr) assigned to it is listed. The linear sedimentation rate (cm/kyr) for each peak is calculated by dividing its length of sedimentation (cm) by Milankovitch cycle length (kyr) and multiplying that value by 1,000. The sedimentation rates for each peak calculated in DR and DR C are averaged to give the averaged sedimentation rate of the core. The average standard deviation divided by average sedimentation rate is included for each core section to show variance between each principal component from the average value. Values marked in red include all peaks, including those with over 30% red noise, while values marked in black only include values that are confidently considered orbital cycles. Figure 3.9 shows how linear sedimentation rates calculated in this
study compare to linear sedimentation rates calculated using magnetostratigraphy of modern MBSF calculations and magnetic reversals, and magnetostratigraphy of 1987 shipboard MBSF values and magnetic reversals. Figure 3.10 compares these values for accumulation rates over time.

**Figure 3.8** Wavelet Spectra for Core 15X DR5 ("697B 15X DR5 1-7") and DR1 C ("697B 15X DR5 Compiled"). Black lines added. See all others in Appendix C.
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<th>Length (cm)</th>
<th>Sed Rate (cm/kyr)</th>
<th>M Cycle (kyr)</th>
<th>Length (cm)</th>
<th>Sed Rate (cm/kyr)</th>
<th>Avg Std Dev</th>
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<td>168.7-176.6</td>
<td>17X</td>
<td>1</td>
<td>1</td>
<td>100</td>
<td>450</td>
<td>4.500</td>
<td>19</td>
<td>120</td>
<td>6.316</td>
<td>0.17</td>
<td>6.193</td>
<td>0.30</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2</td>
<td>2</td>
<td>41</td>
<td>250</td>
<td>6.098</td>
<td>41</td>
<td>200</td>
<td>4.878</td>
<td>2</td>
<td>41</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3</td>
<td>1</td>
<td>23</td>
<td>175</td>
<td>7.609</td>
<td>19</td>
<td>120</td>
<td>6.316</td>
<td>3</td>
<td>23</td>
</tr>
<tr>
<td></td>
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<td></td>
<td>4</td>
<td>1</td>
<td>100</td>
<td>450</td>
<td>4.500</td>
<td>19</td>
<td>120</td>
<td>6.316</td>
<td>2</td>
<td>19</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>5</td>
<td>1</td>
<td>23</td>
<td>175</td>
<td>7.609</td>
<td>19</td>
<td>120</td>
<td>6.316</td>
<td>2</td>
<td>41</td>
</tr>
</tbody>
</table>

Table 3.3 Average Pliocene sedimentation rates 3.78-3.03 Ma at ODP Site 697B.
Figure 3.9 Comparison of Calculated Linear Sedimentation Rates Over Time. Ages in millions of years ago for mid-point of each core listed.

Figure 3.10 Comparison of Calculated Accumulation Rates Over Time. Ages in millions of years ago for mid-point of each core listed.
Chapter 4. DISCUSSION

4.1 Interpretation of Fine Fraction Mineral Associations

Our finding of consistent smectite deposition is consistent with Robert & Maillot (1990) who found abundant smectite in the clay fraction of this site using X-Ray Diffraction (Table 1.2). The same study also found abundant smectite at ODP Site 693, 696 and 695 (Figure 1.7 for locations). ODP Site 693 on the East Antarctic Margin shows an increase in smectite deposition in the late Miocene (5.5 Ma), which is attributed to increased erosion by ice and/or currents due to an increase in West Antarctic ice volume (Robert & Maillot, 1990).

ODP Site 1165 in the Prydz Bay shows high detrital smectite deposition from bottom water transport during peak Mid-Pliocene warming (Junttila & Strand, 2006). ODP Site 695 shows an increase in smectite deposition at ~4.4 Ma and ODP Site 697 shows an increase in smectite abundance after 4.2 Ma (Robert & Maillot, 1990). In the Early Pliocene (4.4-4.2 Ma), smectite was eroded from the Antarctic margin, transported by AABW and WBW and deposited in the Jane Basin and other Weddell Sea locations (Robert & Maillot, 1990). Smectite formation and deposition is aided by ice retreat conditions (Verma et al., 2014). Pudsey et al. (1987) similarly interprets higher silt presence in Jane Basin diatomaceous sediments to suggest that bottom water flow increases during interglacial periods. Thus consistent smectite deposition in Cores 697B 13X-17X is interpreted to indicate high AABW flow and warming conditions between 3.8 and 3.0 Ma.

Common illite identification in this study is consistent with Robert & Maillot (1990), who found common to abundant illite through these sediment samples.
Chlorite was identified only once, conflicting with their findings of common to abundant chlorite. Kaolinite was not found in these sediment samples, which also conflicts with their findings of common to rare kaolinite. A possible explanation for the discrepancy is that the 1990 studies analyzed a \(<2\mu m\) fraction using a Philips Diffractometer. This study analyzed the \(<63\mu m\) sample with a spectro-radiometer. The difference in particle size as well as method could aid the identification of certain minerals and impede the identification of others.

A recent study of East Antarctica Wilkes Land clay \((<2\mu m)\) and silt \((2–53\mu m)\) mineral assemblages found presence of smectite and kaolinite only in the clay size fraction and chlorite and illite in both silt and clay fractions (Verma et al., 2014). This could explain why kaolinite was not identified in this study, but raises further questions as to why smectite was found. They attribute smectite presence only in the clay size fraction to an effective role of sorting due to deposition from distal source in ice retreat conditions (Verma et al., 2014). Similarity in the crystallinity and chemistry of illite in both fractions is interpreted as a sign of negligible sorting, which points to deposition from the waxing of ice sheets (Ehrmann et al., 1992; Junnttila et al., 2005). We interpret consistent deposition of both illite and smectite at ODP Site 697 between 3.8 and 3.0 Ma to indicate dynamic AIS behavior throughout this period. Because this study measured presence and not abundance of fine fraction minerals, the duration of warming and cooling periods cannot be further defined.

4.2 Comparison of Mineral Associations and Paleoproductivity Indicators

In the absence of mineral abundance data, fine fraction mineral assemblages are compared to elemental count ratios and paleoproductivity data to explore any
further relationship between fine fraction mineral assemblages and their depositional environment. Table 4.1 from Kaufman et al. (2016) interprets IRD and biosilica accumulation data as indicating four distinct climatic units. Additional interpretation of diatom, magnetic susceptibility, Si/Ti and Fe/Ti data from this study is included in italics. Across all datasets, depositional Unit III (3.56-3.34 Ma) points to dynamic, possibly obliquity driven, ice sheet behavior and sea ice cover. Si/Ti and Fe/Ti elemental count ratios, biosilica accumulation, coarse fraction accumulation and diatom percent content all display a peak at ~3.35 Ma. Fine fraction mineral assemblages at this time show montmorillonite and not illite presence, which supports our interpretation of peak warming (Figure 3.4, Core 15X 1 Sample 82-84).

In each dataset a dramatic decline follows this peak, which could be interpreted as rapid cooling. The LRO4 stack indicates peak cooling conditions at ~3.31 Ma and an overall warming trend thereafter. Fine fraction mineral data also supports this interpretation with illite deposition from 3.33-3.29 Ma (Core 14X Samples 7_28-30, 6_129-131, 5_94-96 and 4_136-139). However, smectite is also deposited through these samples. Coinciding peaks in biosilica accumulation, diatom percent and Fe/Ti point to dynamic changes in paleoproductivity and sea ice cover through 3.22 Ma. Kaufman et al. (2016) interprets Unit II and I low IRD and high paleoproductivity indicators to indicate peak warming that melts EAIS IRD bearing icebergs before they reach the Jane Basin. Dynamic Fe/Ti, Si/Ti and diatom depositional behavior would also support this hypothesis.
<table>
<thead>
<tr>
<th>Unit</th>
<th>Age (Ma)</th>
<th>Description</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>I 3.03-3.16</td>
<td>Low IRD accumulation, highly variable biosilica accumulation. Peak in Fe/Ti, mag sus and biosilica at 3.09 Ma</td>
<td>Variable productivity with iceberg delivery prevented by melting ice or lack of ice.</td>
<td></td>
</tr>
<tr>
<td>II 3.16-3.34</td>
<td>Low IRD accumulation, increasing biosilica accumulation. Diatom peaks coincide with biosilica peaks and one Fe/Ti peak</td>
<td>An open ocean (high productivity) environment with iceberg delivery prevented by melting ice or lack of ice.</td>
<td></td>
</tr>
<tr>
<td>III 3.34-3.56</td>
<td>Variable IRD &amp; biosilica accumulation, with biosilica peaks leading IRD peaks by 3-10 kyr. Peaks increase in amplitude up section with ~50 kyr pacing. Diatom peaks coincide with IRD and biosilica peaks Fe/Ti, Si/Ti, biosilica, IRD and diatoms peak at ~3.35 Ma</td>
<td>Cyclical changes in productivity and ice-rafting are occurring simultaneously, possibly obliquity-driven. ~3.35 Ma marks peak warming conditions, minimum sea ice cover and high paleoproductivity.</td>
<td></td>
</tr>
<tr>
<td>IV 3.56- 3.78</td>
<td>Variable and low IRD accumulation, consistently low biosilica. Diatom % content also variable and low</td>
<td>Larger sea ice extent with some land-derived ice rafting.</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.1 Interpretation of Depositional Units outlined in Figure 3.4. Table from Kaufman et al. (2016). Supplemental interpretations from this study added in italics.

Figure 4.1 compares Si/Ti, biosilica and overall fine fraction accumulation to identified fine fraction minerals. A spike in fine fraction and biosilica accumulation at 3.46 Ma corresponds to Core 15X Sample 6_62-64 where no illite, but montmorillonite, chert, and chalcedony are identified. This relationship is mirrored again at 3.35 Ma where a peak in fine fraction and biosilica accumulation as well as in the Si/Ti ratio corresponds to Core 15X Sample 1_82-84 where montmorillonite, chalcedony and no illite are identified. Naturally, higher biosilica accumulation would have a relationship with higher silica mineral deposition as well as higher Si/Ti values. However, because we cannot compare silica mineral abundance with biosilica accumulation, this relationship should not be overstated.
Figure 4.1 Comparison of Siliceous Minerals, Si/Ti and Biosilica Accumulation. Each value marked on Si/Ti, biosilica and fine fraction accumulation datasets coincides to a sample number. Si/Ti is weighted on a five point moving average.

To further explore this relationship, Figure 4.2 compares chert deposition and biosilica accumulation. Chert forms from diagenesis of biosilica, so high biosilica accumulation values in sediment samples with chert would point to local chert formation (Behl & Smith, 1992). This plot shows biosilica accumulation rates to be variable in samples where chert was identified, showing that the chert found in these samples was likely not locally formed. Moreover, we see complete diatoms in these samples and in areas of high local chert formation diatoms would be fractured and disturbed.
Figure 4.2 Comparison of biosilica accumulation and chert presence as identified in spectral reflectance.

In Figure 4.3 Pudsey (1990) sand percent content, which is indicative of IRD content, is compared in a crossplot with diatom percent area of a smear slide. Figure 4.4 compares similarly coarse fraction accumulation with biosilica accumulation. Neither crossplot displays a clear linear relationship between increasing paleoproductivity and increasing IRD deposition, showing that this relationship is complex and changes through time. Many factors in addition to temperature, including bottom water flux, surface currents, wind-blown Fe and other factors could affect transport of IRD and diatom growth. Moreover, analysis of elemental counts ratios and Milankovitch cycles could help constrain our temporal understanding of climatic changes during this period and elucidate driving forcings behind them.
Figure 4.3 Crossplots of Sand % Content and Diatom % Area (Pudsey, 1990).

Figure 4.4 Crossplots of Biosilica v. Coarse Fraction (>63 μm) Accumulation.
4.3 Interpretation of Elemental Count Ratio Principal Components

The dominance of land-based elements, including Ba and dissolved trace metals Ni, and Cu (Saager et al., 1997) in DR1 points to a terrigenous sourcing of these sediments (Table 3.2). Continental weathering from glacial advance or retreat likely led to their deposition in the Weddell Sea. Positively correlated loadings of Fe, Si, Ba and Ca furthermore point to paleoproductivity (Sun et al., 2008). High loadings of Mn and Ca also indicate sediment deposition under oxic bottom water conditions (Calvert and Pedersen, 1993). Variations in Zr loadings could reflect the relative contribution of coarse and fine-grained materials in the terrigenous component of these sediments. Barium holds one of the largest loadings in DR1 of Cores 14X and 15X, at a positive correlation of 0.94 and 0.95, respectively. In Core 14X, barium is negatively correlated at -0.83 in DR2. We interpret this to be a potential sign of decreased productivity in Core 13X (3.14-3.03 Ma) compared to Cores 14X and 15X (3.48-3.21 Ma). This could point to the onset of a cooling period of increased sea ice cover.

4.4 Sedimentation Rates and Dominant Milankovitch Forcings

Linear sedimentation rates derived using wavelet analysis are comparable to sedimentation rates derived using magnetostratigraphy (Figure 3.10). These values are different from 1987 shipboard measurements, which is to be expected as this study uses an updated geomagnetic polarity reversal timeline (Gee & Kent, 2007). Sedimentation rates for Cores 697B 14X-17X (3.79-3.21 Ma) range from 6.14 cm/kyr to 6.54 cm/kyr. At Core 13X (3.14-3.03 Ma), the sedimentation rate
decreases to 4.76 cm/kyr. This change points to a shift in depositional processes at 
~3.2 Ma, which is also reflected in a change in dominant Milankovitch forcings.

Results from the wavelet power spectra for these cores shows a transition 
from a dominant eccentricity forcing with obliquity influence to a dominant 
precession forcing at ~3.3 Ma. Cores 697B 17X (3.79- 3.66 Ma) and 16X (3.61- 3.48 
Ma) are driven by a dominant eccentricity forcing with a secondary obliquity forcing. 
Eccentricity is assigned to eight of the ten dominant peaks identified in Core 17X and 
seven of the ten dominant peaks identified in Core 16X. Obliquity is assigned to eight 
of the ten secondary peaks in Core 17X and six of the seven secondary peaks in Core 
16X. It is important to note that Core 17X has a high number of values with over 30% 
red noise, decreasing our confidence in our results for this core.

Cores 697B 14X (3.33- 3.21 Ma) and 15X (3.48- 3.33 Ma) show a transition 
to a dominant precession forcing and secondary eccentricity forcing. Five of the ten 
dominant peaks identified in Core 15X are assigned to eccentricity while the other 
five are assigned to precession. Only two of the Core 15X seven secondary peaks are 
assigned to obliquity while the other five are assigned to precession. In Core 14X, 
eccentricity is assigned to three, precession is assigned to six and obliquity is assigned 
to one of the dominant peaks identified. Obliquity is assigned to two of the six 
secondary peaks, with precession assigned to the remainder.

Core 13X (3.14- 3.03 Ma) shows a strong adherence to a dominant precession 
cycle and secondary eccentricity cycle. Precession is assigned to five of the ten 
dominant peaks identified in Core 13X. It is also assigned to four of the five
secondary peaks identified. Of the remaining five dominant peaks, eccentricity is
assigned twice and obliquity is assigned three times.

4.5 Interpretation of Milankovitch Forcings

A similar study of Dronning Maud Land ODP Site 693 by Hall, 2015 and
True-Alcalá, 2015 find a dominant eccentricity forcing and secondary precession
forcing in cores deposited between 4.2 and 3.6 Ma. The absence of paleomagnetics
prevents further time constraint of their data. Also analyzing EAIS sediments,
Patterson et al. (2014) finds evidence of a transition from a dominant obliquity
forcing to an eccentricity and precession at ~3.5 Ma. The LRO4 benthic δ18O stack
also shows a reduction of obliquity variance at ~3.5 Ma (Lisiecki & Raymo, 2005).

The reduction of obliquity power at ~3.3 Ma and growing influence of
precession and eccentricity in East Antarctic sedimentation is interpreted to reflect a
decline in EAIS sensitivity to oceanic forcings with southern high-latitude cooling
(Passchier, 2011; Patterson et al., 2014). However, WAIS studies show obliquity-
paced glacial retreat retreated continuing to 3.0 Ma (Naish et al., 2009). ODP Site 697
receives sedimentation from both the WAIS and EAIS, but shows greater coherency
with studies of EAIS behavior during this ~3.3 Ma critical climatic point.
Chapter 5. CONCLUSION

5.1 Pliocene Climatic Optimum and Implications for Current Climate Change

Here we show ~3.3 Ma marked a critical threshold in Antarctic glacial behavior and climatic conditions. Kaufman et al. (2016) finds evidence of increased warming after this period while this study finds evidence of precession-paced cooling, as supported by Passchier (2011) and Patterson et al. (2014). These findings have implications for our understanding of the Pliocene Climatic Optimum. Paleoproductivity indicators including barium elemental loadings, diatom percent content, biosilica, Si/Ti and Fe/Ti ratios point to warm and productive marine conditions from 3.6 Ma to 3.35 Ma (Unit III of Figure 3.4). Consistent smectite fine fraction mineral deposition through this period also points to a warm climate. Our site records EAIS dynamism during Pliocene warmth and a growing influence of eccentricity and precession on EAIS at ~3.3 Ma, followed by a decline in EAIS sensitivity to oceanic forcings with climatic cooling. Moreover, our data supports a dynamic EAIS in climatic conditions similar to today’s.

5.2 Future Work

While this site has a high sedimentation rates for a pelagic area, age resolution is not high enough to identify precession and obliquity cycles with high confidence. The average standard deviation from the average sedimentation rate for each core section is moderate, with the lowest value for Core 13X at 0.127 and highest value for Core 16X at 0.227. Improved aged dating would increase the confidence of these findings and likely lower average standard deviation rates.
A study of abundance of fine fraction minerals rather than just presence would further our interpretation of AABW behavior. A similar study of this site for cores deposited between 4.4 and 3.8 Ma would help constrain the dates of the Pliocene Climatic Optimum, as Fedorov et al. (2013) and Pagani et al. (2009) cite peak warmth during this interval. XRF scans at 50 kv voltage for more core sections would provide more data on Ba content, which would also improve our understanding of paleoproductivity during this period. Moreover, identification of diatoms in Site 697B Core 13X and their affinities for cold or warm ocean temperatures would help constrain if 3.3-3.0 Ma was indeed a period of warming, as supported by Kaufman et al. (2016) or cooling, as supported by this work.
Appendix A. List of Acronyms

DR – Detrended and varimax-rotated principal components
EAIS – East Antarctic Ice Sheet
IODP—International Drilling Program (formerly ODP)
IRD—Ice Rafted Detritus
Ma – Million years ago
MBSF—Meters Below Sea Floor
ODP—Ocean Drilling Program (now IODP)
PCA—Principal Component Analysis
PCO—Pliocene Climatic Optimum
TAM—Transantarctic Mountains
WAIS – West Antarctic Ice Sheet
VPC – Varimax-rotated Principal Component
VPCA – Varimax-rotated Principal Component Analysis
Appendix B. Fine Fraction Mineral Spectra

During mineral identification spectra were analyzed separately, but for ease of comparison all spectra for one core are included together. Note the dip in absorption value at ~950µm is due to an instrument error, an absorption at 1900µm is due to H$_2$O content and an absorption at 1400µm is due to OH$^-$ presence. Absorptions between 2200 and 2300µm are usually diagnostic of a certain mineral type.

Site 697B Core 13X Fine Fraction Spectra
697B Core 14X Fine Fraction Spectra

Site 697B Core 15X Fine Fraction Spectra
Site 697B Core 16X Fine Fraction Spectra

Site 697B Core 17X Fine Fraction Spectra
Appendix C. Wavelet Power Spectra

Wavelet spectra are included for Site 697B Cores 13X-17X below for VPCs 1-5. “DR” components are listed as DR1 (indicating VPC number) 1-4 (indicating core sections included). “DR C” components are listed as DR1 Compiled. For example, the wavelet below is for Core 697B 13X, detrended varminax-rotated principal component 1 created by running core sections 1-4 through VPCA as separate data sets, so with the “DR” method. In this specific instance, only core sections 1-4 are included because Site 697B Core 13X 5-7 were too poorly recovered to XRF scan.

In each wavelet figure, Part A shows the interpolated and detrended data for the component. Part B shows the wavelet power spectrum color-coded based on the wavelet’s power. Part C shows the variance in each wavelet, which is another way to show the variance in the wavelet, overlaid on a line that represents 30% noise.
a. 697B 13X DR3 1-4

b. Wavelet Power Spectrum

c. Global Wavelet

Powered by IDL
a. 697B 14X DR1 1-7

Data (units)

-10
-5
0

mbsf (meters)

140 142 144 146

Morlet 6.00
Real (solid) Imaginary (dash)

b. Wavelet Power Spectrum

Period (meters)

0.062
0.125
0.250
0.500
1.000
2.000
4.000
8.000

mbsf (meters)

140 142 144 146

Variance (units)² 10³

Powered by IDL

c. Global Wavelet

a. 697B 14X DR1 Compiled

Data (units)

0
-10

mbsf (meters)

140 142 144 146

Morlet 6.00
Real (solid) Imaginary (dash)

b. Wavelet Power Spectrum

Period (meters)

0.062
0.125
0.250
0.500
1.000
2.000
4.000
8.000

mbsf (meters)

140 142 144 146

Power (units)²

0.0 0.030 0.52 3.2 15

Powered by IDL

c. Global Wavelet
80
Appendix D. RStudio Scripts

Creating Elemental Count Ratio Varimax-rotated Principal Components

1. Setting working folder and importing dataset

```r
setwd("~/Desktop/Wavelet/13X")
wavelet_prep <- read.delim("697B_13X_1-7_TiArInt.txt", na.strings="#N/A")
```

2. Loading necessary packages

```r
library(psych)
library(GPArotation)
library(calibrate)
library(akima)
library(pracma)
library(foreign)
```

3. Creating varimax-rotated principal components

```r
VPCA1 <- principal(wavelet_prep[,14:39], nfactors=5, rotate="varimax")
wavelet_prep$RC1 <- VPCA1$scores[,1]
wavelet_prep$RC2 <- VPCA1$scores[,2]
wavelet_prep$RC3 <- VPCA1$scores[,3]
wavelet_prep$RC4 <- VPCA1$scores[,4]
wavelet_prep$RC5 <- VPCA1$scores[,5]
```

4. Interpolating principal components across shipboard meter below sea floor values

```r
wavelet1 <- aspline(wavelet_prep$mbsf, wavelet_prep$RC1, wavelet_prep$Interpolated)
wavelet2 <- aspline(wavelet_prep$mbsf, wavelet_prep$RC2, wavelet_prep$Interpolated)
wavelet3 <- aspline(wavelet_prep$mbsf, wavelet_prep$RC3, wavelet_prep$Interpolated)
wavelet4 <- aspline(wavelet_prep$mbsf, wavelet_prep$RC4, wavelet_prep$Interpolated)
wavelet5 <- aspline(wavelet_prep$mbsf, wavelet_prep$RC5, wavelet_prep$Interpolated)
```

5. Applying polynomial trend line (TR) to the data and detrending (DR) in order to look at changes in the rate of change as opposed to changes themselves

```r
wavelet <- data.frame(wavelet1)
wavelet$mbsf <- wavelet$x
wavelet$R1 <- wavelet$y
```
wavelet <- wavelet[3:4]
wavelet$R2 <- wavelet2$y
wavelet$R3 <- wavelet3$y
wavelet$R4 <- wavelet4$y
wavelet$R5 <- wavelet5$y

p1 <- polyfit(wavelet$mbsf, wavelet$R1, 2)
p2 <- polyfit(wavelet$mbsf, wavelet$R2, 2)
p3 <- polyfit(wavelet$mbsf, wavelet$R3, 2)
p4 <- polyfit(wavelet$mbsf, wavelet$R4, 2)
p5 <- polyfit(wavelet$mbsf, wavelet$R5, 2)

wavelet$TR1 <- polyval(p1, wavelet$mbsf)
wavelet$TR2 <- polyval(p2, wavelet$mbsf)
wavelet$TR3 <- polyval(p3, wavelet$mbsf)
wavelet$TR4 <- polyval(p4, wavelet$mbsf)
wavelet$TR5 <- polyval(p5, wavelet$mbsf)

wavelet$DR1 <- wavelet$R1 - wavelet$TR1
wavelet$DR2 <- wavelet$R2 - wavelet$TR2
wavelet$DR3 <- wavelet$R3 - wavelet$TR3
wavelet$DR4 <- wavelet$R4 - wavelet$TR4
wavelet$DR5 <- wavelet$R5 - wavelet$TR5

6. Exporting data set

write.csv(wavelet, "697B_13X_1-7_DR.csv")

Converting .ms Magnetic Susceptibility Files to .csv Files

1. Setting working folder and importing dataset

setwd("/Volumes/TOURO MOBIL/Magnetic susceptibility/MS Data 2015")

a = readLines("697B_2H_7_697B_2H_7_20150607143016.ms")
p = "offset=((0-9]+\([0-9]\)+), magnetic_susceptibility=((0-9]+\([0-9]\)+),
time_stamp=(.*)"
r = grepl(p, a, perl=TRUE)
sum(r)
m = gregexpr(p, a, perl=TRUE)
m[[35]]
p2 = "text_id=.*"
m2 = gregexpr(p2, a)
str(m2)
k = which(m2 > 0)
k
s = regmatches(a[k], m2[[k]])
txt_id = sub("text_id=", ",", s)
n_entries = sum(r)

2. Creating magnetic susceptibility data frame

df = data.frame(stringsAsFactors=FALSE)

for (i in 1:length(a)) {
  ## r is the vector of logical values r[i] is TRUE if the pattern was matched in line a[i]
  if (r[i]) {
    w = attributes(m[[i]])

    3. Extract the contents of the first, second and third capture groups

    s1 = w$capture.start[1]
s2 = w$capture.length[1]
    offset = substr(a[i], s1, s1+s2-1)
    ## as.numeric() converts a text string into a number
    offset_n = as.numeric(offset)

    s1 = w$capture.start[2]
s2 = w$capture.length[2]
    ms = substr(a[i], s1, s1+s2-1)
    ## as.numeric() converts a text string into a number
    ms_n = as.numeric(ms)

    s1 = w$capture.start[3]
s2 = w$capture.length[3]
    ## we don't need time_stamp as a number, so we do not convert it
    t_stamp = substr(a[i], s1, s1+s2-1)

    4. Appending the new row to the data frame

    df = rbind(df, data.frame(text_id = txt_id,
                              offset = offset_n,
                              mag_s = ms_n,
                              time_stamp = t_stamp,
                              stringsAsFactors=FALSE)) 
  }
}

5. Exporting .csv file

write.csv(df, "697B_2H_7.csv", row.names=FALSE)
Works Cited


Hillenbrand, C-D., and Werner Ehrmann. "Distribution of clay minerals in drift sediments on
the continental rise west of the Antarctic Peninsula, ODP Leg 178, Sites 1095 and 1096." Barker, PF, Camerlenghi, A., Acton, GD, and Ramsay, ATS (Eds.), Proc. ODP, Sci. Results, 178.


Iacoviello, Francesco; Giorgetti, Giovanna; Nieto, Fernando; and Memmi, Isabella Turbanti, "Evolution with depth from detrital to authigenic smectites in sediments from AND-2A drill core (McMurdo Sound, Antarctica)" (2012).


Landsat Image Mosaic of Antarctica (LIMA) by NASA “Meet Antarctica”
www.lima.nasa.gov/antarctica.


USGS Open-File Report 2006-1195 “Nomenclature”


