East Antarctic Ice Sheet and Southern Ocean Response
to Orbital Forcing from Late Miocene to
Early Pliocene, Wilkes Land, East Antarctica

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ABSTRACT

Geological and ice sheet models indicate that marine-based sectors of the East Antarctic Ice Sheet (EAIS) were unstable during periods of moderate climatic warmth in the past. While geological records from the Middle to Late Pliocene indicate a dynamic ice sheet, records of ice sheet variability from the comparatively warmer Late Miocene to Early Pliocene are sparse, and there are few direct records of Antarctic ice sheet variability during this time period. Sediment recovered in Integrated Ocean Drilling Program U1361 drill core from the Wilkes Land margin provides a distal but continuous glacially-influenced record of the behaviour of Antarctic Ice Sheets.

This thesis presents marine sedimentological and x-ray fluorescence geochemical datasets in order to assess changes in the dynamic response of the EAIS and Southern Ocean productivity in the Wilkes Land sector during Late Miocene and Early Pliocene to climatic warming and orbital forcing between 6.2 and 4.4 Ma. Two primary lithofacies are identified which can be directly related to glacial–interglacial cycles; enhanced sedimentation during glacials is represented by low-density turbidity flows that occurred in unison with low marine productivity and reduced iceberg rafted debris. Interglacial sediments contain diatomaceous muds with short-lived, large fluxes of iceberg rafted debris preceding a more prolonged phase of enhanced marine productivity. Interglacial sediments coincide with a more mafic source of terrigenous sediment, interfered to be associated with an inland retreat of the ice margin resulting in erosion of lithologies that are currently located beneath the grounded EAIS. Poleward invigoration of the Antarctic Circumpolar Current during glacial–interglacial transitions is proposed to have intensified upwelling, enhancing nutrient availability for marine productivity, and increasing oceanic heat flux at the ice margin acting to erode marine ice sheet grounding lines and triggering retreat.

Spectral analysis of the datasets indicated orbital frequencies are present in the iceberg rafted debris mass accumulation rates at all three Milankovitch frequencies, with a dominant 100 kyr eccentricity driven ice discharge. Prolonged intervals of marine productivity correlate to 100 kyr cyclicity occurring at peaks in obliquity. The response of both ice sheet and biological systems to 100 kyr cyclicity may indicate eccentricity-modulated sea ice extent controls the influx of warm water onto the continental shelf.

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Text Abbreviations

AABW	Antarctic Bottom Water
ACC	Antarctic Circumpolar Current
AIS	Antarctic Ice Sheet
Al/Ti	a provenance or grain size proxy; the ratio of aluminium to titanium
AND-2A	second Antarctic Drilling Project hole
ANDRILL	ANtarctic DRILLing Project
APC	advance piston coring system
APF	Antarctic Polar Front
APFZ	Antarctic Polar Front zone
ASB	Aurora Subglacial Basin
ASF	Antarctic Slope Front
Ba/Al	a marine productivity proxy; the ratio of barium to aluminium
BaSO ₄	barite
BP	years before present
BSi	biogenic silica
CaCO ₃	calcium carbonate
CDW	Circumpolar Deep Water
CO ₂	carbon dioxide
cs%	coarse sand percent
DI	deionised
DIC	dissolved inorganic carbon
DBD	dry bulk density
EAIS	East Antarctic Ice Sheet
EHA	evolutive harmonic analysis
EOCT	Eocene–Oligocene Climatic Transition
ETP	eccentricity tilt precession
FLIP	Ferrar Large Igneous Province
GRA	bulk density
H ₂ O	water
H ₂ O ₂	hydrogen peroxide
HCl	hydrochloric acid
HSSW	high salinity shelf water
IRD	ice rafted debris
IBRD	iceberg rafted debris
IBRD MAR	iceberg rafted debris mass accumulation rate
IODP	Integrated Ocean Drilling Program
K/Ti	a provenance proxy; the ratio of potassium to titanium
LPSA	laser particle size analyzer
LR04	Lisiecki-Raymo 2004 benthic $\delta^{18}\text{O}$ stack
LSR	linear sedimentation rate
MCDW	modified circumpolar deep water
MCO	Mid-Miocene Climatic Optimum
MMCT	Mid-Miocene Climate Transition

MTM	multitaper method
NADW	North Atlantic Deep Water
NaOH	sodium hydroxide
NGR	natural gamma ray
PCO	Pliocene Climatic Optimum
SAF	Subantarctic Front
Si/Al	a biogenic silica proxy; the ratio of silica to aluminum
SST	sea surface temperature
UCDW	upper circumpolar deep water
v%	volume percent
WAIS	West Antarctic Ice Sheet
WSB	Wilkes Subglacial Basin
wt%	weight percent
XRF	x-ray fluorescence
$\delta^{18}\text{O}$	ratio of oxygen-18 to oxygen-16

Units of Measurement

$^{\circ}\text{C}$	degrees Celsius
μm	micron; unit of length, equivalent to one millionth of a metre
‰	parts per thousand
cc	centimeter cubed
cm/kyr	centrimetres per thousand years
cm^3	centimetre cubed
cm s^{-1}	centimetres per second
dm	decimeter; a unit of length; equivalent to one tenth of a metre
g	gram; unit of mass, equivalent to one thousandth of a kilogram
$\text{g/cm}^2/\text{kyr}$	grams per centimetre squared per thousand year
g/cm^3	grams per centimetre squared
km	kilometer; unit of length, equivalent to one thousand metres
kV	kilovolts
kyr	thousands of years
M	molar concentration
m	metre
m/myr	meters per million years
Ma	millions of years before present
mbsf	meters below sea floor
ml	millilitre; unit of volume, equivalent to one cubic centimetre
mm	millimetre; unit of length, equivalent to one millionth of a metre
myr	millions of years
ppm	parts per million
rpm	revolutions per minute

CHAPTER 1: INTRODUCTION

1.1 Thesis context and motivation

The Antarctic Ice Sheets (AIS) are a significant modulator of global climate — how the ice sheets respond to variations in the Earth system have direct implications for global sea level, and global atmospheric and oceanic circulation (Golledge *et al.* 2018). Understanding ice sheet dynamics under a range of climate scenarios provides context for modelling the response of the ice sheets to a warming climate.

The Early Pliocene warm period (~5.3–3.5 Ma) is commonly cited as an analogue for future climate change, with similar geographic boundary conditions to today, atmospheric CO₂ concentrations at ~400 ppm, and a global climate that was 2–3°C warmer than present (Seki *et al.* 2010; Zhang *et al.* 2013). The instability of marine-based sectors of the West Antarctic Ice Sheet (WAIS) during the Late Miocene and Early Pliocene was confirmed by ANDRILL AND-1B drill core in the Ross Sea, which obtained direct sedimentary records of cyclic retreat and advance of the WAIS fluctuating at orbital timescales (McKay *et al.* 2009; Naish *et al.* 2009). This instability of the WAIS was proposed to be the result of shifting latitudinal temperatures resulting in movements of Southern Ocean wind fields that are controlled by the tilt of the Earth's axis (i.e. obliquity), that lead to invigorated upwelling of warm circumpolar deep water (CDW) onto West Antarctica's continental shelves (Naish *et al.* 2009). Once ice sheet retreat was initiated, internal marine ice sheet instability processes relating to the loss of ice shelf buttressing, and the reverse slope nature of the continental shelf in West Antarctica, resulted in the loss of the Ross Ice Shelf and most of the WAIS (Pollard & DeConto 2009).

Geological and modelling studies have since recognised similar retreat patterns from marine-based sectors of the East Antarctic Ice Sheet (EAIS), such as the Wilkes Subglacial Basin (WSB), occurring in the Early Pliocene, with significant retreat inferred from geochemical provenance tracers (Cook *et al.* 2013; Mengel & Levermann 2014; Golledge *et al.* 2017). This response of the WSB to oceanic forcing highlights the importance of understanding the drivers of ice sheet variability in these marine-based sectors, as they are key regions for understanding ice dynamics in a world with warmer-than-present atmospheric and oceanic conditions.

In a natural climate state, paleoclimate records have established a link between incoming solar radiation (insolation) and ice sheet variability to the three major astronomical cycles of the Earth's orbit around the sun — 100 kyr eccentricity, 40 kyr obliquity, and 20 kyr precession (e.g. Hays *et al.* 1976). However, the response of AIS to astronomical forcing has varied significantly through the Late Cenozoic, and at times it has been more sensitive to different astronomical frequencies. This is likely to be the consequence of changing climate states, which acts to modify the processes that influence ice sheet mass-balance (Huybers 2006; Raymo *et al.* 2006). A recent hypothesis by Levy *et al.* (2019), proposes that shifts in orbital pacing of the AIS between 34 and 5 Ma were primarily controlled by their linkage to the ocean. Boundary conditions such as whether the ice sheet was predominantly terrestrial or marine based, or if there was the presence/absence of a significant sea ice belt in the Southern Ocean, controlled ice sheet response (Levy *et al.* 2019).

Throughout most of the Plio–Pleistocene, the benthic foraminiferal $\delta^{18}\text{O}$ stack displays a dominant 40 kyr obliquity response of global ice sheet volumes, although there are short-lived periods of eccentricity/precession forcing (LR04; Lisiecki & Raymo 2005). A Middle to Late Pliocene record of iceberg rafted debris (IBRD) from Site U1361 on the Wilkes Land margin closely resembled the LR04 stack, and IBRD peaks could be visually correlated to mean annual insolation at 40 kyr cyclicity, but documented a fundamental shift towards eccentricity/precession pacing in the Late Pliocene to Early Pleistocene (Patterson *et al.* 2014). This thesis builds on previously published records from the Wilkes Land sector from the Middle Pliocene to Late Pleistocene to assess if the high fidelity orbital signal of the IBRD datasets presented in Patterson *et al.* (2014) was also evident in the latest Miocene and into the Early Pliocene at IODP Site U1361, and if the variances in ice sheet volume is consistent with obliquity forcing of WAIS from the AND-1B core (Naish *et al.* 2009).

1.2 Study location

Site U1361 was drilled by Integrated Ocean Drilling Program (IODP) Expedition 318 during the austral summer of 2010 on the lower continental rise of Wilkes Land, East Antarctica (Figure 3.1; Escutia *et al.* 2011). The site is influenced by sediment derived directly from the continental shelf (turbidites and ice rafted debris), and thus provided a near-

continuous Late Miocene to Pleistocene record that is proximal enough to record a direct signal of AIS dynamics, but also contains a pelagic component that will capture a record of surface water productivity (Escutia *et al.* 2011; Patterson *et al.* 2014). This site is sensitive to ice extent on the continental margin and thus provides a continuous history of AIS variability, without the erosional breaks that characterised the more direct record of ice sheet advance and retreat at AND-1B site (Naish *et al.* 2009).

1.3 Thesis aims and objectives

The overarching aim of this thesis is to examine the response of ice dynamics and marine productivity at the East Antarctic margin during the Late Miocene to Early Pliocene to orbital forcing to test existing hypotheses of orbital drivers for marine-based ice sheet variability in warmer-than-present climates. This will be achieved by assessing the primary controls on sedimentation at the site through facies and grain size analysis, and by inorganic geochemical methods (e.g. x-ray fluorescence (XRF)) and core physical properties to assess biological productivity and oceanography. Analysis of paired samples allows the leads and lags of physical and biological systems to glacial/interglacial cyclicity to be assessed. The key objectives of the thesis are:

1. develop a sedimentary depositional model for the Late Miocene to Early Pliocene (6.2–4.4 Ma) at Site U1361 in order to reconstruct ice dynamics and ocean productivity at this time.
2. identify how the EAIS and physical/biological systems in the adjacent oceanic margin in the Wilkes Land region responded to insolation forcing during Late Miocene to Early Pliocene (6.2–4.4 Ma) climates.

A secondary objective of this thesis is to develop a complete Late Miocene to Early Pleistocene (6.2–2.2 Ma) record of IBRD from U1361, by combining results from this study with previously published work by Patterson *et al.* (2014). This record will assist in understanding some of the longer-term drivers that may have led to shifts in the frequency of orbital pacing of past ice age cycles.

CHAPTER 2: BACKGROUND

2.1 Late Cenozoic climate and ice sheet history

2.1.1 Initiation of Antarctic Ice Sheets

The Eocene–Oligocene Climatic Transition (EOCT; Figure 2.1A) at ~34 Ma marks a major step in Cenozoic climate, with a shift from early Cenozoic ‘greenhouse’ conditions to a glaciated Antarctic climate state (Lear et al. 2008). Far-field benthic foraminifera $\delta^{18}\text{O}$ record the onset of Antarctic glaciation as a two-step 1.5 ‰ increase in deep-marine oxygen isotope values associated with a decrease in atmospheric CO_2 to <750 parts per million (ppm) (Pagani et al. 2005; DeConto et al. 2008). CO_2 decline and the tectonic opening of Southern Ocean ‘gateways’ are proposed to be primary boundary conditions required for triggering continental-scale Antarctic glaciation (Kennett 1977; Bijl et al. 2013; Anagnostou et al. 2016). Once these boundary conditions were in place, an optimally cold orbital configuration at the Eocene–Oligocene boundary, characterised by prolonged low-variability ‘nodes’ in orbital eccentricity and obliquity, acted to trigger the exact timing and step-wise nature of ice sheet development (Zachos et al. 2001). However, the relative importance of different climatic mechanisms and associated feedbacks in driving ice sheet development are complex and remain poorly resolved.

Oxygen isotope records are commonly used as a proxy for ice volume because the isotopic fractionation of ^{18}O and ^{16}O within the oceans is approximately proportional to global ice volume. As ^{16}O is lighter, it is more readily evaporated from the ocean’s surface. When it is precipitated as snow on polar ice caps (i.e. not returned to the ocean as river runoff from land), global oceans are depleted in this isotope. Thus, during glacial periods, the ocean is relatively enriched in ^{18}O compared to the present day and in ^{16}O during warmer-than-present interglacials as polar ice melts. However, there is ambiguity in $\delta^{18}\text{O}$ records as this ratio is also influenced by the temperature of the water the foraminifera grew in. The unreliability of separating the effects of ice volume and temperature in $\delta^{18}\text{O}$ records is a major impediment to its use as a recorder of ice sheet dynamics. A further compounding factor is that it is a globally integrated proxy of ice volume, and even if the effects of temperature on $\delta^{18}\text{O}$ can be accounted for, it is not possible to identify the location of where (i.e. which hemisphere) the ice volume changes are occurring.

In contrast, ice-proximal records in Antarctica provide direct evidence of AIS evolution, although they are geographically limited and often temporally discontinuous. Thus, a combination of near continuous far-field and near-field records is required to advance understanding of AIS development and variability through time.

Figure 2.1 Integrated oxygen isotope and evolutionary power spectra from the Early Oligocene to Early Pliocene. Arrows delineate significant shifts in the $\delta^{18}\text{O}$ record indicating a response in global climate (blue = cooling, red = warming). A–D are events referred to in text. Horizontal arrows show phases of AIS evolution and approximate CO_2 range (black = terrestrial; blue = episodic marine; red = perennial sea ice). Modified from Levy *et al.* (2019).

The widespread continental-scale expansion of the AIS began at the Oi-1 glaciation, as defined by two distinct excursions in $\delta^{18}\text{O}$, with an initial $\sim 0.7\%$ increase at 34 Ma — inferred to be the consequence of a 2.5°C cooling of tropical surface and bottom waters (Lear *et al.* 2008). Such a temperature shift implies most of the initial $\delta^{18}\text{O}$ shift represents a global cooling and only a moderate ($\sim 0.2\%$) portion represented an increase in continental ice volume. The second $\sim 1\%$ shift at ~ 33 Ma is interpreted as the major expansion in AIS ice-volume. At this time the ice sheet grew to approximately modern-day size and was associated with a 40–50 m lowering of eustatic sea level and a $\sim 2^\circ\text{C}$ deep sea cooling (Pekar *et al.* 2002; Katz *et al.* 2008). Glacimarine strata from Cape Roberts Project drill cores indicate a smaller, terrestrial AIS, responding to eccentricity-paced insolation forcing was in place prior to 32.8 Ma (Galeotti *et al.* 2016). The first evidence for a stable, continental scale, marine-calving ice sheet did not occur until ~ 32.8 Ma when atmospheric CO_2 levels dropped below 600 ppm and ocean temperatures were cold enough to sustain ice sheets expanding across the continental margin, about 1 Myr after glacial maxima inferred by $\delta^{18}\text{O}$ records (Galeotti *et al.* 2016).

Following the Oi-1 glaciation, $\delta^{18}\text{O}$ values remained relatively high ($>2.5\text{‰}$) indicating a persistent presence of ice sheets on the continent, albeit highly oscillatory and transient in nature (Zachos *et al.* 2011), until the Early Miocene (Figure 2.1B). The Mi-1 isotopic excursion marks the onset of a period of strong and rapid climate variability. Although CO_2 concentrations remained relatively high (600–400 ppm), Levy *et al.* (2019) proposed that a climatic cooling threshold was reached at this time that allowed marine terminating ice sheets to advance and retreat across the continental shelf. At transitions from glacial to interglacials, the retreat of the AIS is thought to be a result of the propagation of relatively warmer waters onto the Antarctic shelf. This may be in response to changes in the latitudinal heat gradient, which is strongly regulated by obliquity on these timescales.

2.1.2 Mid-Miocene Climatic Optimum and Transition (16.9–13.8 Ma)

The Mid-Miocene Climatic Optimum (MCO; Figure 2.1C) was the warmest interlude in the general Antarctic cooling trend of the Late Cenozoic, with ocean temperatures up to 10°C in the Ross Sea and Wilkes Land margin (Zhang *et al.* 2013; Sangiorgi *et al.* 2018). Drill core records from the Ross Sea (AND-2A) and Wilkes Land (U1356) indicate fundamentally different marine and terrestrial environments in Antarctica during MCO than any time since (Sangiorgi *et al.* 2018). This climatic warming event was characterised by a humid, ice-free coast covered by temperate vegetation, a significantly smaller AIS, and warm, ice-free marine conditions with temperatures similar to those observed north of modern-day southern hemisphere subtropical fronts (Warny *et al.* 2009; Sangiorgi *et al.* 2018). The Mid-Miocene Climatic Transition (MMCT Figure 2.1D; 14.2–13.8 Ma) represents the termination of MCO, and was the beginning of significant Cenozoic climatic cooling associated with a 200–300 ppm decrease in atmospheric CO_2 concentrations to below 300 ppm and sea surface temperatures cooling by $6\text{--}7^\circ\text{C}$. Climatic cooling at the MMCT had a significant impact in Antarctic environments, with sea surface temperatures (SST) and atmospheric temperatures around the continent cooling enough for perennial sea ice to form (Levy *et al.* 2016; Sangiorgi *et al.* 2018).

The climatic cooling across the MMCT has been suggested to have decreased the influence of local insolation on surface melt of the ice sheet, by increasing the insolation melt threshold and thereby reducing the magnitude of terrestrial ice-sheet variability

(Levy *et al.* 2019). Surface melting (solar radiation impinging directly on the surface of the ice) is most likely to dominate in terrestrial ice sheets that are not marine terminating (Huybers 2006). Instead, the expansion of AIS marine-based sectors after MMCT, and tectonic subsidence combined with glacial erosion of the continental shelf, meant the ice sheet was highly sensitive to oceanic influences, especially to excursions of warm water onto the continental shelf (Levy *et al.* 2019). This sensitivity to ocean forcing is combined with the feedbacks of increased oceanic sensitivity due to an increased deepening and expansion of the Antarctic continental margin since the Middle Miocene (Colleoni *et al.* 2018). Ocean-ice sheet feedbacks are the primary influence of the dynamical state of the AIS after MCO, with atmospheric forcing a modulating factor of the magnitude of AIS response. A reduction in sediment supply to sedimentary records from the East Antarctic continental rise (Escutia *et al.* 2005; Gulick *et al.* 2017) and in the Ross Sea (Passchier 2011; Kim *et al.* 2018) are thought to represent a transition from a highly-erosive, wet-based to a less-erosive, cold-based 'permanent' East Antarctic Ice Sheet.

2.1.3 Late Miocene cooling and Early Pliocene warmth

Following MCO, the Late Miocene saw a progressive long-term $\sim 6^{\circ}\text{C}$ cooling trend of SST in the high latitudes of both hemispheres to approximately near-modern values between 7 and 5.4 million years ago (Herbert *et al.* 2016). This cooling coincides with records of IBRD pulses in the Northern Atlantic Ocean interpreted as the onset of ephemeral Greenland glaciation (Thiede *et al.* 1998; Pérez *et al.* 2018). Low spectral power at obliquity frequencies within the $\delta^{18}\text{O}$ record (Figure 2.1) during the Late Miocene are hypothesised to reflect times when sea ice extent increased, such that it acted as a barrier to warm water incursions onto the continental shelf, and thereby reduced the oceanic influence on AIS (Levy *et al.* 2019). Increase in obliquity sensitivity at ~ 5.8 Ma may be related to interglacials warm enough to melt sea ice, but cold enough to maintain an obliquity-controlled marine ice sheet.

The Early Pliocene warm period was a ~ 2.2 million year reversal of the general Late Cenozoic cooling trend, with prolonged warmer-than-present atmospheric and marine conditions. Atmospheric CO_2 concentrations ranged between 300 and 415 ppm by ~ 4.5 Ma (Figure 2.2; Seki *et al.* 2010; Zhang *et al.* 2013) and global mean temperatures were

2–3°C higher than pre-industrial levels (Haywood & Valdes 2004). At ~4.6 Ma, SSTs in the Southern Ocean were >4°C above modern-day temperatures (Whitehead & Bohaty 2003), up to ~5°C in the Ross Sea (McKay *et al.* 2012), and were associated with a >55% reduction in sea ice extent in Prydz Bay (Whitehead *et al.* 2005).

Figure 2.2 Plio–Pleistocene integrated CO₂ (top) and Lisiecki & Raymo (2005) benthic δ¹⁸O (bottom) records. The blue arrow outlines the increase in δ¹⁸O associated with global Late Pliocene cooling and intensification of Northern Hemisphere ice sheets. Yellow: Zhang *et al.* (2013); red: Seki *et al.* (2010).

Several ice-proximal Antarctic sites record a dynamic response of ice sheet behaviour and commensurate biological activity in response to higher-than-present atmospheric and SSTs (Bohaty & Harwood 1998; Billups *et al.* 2013; Cook *et al.* 2013; Reinhardy *et al.* 2015; Hansen & Passchier 2017). Sedimentary facies from drill cores in the Ross Sea (AND-1B) and Prydz Bay (ODP Site 1164) both indicate warmer-than-present marine-based sectors of the AIS in the Early Pliocene. SSTs were sufficiently warm enough to restrict the development of prolonged ice-shelf conditions during glacial minima (Quilty *et al.* 2000; Naish *et al.* 2009; McKay *et al.* 2012). ANDRILL provided the first evidence of periodic collapse of the Ross Ice Sheet during the Early to Middle Pliocene, with regular obliquity paced advances and retreats of the Ross Ice Shelf between ~4.6 and 4.8 Ma. This interval had a low abundance of sea ice associated diatoms during interglacials, suggesting summer SSTs were above freezing (Winter *et al.* 2010; McKay *et al.* 2012). The appearance of silicoflagellates (*Dictyocha* genus) that predominantly reside north of the Antarctic Polar Front (APF) at Prydz Bay occurred in discrete intervals in the Early to Middle Pliocene (Whitehead & Bohaty 2003) and coincides with a 90 m thick section of diatomite in the

Ross Sea containing subantarctic diatom flora (Naish *et al.* 2009). The influx of these subantarctic species suggests warming of surface water masses and sea ice conditions around the continent during the Early to Middle Pliocene. A possible mechanism for the influx of warmer waters is the poleward migration of the Southern Hemisphere westerly wind belt (see section 3.3.1), increasing Southern Ocean temperatures via the increased influence of subtropical gyres (McKay *et al.* 2012). This process has been modelled to occur in the Middle Pliocene warm period (3.3–3 Ma; Li *et al.* 2015).

The seismic stratigraphic geometry of the Wilkes Land continental shelf indicates the EAIS expanded towards the continental shelf edge during glacial maxima in the Pliocene. Therefore, it is thought most AIS variability was a consequence of the growth and retreat of the marine-based sectors of the ice sheets which are highly susceptible to oceanic change (Escutia *et al.* 2005). IBRD records from Prydz Bay suggests both local and distant EAIS retreat during the Early Pliocene, with significant Adélie Land-sourced IBRD (WSB) at 4.8 Ma and Wilkes Land-sourced Aurora Subglacial Basin (ASB) IBRD at 4.6 Ma, coinciding with retreat in the Ross Embayment and local Prydz Embayment (Hambrey & McKelvey 2000; Naish *et al.* 2009; Williams *et al.* 2010). These peaks in IBRD up to 3000 km from source during very warm interglacials suggest large, rapid destabilisation events produced massive iceberg fluxes from marine-based EAIS sectors such as WSB and ASB (Cook *et al.* 2014). Similar EAIS retreats were recognised from the Wilkes Subglacial Basin using sediment provenance proxies that indicate several hundred kilometres of retreat of the ice margin (Cook *et al.* 2013). These retreats coincided with periods of warm, highly productive surface waters and Ross Sea derived IBRD at Wilkes Land suggesting a large iceberg flux from the WAIS (Cook *et al.* 2017).

A Wilkes Land IBRD record from the Middle to Late Pliocene recorded strong 40 kyr obliquity pacing of the EAIS during peak Pliocene warmth between 4 and 3.5 Ma. This may be a response to obliquity controlled insolation in higher atmospheric CO₂ conditions, which acts to increase the duration and intensity of austral summer (Patterson *et al.* 2014). An abrupt switch to eccentricity and precession pacing of IBRD at ~3.5 Ma, and a change in the sedimentation patterns at the site suggests a profound change in ice sheet dynamics. This was hypothesised to be related to decreasing atmospheric CO₂ concentrations decreasing the length of summer, and an expansion of sea ice cover as

regional Southern Ocean SSTs cooled (McKay *et al.* 2012), and was characterised on the Wilkes Land margin as a major decrease in IBRD after 2.5 Ma (Patterson *et al.* 2014). Regular expansions and retreats of the marine-based WAIS into the Ross Sea occurred from ~3.3 Ma, coinciding with an intensification of sea ice conditions as a result of SST cooling between 3.3 and 2.5 Ma (McKay *et al.* 2012).

The benthic foraminiferal $\delta^{18}\text{O}$ LR04 stack also reveals a switch from obliquity to eccentricity/precession pacing at 3.5 Ma, until a reemergence of obliquity at ~3 Ma. This is in contrast to the Wilkes Land IBRD record from Patterson *et al.* (2014), and may be a reflection of the increased influence as Northern Hemisphere ice sheets on the $\delta^{18}\text{O}$ record as they began to oscillate at obliquity-paced periods (Lisiecki & Raymo 2005). IBRD studies support a ~2.57 Ma age for the onset of major Northern Hemisphere glaciation in the northern Pacific Ocean (Rea & Schrader 1985), although minor IBRD has been found in the Norwegian Sea as early as 3.3 Ma, suggesting localized ice caps may have been present earlier (Jansen *et al.* 2000). In the cooler Late Pleistocene, the pacing of climatic variability moved from an obliquity-paced 40 kyr world to large amplitude 100 kyr eccentricity cycles (Figure 2.2), with large variability in ice sheet extent between glacial-interglacial periods in both hemispheres. Recent studies have indicated that despite relatively cool temperatures through the Pleistocene, a significant retreat occurred in the WSB during the Last Interglacial Period (120 ka BP), when global atmospheric temperatures were ~1–2°C warmer than present, and along with WAIS and Greenland retreat may have contributed to sea level up to 6–9 m higher than present (Wilson *et al.* 2018). This suggests marine-based sectors of the Antarctic Ice Sheets, including the Wilkes Land EAIS, will remain sensitive to even modest warmings in future climate change.

2.2 Internal and external drivers of climate change

2.2.1 Southern Ocean and the carbon cycle

Atmospheric CO_2 exchanges rapidly with oceans and terrestrial ecosystems and the rate at which this exchange occurs determines atmospheric concentrations. The global oceans are a much larger carbon reservoir than the atmosphere, and therefore the oceans, particularly the Southern Ocean, are important in regulating long-term carbon and climate systems (Broecker & Peng 1982; Falkowski *et al.* 2000; Sigman *et al.* 2010). During

interglacials, it is theorised that the poleward movement of westerlies invigorates the Antarctic Circumpolar Current (ACC), which increases upwelling of carbon-rich water to the surface where it is available for exchange to the atmosphere (Toggweiler *et al.* 2006). This process may be reduced during glacials due to enhanced stratification limiting the exposure of deep carbon-rich waters to the atmosphere (Köhler *et al.* 2005), large changes in vertical ocean mixing (Toggweiler 1999), and/or sea ice limiting gas exchange rates (Stephens & Keeling 2000).

CO₂ is more soluble in cold and saline waters. Therefore, the ability to sequester atmospheric CO₂ is controlled by the formation of these water masses at high latitudes (the 'solubility pump'; Falkowski *et al.* 2000), especially in the North Atlantic and Southern Ocean (Broecker & Peng 1992; Stocker & Schmittner 1997). The efficiency of the solubility pump depends on latitudinal and seasonal changes in ocean ventilation and strength of the overturning circulation. Cold, dense water masses sink to depths below which dissolved inorganic carbon (DIC) within the water can no longer interact with the atmosphere. The DIC is then stored within the deep ocean until the waters are upwelled to the surface, potentially centuries later. Increased stratification in the oceans reduces the ability of surface waters to sink and as such decreases the exchange of carbon to the deep ocean (Köhler *et al.* 2005).

Another critical aspect relating to changes in upwelling of nutrients in the Southern Ocean is the sequestering of DIC to the ocean occurring as a result of the biological pump, due to photosynthesising phytoplankton in the ocean's surface waters. Photosynthesis by phytoplankton reduces the partial pressure of surface water, which then sequesters CO₂ from the atmosphere to restore equilibrium (Falkowski *et al.* 2000). Upon death, sinking of the remains of these plankton acts to transfer both organic and inorganic carbon to the ocean reservoir and underlying sediments. This process is enhanced by the formation of particulate organic carbon in the top 200 m which is processed by biological communities into organic aggregates which sink to depth (Turner 2015; Basu & Mackey 2018).

2.2.2 Orbital forcing

2.2.2.1 Astronomical variability

The variability of the Earth's axis of rotation and orbit around the sun results in the amount and distribution of solar radiation across Earth's surface varying over time. The prevailing theory of the glacial–interglacial cycles, or 'Milankovitch theory', is that orbitally driven variability in insolation serves as a pacemaker in forcing ice sheet variance, specifically, the astronomical cycles of precession (19 and 23 kyr periods), obliquity (41 kyr), and eccentricity (95 and 120 kyr) (Figure 2.3; Milankovitch 1941). The identification and quantification of Milankovitch orbital frequencies were first recognised by Hays *et al.* (1976) and has since been recognised as a pervasive signal in paleoclimate.

For a given latitude, solar radiation received at the top of the Earth's atmosphere depends on the Earth's orbit and rotational parameters, which are a function of the gravitational effects of the sun, the moon, and the planets (Berger 1978). Eccentricity is the only cycle to vary total insolation over time as it changes the Earth's distance from the sun and orbital speed; both obliquity and precession control the latitudinal distribution of insolation at the Earth's surface.

Figure 2.3 The three major astronomical orbits responsible for climate variability. Eccentricity (95 and 120 kyr, averaging 100 kyr), obliquity (40 kyr), and precession (19 and 23 kyr).

Precession varies at a frequency of 19 kyr and 23 kyr, and is the 'wobble' of the Earth's axis. Precession changes the direction of the Earth's tilt towards the sun and controls the

timing of seasons (equinoxes), relative to when Earth is at perihelion (closer to the sun), or aphelion (further from the sun). Precession controls the intensity and duration of the seasons, or mean summer insolation, and is out-of-phase between hemispheres — i.e. when a northern hemisphere summer occurs at perihelion, the summer will be warmer than the southern hemisphere summer due to the distance from the sun, and vice versa.

Obliquity is the tilt of the Earth's rotational axis and oscillates between 22.1° and 24.5° at a dominant frequency of 41 kyr. Obliquity is the main control on mean annual insolation as it controls the latitudinal gradient of insolation and thus affects seasonal contrast. An increased tilt increases the amount of heat at the poles and increases the amplitude of the seasonal cycle. A reduced tilt distributes insolation more evenly between latitudes, resulting in less heat at the poles and milder seasonality. Obliquity is in-phase in both hemispheres and has a maximum effect on the variability of insolation at high latitudes.

Eccentricity changes the shape of the Earth's orbit around the sun and is a combination of 95 kyr, 120 kyr, and 400 kyr periodicities. Eccentricity controls the relative distance between the Earth and the sun during perihelion and aphelion, and as such acts to modulate the amplitude of the precession cycle. Variation in the eccentricity cycle is the only orbit to affect the total solar insolation; following Kepler's second law of planetary motion, a planet sweeps out equal areas in equal times, thus the speed of a planet increases as it nears the sun and decreases as it recedes. So, during periods of high eccentricity (i.e. more elliptical), the Earth takes longer to transit the half of its orbit where it is further from the sun than it does where it is closer to the sun. The duration of the summer is inversely proportional to the Earth's distance from the sun, and total solar insolation averaged over a year is less when the orbit is more elliptical.

2.2.2.2 Milankovitch theory and paleoclimate records

A relationship between climate and insolation was observed by Emiliani (1966) who found a strong correlation between oxygen isotope-based temperature estimates and insolation minima at 65°N . This relationship was quantitatively assessed by Hays *et al.* (1976), who revealed the independent contribution of the three orbital elements within paleoclimate time series. The prevailing theory of Milankovitch climate cycles is that major oscillations of the global climate are caused by insolation variations at critical latitudes and seasons.

In particular, Milankovitch theorised that the advance and retreat of polar ice sheets were in response of climate sensitivity to Northern Hemisphere high latitude summer insolation, as controlled by precession — if snow and ice survive the boreal summer, the ice sheet will grow, and with enhanced snow-albedo feedbacks will cause full ice sheet development (Milankovitch 1941). This theory of insolation control on glacial—interglacial cycles is well established, however, the mechanics of how the Earth system translates insolation forcing to drivers of ice sheet variability is still under debate. Theoretical models of insolation demonstrate eccentricity modulated precession has the greatest influence on high-latitude insolation intensity (Figure 2.4a; Milankovitch 1941; Laskar et al. 2004), however in many climate records in the Plio–Pleistocene (e.g. LR04 $\delta^{18}\text{O}$ stack, Figure 2.4b), a strong, persistent precession signal is missing, and is instead dominated by strong obliquity with a subordinate precession/eccentricity influence.

The pervasive obliquity signal within paleoclimate records has been explained through the variation in summer insolation. One hypothesis is that ice sheets are sensitive to insolation integrated over the duration of the summer, that is ice sheets are sensitive to the duration of the summer melt season (defined by the number of positive degree days). In warmer climate states (e.g. higher CO_2), the duration of summer increases, which skews the integrated summer insolation towards an obliquity signal (Huybers 2006). Alternatively, it is proposed changes in Antarctic ice volume are driven by the response to summer intensity, which could counter-balance the precession component of Northern Hemisphere ice volume changes, as precession is out-of-phase between hemispheres (Raymo *et al.* 2006). This effectively removes any strong precession signal from globally integrated measures of ice volume, leaving the obliquity component of the signal to dominate. However, both of these theories are theoretical models focused on the top-down melting control on ice sheet behaviour and are difficult to resolve in an Antarctic context. Raymo *et al.* (2006) provided a conceptual model for the period between 1 to 4 Ma but this does not account for the lack of strong precession in the unipolar world prior to the onset of major Northern Hemisphere glaciation at 3 Ma. However, Huybers (2006) requires an increase in the surface melt threshold, such that the melt season becomes very short and only occurs during peak insolation intensity. Both models oversimplify processes and feedbacks relating to AIS variability, which is largely driven by oceanic

processes, whereby marine-based sectors of the ice sheet lose mass through calving and basal melt of ice sheets (Pollard & DeConto 2003; Rignot *et al.* 2013).

These hypotheses were refined by Patterson *et al.* (2014), who examined the orbital pacing of EAIS variations using an IBRD record from Wilkes Land (U1361) from the Middle to Late Pliocene. Between 4.2 and 3.5 Ma, obliquity-forced 40 kyr cycles dominated the record. They theorised that under a warmer climate state, high CO₂ concentrations required less insolation to melt sea ice, thus extending the austral summer and the climate signal was more strongly influenced by mean annual insolation which is controlled by obliquity. Sea ice is an important buffer for oceanic induced melting of marine based ice sheets, and thus the duration of the sea ice season can provide an important control on the mass balance of the marine-based sectors of the WAIS and EAIS. In the Late Pliocene, IBRD switched to 23 kyr precession pacing, explained by the cooling climate raising the melt threshold such that the duration of the sea ice melt season was restricted to a shorter period (i.e. an overall decrease in seasonal sea ice melt duration), and thus the signal is dominated peak summer insolation controlled by precession. Consequently, this increase in sea ice acted to restrict the inflow of warmer water masses onto the continental shelf.

Figure 2.4 Power spectra of ETP (6.2–4.4 Ma) and LR04 (5.3–0 Ma). a. ETP displays strong 23 and 19 kyr cycles of precession, with minor obliquity and eccentricity (Laskar *et al.* 2014). b. LR04 displays strong 40 kyr cycles of obliquity with only minor eccentricity and precession (Lisiecki & Raymo 2005).

Prior to the Plio–Pleistocene, the $\delta^{18}\text{O}$ (and by extension the AIS) appear to be sensitive to different astronomical frequencies at different times (Figure 2.1). A recent quantitative examination of the composite $\delta^{18}\text{O}$ record from the Late Eocene to Early Pliocene shows a variable response to obliquity forcing, with long periods of eccentricity forcing with less of an influence of obliquity. This is theorised to be the result of the obliquity-dominated

signal being a response of the ice sheets to marine influence, and eccentricity/precession only occurs when marine influence is reduced (Vleeschouwer *et al.* 2017; Levy *et al.* 2019). Enhanced obliquity-forcing on the ice sheets is thought to occur through changes in the meridional SST gradient inducing changes in the strength of the Southern Hemisphere westerlies. The poleward expansion of the westerly wind field over the Antarctic circumpolar convergence zone has been associated with a reduced sea ice field (McKay *et al.* 2012), and the upwelling of warm, CO₂ rich circumpolar deep water onto the continental shelf (Toggweiler *et al.* 2006), with consequences for the stability of marine ice sheet grounding lines (Pritchard *et al.* 2012). The dominance of precession-paced variability and the corresponding reduction in obliquity influence is interpreted to reflect a declining influence of oceanic forcing on EAIS stability and extent as the globe cooled and enhanced sea ice restricted warm water incursions sourced from the lower latitudes (Patterson *et al.* 2014).

2.3 Depositional models for glaciomarine margins

In the high latitudes, processes relating to ice sheet advance and retreat on the continental shelf will have a direct influence on sediment deposition in the more distal continental rise (Lucchi *et al.* 2002; Cowan *et al.* 2008; Passchier *et al.* 2018). The primary source of sediment and the processes of deposition in glaciomarine settings reflect variations in the proximity of the grounding line (availability of terrigenous glaciogenic sediments) and oceanic processes such as the presence of sea ice or biogenic productivity (biosiliceous sediments in particular). Continental shelf sedimentary records are directly influenced by glacial processes, but are often compromised by erosional hiatuses during glacial advance. Continental rise records from the high latitudes are inevitably influenced by down-slope and/or oceanic processes, but the stratigraphic sequence of facies is controlled by the long-term modulation of glaciogenic influences. Examining sedimentary facies from continental rise sites are often more continuous records than continental shelf sites, and thus allow for the interpretation of the glacial and oceanographic processes through time.

The main depositional processes in deep water continental margin settings are the deposition of fine-grained sediment and occur as a continuum between horizontal,

current driven (contourite), down-slope gravity driven (turbidite), or slow vertical settling (pelagite) of sediment (Stow & Piper 1984; Figure 2.5). Sedimentation from iceberg rafting can be superimposed onto these processes in glaciated marine settings.

Figure 2.5 Schematic diagram of the continuum of depositional processes and associated sedimentary deposits in the deep ocean. From Rebesco *et al.* (2014).

2.3.1 Pelagic and hemipelagic settling

Pelagic and hemipelagic sedimentation is widespread throughout the world's oceans and represents 'background' sedimentation by slow settling of particles through the water column (Stow & Piper 1984). Hemipelagites comprise of a mixture of biogenic and terrigenous material, and occur in low-energy regions that are still proximal to a source of terrigenous material; whereas true pelagic sediments comprise entirely of biogenic material. Sediment accumulation is slow ($\sim$ cm/kyr) and as a result is often intensely bioturbated, except in anoxic basins devoid of benthic fauna (Stow & Piper 1984). Compositional criteria are based on the relative proportion of terrigenous versus biogenic components (Berger 1974), while textural features are defined by Stow & Piper (1984) as:

1. low sedimentation and continuous bioturbation (except in anoxic basins)
2. no primary sedimentary structures

3. more or less uniform composition within any one succession
4. a variable biogenic component with a fine-grained, often far-travelled, terrigenous component.

Texturally, hemipelagites are generally poorly sorted silty clays, with 1–15% sand, and may contain irregularities such as authigenic minerals, aeolian sediment, or ice-rafted debris (Stow & Piper 1984).

2.3.2 Turbidity currents

Turbidity currents are episodic, high- to low-energy fluid density flows that initiate on the continental rise and may transport coarse-grained sediment into the deep ocean. Turbidite deposits are widespread in the deep ocean and occur as very thin (mm) to very thick beds (m scale) that were deposited rapidly from a single resedimentation event, and may be superimposed or interact with both background sedimentation and along-slope processes (Stow & Piper 1984). The 'classic' sandy turbidite sequence was described by Bouma (1962) and contains massive, medium-to-coarse sand fining up to 'featureless' mud (Figure 2.6a). However, the classic Bouma sequence is rarely preserved in the geological record, and further work recognised a continuum of turbidity flows related to energy, sediment supply, and distance from the continental margin (Figure 2.7). In distal environments, silt beds (>70% silt) are more abundant than sand beds and commonly occur as thin mm-scale laminae to thick cm-scale bedded turbidites (Figure 2.6b). Fine-grained turbidites commonly exhibit the same suite of structures as the classic sandy turbidite, but the complete sequence is also rarely found and more commonly occur as incomplete base-cut-out sequences. Stow & Piper (1984) note that the chief criteria for distinguishing turbidites from other similar facies are:

1. a regular sequence with positive grading
2. presence of sedimentary structures indicating rapid deposition, with bioturbation restricted to the top of the beds
3. compositional, textural, or other features which indicate that they are exotic to their depositional environment.

Figure 2.6 Schematic diagram of a classic Bouma a. sand turbidite b. and mud turbidite. Note the scale changes between the two stratigraphic columns. From Stow & Piper (1984).

Figure 2.7 Schematic distribution of fine-grained sediment facies across a typical slope and rise. Note the irregular distribution of mixed facies types. From Stow & Piper (1984).

2.3.3 Contourite currents

Contouritic deposits were originally defined as sediments deposited in the deep sea by contour-parallel thermohaline currents (Heezen *et al.* 1966), where sedimentation is related to the movement of along-slope bottom water flows. Contourite-forming currents are relatively permanent flows, however mean velocities ($<10 \text{ cm s}^{-1}$) are typically less than that is required to rework coarse sediment, and only over steep slopes or constricted flows do velocities reach the $\sim 20 \text{ cm s}^{-1}$ that is required to rework fine sand (Shanmugam 2008; Stow *et al.* 2008). Where higher velocities occur, large-scale deposition and erosional features may result (Stow *et al.* 2002; Rebesco *et al.* 2014).

Contourites are widely recognised in seismic sections which capture broad-scale spatial features, but there is a lack of commonly accepted drill-core-scale criteria for uniquely identifying contourites in sedimentary sequence. In particular, it is especially difficult to distinguish between contourites and turbidites, and deposits with a mixed component of both (Rebesco *et al.* 2014). Proposed criteria by Stow & Piper (1984) include:

1. an irregular vertical arrangement of facies types and structure with both negative and positive grading, but no regular structural sequence
2. evidence for more-or-less continuous bioturbation that has kept pace with deposition
3. composition, texture, or other features that indicate a combined in situ and exotic origin.

Contourite drift sediments are usually continuously deposited and thus have higher accumulation than adjacent pelagic sedimentation. As a result, they may record variability of ocean circulation patterns and current velocities with seasonal to millennia variability related to changes in climatic and oceanic variability (Shanmugam 2008; Stow *et al.* 2008; Rebesco *et al.* 2014). The classic contourite facies are related to the strength of the bottom current that formed them. Negatively graded sedimentation occurs as current velocity increases and positive grading occurs as velocity decreases towards the background sedimentation state (Figure 2.8; Stow *et al.* 2008). Contourites are usually highly bioturbated, which destroys primary sedimentary structures, but there may be indistinct and discontinuous parallel laminae and lenses of coarser material. Pervasive

burrowing may be used as a distinguishing feature between contourites and turbidites, as contour currents are more continuously deposited compared to turbidity currents which are episodic and ‘instantaneous’ in deposition, allowing no time for benthic fauna to burrow (Stow *et al.* 2002). The continuous bed-load transport of contour currents classically result in better sorting of sandy contourites relative to sandy turbidites (Shanmugam 2017).

Figure 2.8 Schematic diagram of the stacking pattern of sediment deposited by contourite currents due to the variation in current speed. From Stow *et al.* (1998).

2.3.4 Glaciomarine influences

In high latitude glaciomarine settings, delivery of terrigenous sediment to more distal continental regions are often highly influenced by the presence of ice sheets which may introduce idiosyncrasies to the classic sedimentary models of continental margins. Low-density turbid surface plumes related to meltwater released from beneath ice sheets can deposit fine-grained ‘plumites’ (Hesse *et al.* 1997). These deposits were first identified along the Labrador slope in Greenland, but occur only within tens of kilometres of the ice margin as 1–2 cm thick interbedded coarse silt and silty clay. They are distinguishable from background hemipelagic sedimentation as they appear as diffuse banding from high sedimentation rates and have low bioturbation. Similar deposits were identified in the Antarctic Peninsula as mm-thick faintly laminated silty layers with sharp, non-erosive bases (Lucchi *et al.* 2002).

In glaciogenic settings, contourites often appear as laminated, unfossiliferous, non-bioturbated muds in regions with low primary productivity and oxygen-poor waters (Lucchi & Rebesco 2007). An important exception to this is areas where polynyas were maintained during glacials, such as the Wilkes Land margin, where contourites appear as bioturbated as a consequence of the Mertz Polynya producing high-salinity shelf water which transports oxygen and nutrient-rich deep water to the continental rise, promoting productivity and bioturbation (Bindoff *et al.* 2000; Caburlotto *et al.* 2010).

Detailed sedimentary studies have identified significant differences in sedimentary facies between interglacial and glacial periods in many regions of Antarctica, depending on local ice sheet and oceanic processes (Hambrey & McKelvey 2000; Hillenbrand *et al.* 2009; Patterson *et al.* 2014). Studying contourite drift deposits on the flanks of continental rise channels in the Antarctic Peninsula region, Lucchi *et al.* (2002) identified four distinct sedimentary facies in a levee setting relating to glacial advance and retreat during glacial–interglacial periods:

Table 2.1 Continental rise lithofacies characteristics and glaciomarine sedimentary processes related to climate state. After Lucchi *et al.* (2002).

Interglacials	bioturbated mud with abundant IBRD and limited evidence for down-slope and along-slope input (e.g. reduced turbidity and bottom currents).
Deglaciations	terrigenous, structureless mud with distinct layers of IBRD caused by the disintegration of a grounded ice sheet. Absence of calcareous microfossils related to strong carbonate dissolution and rise of carbonate compensation depth as a result of major meltwater discharge.
Glacials	unfossiliferous, laminated silty layers with laminae and lenses, high-energy turbidites containing sharp-based and well-defined laminations, no bioturbation, and no IBRD.
Glaciations	terrigenous, structureless mud with rare or sparse IBRD, and biological activity reduced to polynyas.

2.3.5 Iceberg rafted debris on glaciomarine margins

Iceberg rafted debris is coarse (e.g. sand or gravel sized), terrigenous material delivered to the ocean via the transport of icebergs. Material entrained in icebergs is consequently dropped to the sea floor as the iceberg melts, and results in coarse-grained material deposited within finer-grained (e.g. clay to fine sand), regular background sedimentation.

In this study, IBRD is differentiated from ice rafted debris (IRD) which can also refer to sea ice rafted debris.

Terrigenous material transported by icebergs, or ‘dropstones’, were first recognised in the Precambrian rock record by the observation of and comparison to coarse material deposited by icebergs in modern glacier settings. From these observations, it was concluded IBRD could be reliably used to indicate the presence of glacial ice in the geological record (e.g. Owenshine 1970). More recent records from the Northern Atlantic Ocean contained concentrated layers of IBRD (‘Heinrich layers’). These were interpreted as cyclic discharges of icebergs during glacial expansion and retreat of the Laurentide ice sheet during the Late Pleistocene (Heinrich 1988). Observations of these anomalous layers spurred subsequent research to understand the link between millennial-scale climate variability to ice sheet dynamics and ocean–atmospheric interactions (Broecker 1994; Bond & Lotti 1995; Raymo *et al.* 1998; Hemming 2004). IBRD records have since been interpreted to represent the flux of melting icebergs calving from marine-terminating ice sheets in both the Northern Hemisphere and Antarctic ice sheets, and reflect the variations in ice sheet stability and extent through time (Jansen *et al.* 2000; Kanfoush *et al.* 2000).

The advantages of using IBRD records from ice-proximal marine records is that they represent a near-continuous signal of ice discharge and variability. It is estimated the AIS currently loses 50–80% of mass to iceberg calving (Depoorter *et al.* 2013), thus IBRD flux can be used to infer iceberg flux and be interpreted to reflect the destabilisation, surge, and breakup of ice streams (Williams *et al.* 2010). IBRD records from the Antarctic margin have revealed the timing of major climatic change throughout the Miocene, Pliocene, and Pleistocene (Hillenbrand *et al.* 2009; Williams *et al.* 2010; Patterson *et al.* 2014; Hansen *et al.* 2015; Wilson *et al.* 2018). Moreover, mixed-proxy studies highlight the strength of IBRD records when combined with other records to resolve the response of oceanography or biological productivity to climate (or ice sheet) variability (Bertram *et al.* 2018; Sangiorgi *et al.* 2018; Wilson *et al.* 2018). Provenance studies of IBRD have provided insight into iceberg source areas, which on millennial timescales may reflect significant erosion and retreat of individual sectors (Licht *et al.* 2005; Williams *et al.* 2010; Cook *et al.* 2013; Gulick

et al. 2017; Bertram *et al.* 2018) and reconstruct iceberg trajectories through time (Williams *et al.* 2010).

The ability of icebergs to deposit sediment is linked to both the glaciological regime of the ice sheet and to the oceanography of the offshore marine environment. Beneath grounded ice sheets, the bulk of the debris is limited to, and transported in, a basal debris-rich zone that typically drops out as the ice moves through the grounding line. In some regions (e.g. regions lacking large ice shelves or ice tongues), basal debris can be entrained within icebergs alongside englacial and supraglacial debris (Drewry & Cooper 1981; Dowdeswell & Murray 1990). Glacial erosion and entrainment of basal debris occur in the marginal area of a glacier or ice stream where driving stress increases to overcome the strength of the subglacial bed. The lithology of the sediment eroded by ice streams, as well as the distance the clast has travelled via glacial transport, influence the grain size distribution entrained in ice (Hemming 2004).

The preservation and transport of an iceberg relies on sea surface temperatures to be cold enough that icebergs survive transport to the deep ocean (Hemming 2004), although icebergs travel greater distances in ‘warmer’ waters if transported by fast-moving ocean currents (Powell & Domack 2002). Icebergs derived from large ice shelves travel larger distances but have low debris concentrations, whereas outlet glaciers contain higher concentrations of debris and calve debris-rich icebergs regularly, but the icebergs tend to be smaller and do not travel significant distances from source (Powell 1984; Dowdeswell & Murray 1990; Ó Cofaigh *et al.* 2001; Stuart & Long 2011). Hence ice-proximal continental margin settings typically capture local ice sheet dynamics, but IBRD accumulating at sites further to the north (e.g. north of the southern Boundary Front of the ACC) is likely responding to a SST signal (Cook *et al.* 2014).

To interpret IBRD as a proxy for ice dynamics, the record must be considered in the full stratigraphic context as IBRD-bearing icebergs can be produced during glacial maxima (Marshall & Koutnik 2006), glacial advance, glacial retreat, sub-millennial scale events (e.g. Heinrich events in the Northern Hemisphere), and orbital-scale ice sheet dynamics. However, several Antarctic margin records show a flux of IBRD coinciding with bioturbated, diatom-rich sediments (Passchier 2011; Patterson *et al.* 2014; Weber *et al.*

2014), at times of reduced sea ice (Whitehead *et al.* 2005), and under warming conditions at glacial terminations (Cook *et al.* 2014; Wilson *et al.* 2018). As such, increased IBRD accumulations at the drill site may be interpreted as reflecting instability at the ice margin.

2.3.5.1 Iceberg rafted debris methods

Using coarse-grained sediment as a proxy for ice dynamics requires the assignment of an appropriate grain size distribution to accurately estimate sediment delivery by icebergs, to ensure the fraction can be differentiated from other modes of transport for coarse-grained material such as sea ice or aeolian delivery. Modern studies of sediment transported within Arctic sea ice estimate the amount of sand ($>63\ \mu\text{m}$) is less than 5–10% and is potentially overestimated due to sampling bias (Polyak *et al.* 2010). Wind-blown sediment accumulation generally fines with distance from the coast, with sand not travelling more than 5 km beyond source areas, and the average grain size beyond this is much finer than sand (Chewings *et al.* 2014). Ultimately, icebergs are capable of carrying sediment of all grain sizes, so inevitably choosing a cutoff point and removing the fine fraction for IBRD may reduce the actual signal of iceberg activity, however, is necessary to choose larger sizes to remove the potential of capturing the signal of other sedimentary processes.

Several different grain size fractions have been used to calculate IBRD from sediment cores in both the Arctic and Antarctic, including $>125\ \mu\text{m}$, $250\ \mu\text{m}$ – $2\ \text{mm}$, $>250\ \mu\text{m}$, and $>2\ \text{mm}$. Methods for processing IBRD include calculating the mass accumulation rate using the weight percent ($>250\ \mu\text{m}$; Krissek 1995; Patterson *et al.* 2014), or volume percent of IBRD ($>125\ \mu\text{m}$; Hansen *et al.* 2015) and visually counting $>2\ \text{mm}$ clasts from x-radiographs of the sediment core (Grobe 1987; Weber *et al.* 2014). All methods have advantages and disadvantages. The visual $>2\ \text{mm}$ counting requires a higher amount of IBRD material to get a significant result, which may result in underestimation of ice discharge, and therefore only be reliable under intense iceberg activity (Grobe 1987). Weight and volume percent methods only process discrete samples and relies on interpolation between samples so may underestimate or overestimate sections of the core, depending on the sampling resolution. Albot (2016) compared these three methods to test the reliability and comparability between methods. They found the volume $>125\ \mu\text{m}$ estimation, as

used by Passchier (2011) and Hansen *et al.* (2015), overestimated the IBRD component of the sample and was in lower agreement with the other two methods. This may be a result of the amount of sample used (<0.1 g vs 10-20cc), whereby low sand-percent values can be easily biased by sampling as a single grain can represent a large volume when sample sizes are small. This thesis will use the 250 μm –2 mm fraction to represent IBRD.

2.3.6 Inorganic geochemical proxies using X-Ray Fluorescence in Antarctica

2.3.6.1 *Paleoproductivity: biogenic barium and silica*

Barium (Ba) is mainly present in marine sediments as the detrital form barite (BaSO_4) and is considered as a proxy for paleoproductivity due to the close association of barite production and organic matter decay (Dymond *et al.* 1992; Francois *et al.* 1995). The exact mechanism of precipitation of Ba as BaSO_4 is poorly understood, but is thought to occur in saturated micro-environments generated by bacterial decay of organic aggregates as the majority of the global ocean is under saturated with Ba (Dehairs *et al.* 1980; Bishop 1988; Monnin *et al.* 1999; Martinez-Ruiz *et al.* 2018). BaSO_4 has relatively high preservation rates compared to other biogenic tracers (e.g. opal, organic carbon, calcium carbonate) and can be easily measured using non-destructive methods (e.g. XRF), therefore biogenic Ba accumulation has been widely used to estimate paleoproductivity. However, there are uncertainties associated with using this proxy in ice marginal settings as regional differences in the correlation of Ba to productivity have been noted within the Southern Ocean (McManus *et al.* 2002; Presti *et al.* 2011; Sun *et al.* 2013), and therefore it is recommended to use a multi-proxy approach for inferring paleoproductivity (Bonn *et al.* 1998). On the Wilkes Land margin, high Ba/Al has been correlated to high productivity occurring in interglacials (Presti *et al.* 2011; Cook *et al.* 2013).

Biogenic silica ratios (e.g. Si/Al or Si/Ti) have been linked to biogenic silica content (diatom content) and nutrient content of ocean waters (Marsh *et al.* 2007; Agnihotri *et al.* 2008; Dickson *et al.* 2010). However, as both silica and barium occur in terrigenous sediment, a terrigenous normalizing factor must be used, such as aluminum (Al), which assumes 1) all Al is from a terrestrial source (Calvert & Pedersen 2007; Paytan & Griffith 2007), and 2) the terrestrial source is unvarying through time (Reitz *et al.* 2004).

2.3.6.2 *Provenance tracers*

The geochemical composition of sedimentary rocks depends on the type and weathering of source rocks, transport, depositional conditions, and diagenesis. Many Antarctic studies have used inorganic geochemical studies to determine sediment provenance and have revealed variations in sediment source through time, largely related to glacial–interglacial regimes. Al/Ti ratios can be used to investigate provenance of sediments as the elements are relatively immobile, undergo similar weathering patterns, but greatly vary in igneous source rocks, with higher ratios generally corresponding to a more mafic composition (Young & Nesbitt 1998). Ti content spans a much wider range than Al and is primarily hosted in Ti-bearing heavy minerals (Calvert & Pedersen 2007). Downcore variability in this ratio has been used to identify changes in source regions through time (Yarincik et al. 2000; Latimer et al. 2006; Orejola & Passchier 2013). However, Al/Ti may also reflect grain size variability, in that Ti is enriched in the coarser fractions, so a multi-proxy synthesis is required to interpret the data. Similarly, K/Ti has been used to reveal changes in sediment provenance in both the Ross Sea and Wilkes Land (Diekmann et al. 2008; Monien et al. 2012; Bertram et al. 2018).

CHAPTER 3: REGIONAL SETTING AND STRATIGRAPHIC FRAMEWORK

IODP Expedition 318 was undertaken to establish a history of climate and paleo-oceanographic variability through a transect of continental shelf to rise drill sites off the coast of Wilkes Land, East Antarctica. Site U1361 (Figure 3.1) was drilled with the aim of understanding the history of the EAIS and to assess the stability of the ice sheet during warmer-than-present periods during the Miocene and Plio–Pleistocene (Escutia *et al.* 2011). The Wilkes Subglacial Basin is the largest marine-based sector of the East Antarctic Ice Sheet, and is considered a key region for understanding the behaviour of the EAIS as modelling and data-based studies have shown this sector may be vulnerable to marine ice sheet instabilities (Cook *et al.* 2013; Mengel & Levermann 2014; Austomann *et al.* 2015; Golledge *et al.* 2017). High-latitude continental shelf settings are commonly hindered by hiatuses from glacial erosion (Naish *et al.* 2001, 2009; Passchier *et al.* 2017), so drilling of continental rise sediments provides an opportunity to assess a near-continuous record of glacially influenced sedimentation. These sediments may help to address questions of Late Cenozoic ice sheet climatic response, as the lithofacies associations are likely to represent a continuum of processes related to glacial extent and oceanography along the Wilkes Land margin.

Figure 3.1 Site location of U1361 and offshore Wilkes Land bathymetry, Antarctica. Black arrows denote down-slope currents flowing into Jussieu Canyon. White line is the position of the seismic line shown in **Figure 3.6**. Modified from Patterson *et al.* (2014).

3.1 Site location

IODP Site U1361 (64°14'42.0"S 143°31'51.6"E; Figure 3.1) is located on the continental rise at 3454 m water depth approximately 315 km offshore Adélie Land, East Antarctica. The drill site is situated on the right (east) levee of the Jussieu submarine canyon system where the levee relief is ~195 m. U1361 penetrated to a depth of 388 m below seafloor — spanning the Middle Miocene to Pleistocene (Tauxe *et al.* 2012).

3.2 Physiological setting

3.2.1 Wilkes Land Margin: Adélie Land and George V Coast

The Wilkes Land continental margin formed during the rifting of the Antarctic and Australian continents during the breakup of Gondwana (~85 Ma), which resulted in the faulting and extension of Early Cretaceous basement rock (Cande & Mutter 1982). Outcrop and offshore marine sediment provenance studies indicate the continent is geologically heterogeneous and contains sedimentary, metamorphic, and extrusive and intrusive igneous rocks with ages spanning the last 3 billion years (Goodge & Fanning 2010). These include Precambrian Terranes, the Beacon Supergroup, and Ferrar Large Igneous Province (FLIP) which have proved useful in sediment provenance studies tracing past changes in ice sheet dynamics (Cook *et al.* 2013, 2017). The continental shelf has an average water depth of 450–500, and an average width of 125 km that exhibits an over-deepened and landward-sloping bathymetric profile caused by glacial erosion and flexural loading by the ice sheet (Anderson 1991; Beaman *et al.* 2011).

The upper continental slope of the Wilkes Land margin is incised by several gullies that converge into submarine canyon and levee systems at 1000 m water depth, including the Jussieu and WEGA Canyons (Figure 3.2). The levee-canyon system is interpreted through seismic sections to have built up as a result of fine-grained turbidity flows beginning in the Early Miocene (Donda *et al.* 2003; Santis *et al.* 2003). The Jussieu Canyon is seaward of a deep (>1200 m) continental shelf-depression, termed the George V Basin or Adélie Depression (Beaman *et al.* 2011; Figure 3.3). Plumes of cold, dense water formed in the Adélie Depression flow over the continental shelf and are channelled through the canyon systems (Rintoul 1998; Bindoff *et al.* 2000).

Figure 3.2 Bathymetry of the Wilkes Land continental shelf and continental rise and site location of U1361. Modified after Beaman *et al.* (2011).

3.2.2 Wilkes Subglacial Basin

On the Wilkes Land margin, the EAIS drains through the Cook, Ninnis, and Mertz outlet glaciers that extend seaward as ice tongues (Rignot *et al.* 2011). Ice velocities in outlet glaciers range from 0.5 km/yr to 3.7 km/yr and have an important role in ice drainage and sediment delivery to the continental shelf (Drewry & Cooper 1981; Escutia *et al.* 2005). Ice cover within the WSB is up to 3 km thick and overlies ~1 km of sedimentary infill (Studinger *et al.* 2004). The underlying basin is a structurally controlled ~1400 km long basin, which at 600 km wide at the George V coast and narrows southward (Figure 3.3 Studinger *et al.* 2004; Ferraccioli *et al.* 2009). Aeromagnetic mapping of the basin shows inland-sloping topography and the average depth of the ice sheet bed is 500 m below sea level, with extensive sub-basins reaching depths of 2100 m below sea level (Ferraccioli *et al.* 2009).

The WSB is the largest region of the EAIS with topography below sea level and holds 3–4 m sea level equivalence of ice (Figure 3.3; Mengel & Levermann 2014). In recent years, the WSB has been a focus of attention because of its similar topographic configuration, and a similar amount of ice to the WAIS (Bamber *et al.* 2009), which may indicate this sector of the EAIS responds to similar drivers as the WAIS, and has an increased vulnerability to marine ice sheet instabilities (Schoof 2007; Pollard & DeConto 2009). The WSB has been identified in climate models as susceptible to undergoing ‘threshold’

behaviour, and that an only minor retreat of the grounding line under moderate warming may destabilise the entire sector of the ice sheet and retreat rapidly into the deepest sections of the basin (Mengel & Levermann 2014; Golledge *et al.* 2017). Modelling experiments that are parameterised by geological proxies of past temperature (see Golledge *et al.* (2017)) suggest an increase of sea level by as much as 8.5 m in the Early Pliocene (~4.23 Ma), primarily sourced from the WAIS and WSB. Provenance and sedimentological studies suggest the WSB underwent significant retreat several times during Early to Middle Pliocene warmth (Cook *et al.* 2013; Patterson *et al.* 2014). This highlights the importance of marine-based sectors of ice sheets, such as the WSB, as key regions for understanding ice dynamics in a warming climate and ocean.

Figure 3.3 Subglacial topography of WSB showing the substantial area of ice beneath sea level in this sector, making it vulnerable to marine ice sheet instabilities. Red signifies floating ice. U1361 is ~350 km northwest of Mertz Glacier. Modified from Mengel & Levermann (2014).

3.3 Southern Ocean and Wilkes Land oceanography

3.3.1 The Southern Ocean

The Southern Ocean is a driving force in the world's ocean and climate; cold, dense Antarctic Bottom Water (AABW) is formed on the Antarctic margin, which is then entrained by the ACC and transported throughout the global oceanic basins by deep

western boundary currents, becoming a vital component of the global thermohaline circulation (Figure 3.4; Carter *et al.* 2008). The ACC is the largest and strongest current system and flows west to east. It is driven by the Southern Hemisphere westerlies, and its pathway strongly coupled to bathymetry, with a top-to-bottom flow that extends to the seabed and may interact with seafloor sedimentation (Barker 2001). The ACC is also responsible for the exchange of water between the Pacific, Atlantic, and Indian Oceans around the Southern Ocean (Orsi *et al.* 1995; Rintoul *et al.* 2001). This inter-basinal exchange exports fresh water, heat, salt, and nutrients globally. The modern-day position of the southern boundary of the ACC is ~200 km north of Site U1361 (Orsi *et al.* 1995).

Figure 3.4 A map of major oceanographic currents surrounding Antarctica and the distribution and direction of the ACC and Deep Water Boundary Current (DWBC) inflow, which has a modulating component of the global thermohaline circulation. From Carter *et al.* (2008).

The Antarctic Polar Front Zone spans the area of the Subantarctic Front (SAF) and Antarctic Polar Front (APF), and in the modern day encompasses a transition from 2 to 5°C surface water temperatures and an increase of 0.2% in salinity (Gordon 1971; Gordon

et al. 1982). This change in environmental conditions results in a considerable difference in phytoplankton (Bohaty & Harwood 1998). South of the polar front, relatively warmer upper circumpolar deep water is upwelled, and as a result is an area of high biological activity (Whitworth *et al.* 1998; Bindoff *et al.* 2000). Near the continent, the Antarctic Slope Front (ASF) separates the cold, salty water forming on the shelf, from warm circumpolar deep water (CDW) offshore (Figure 3.5; Jacobs 1991)

Figure 3.5 Cross section of southern high latitudes oceanography. ACC-regulated intrusions of CDW may have a major effect on the stability of AIS. Water masses are AABW = Antarctic bottom water. LCDW = lower circumpolar deep water. NADW = North Atlantic Deep Water. UCDW = upper circumpolar deep water. AAIW = Antarctic Intermediate Water. SAMW = subantarctic mode water. ASF = Antarctic slope front. Frontal systems are SB = southern boundary of the ACC. SF = southern front. PF = polar front. SAF = subantarctic front. STF = subtropical front. From Carter *et al.* (2008).

3.3.2 Southern Ocean and Antarctic Ice Sheets

The ACC and its associated water masses within the Southern Ocean are thought to be intrinsically linked to the westerly winds, such that shifts in the intensity and location of the winds may have a direct influence on Southern Ocean circulation and modulate glacial–interglacial climate (Toggweiler & Russell 2008; Anderson *et al.* 2009; Bard & Rickaby 2009; Sikes *et al.* 2009; Martínez-García *et al.* 2011). Displacement of the westerly winds and intensification of the ACC may regulate intrusions of CDW onto the continental shelves around the Antarctic margins by enhancing the upwelling of warm deep water (Figure 3.5; Toggweiler & Russell 2008). Intrusions of CDW may be reduced by the Antarctic Slope Current (ASC) at the shelf break. The ASC creates a ‘v-shaped’ isopycnal

that extends into intermediate water depths and restricts sub-surface CDW incursions onto the continental shelf (Figure 3.6; Bindoff *et al.* 2000; Spence *et al.* 2014).

Changes to the strength or location of the polar easterly winds may change the location and intensity of the ASC (Thoma *et al.* 2008; Dinniman *et al.* 2012; Rintoul 2018). As Antarctic easterlies weaken and migrate southwards during interglacials, the advection of CDW onto the continental shelf may be enhanced. The interaction of warm CDW and grounding lines can increase melt rates by tens of meters a year (Joughin *et al.* 2012), resulting in the retreat of marine ice sheets (Naish *et al.* 2009; McKay *et al.* 2012; Pritchard *et al.* 2012; Hillenbrand *et al.* 2017). This process is thought to be responsible for modern-day retreat on the Antarctic Peninsula (Jenkins & Jacobs 2008; Wåhlin *et al.* 2010; Jacobs *et al.* 2011) and Amundsen Sea (Steig *et al.* 2012; Jenkins *et al.* 2018).

3.3.3 Antarctic Bottom Water (AABW)

The Wilkes Land margin is one of the three largest producers of AABW, alongside the Weddell and Ross Seas (Rintoul 1998, 2007; Bindoff *et al.* 2000). High-salinity shelf water (HSSW) is produced by brine rejection under enhanced sea ice production within polynyas (Rintoul 1998). These dense waters spread across the shelf and flow down the continental slope, and are funnelled into slope canyons, such as the Jussieu canyon, to produce AABW (Williams *et al.* 2008).

Modern observations suggest basal melting of the Mertz Glacier Tongue doubles under weak polynya activity because of warm water reaching the base of the ice shelf. Weaker polynya activity allows increased intrusions of warm modified CDW within the ice shelf cavity and increases basal melting, which in turn limits the formation of Dense Shelf Water (Cougnon 2016). A similar process has been observed in Prydz Bay (Williams *et al.* 2016).

3.4 Stratigraphic framework

Comprehensive seismic surveying of the Wilkes land margin identified nine seismic sequences representing four seismic units associated with the development and stabilisation of the East Antarctic Ice Sheet (Table 3.1; Donda *et al.* 2007). U1361 targeted and achieved penetration of unconformity WL-U6 in order to establish the timing and nature of a shift in sedimentary depocenters associated with the transition from a

dynamic EAIS to a more persistent cold-based EAIS (Escutia *et al.* 2011). Drilling confirmed that the youngest unit (Unit 4) contains a continuous sediment record for the Late Miocene to Pleistocene, with erosional surfaces WL-U7 at ~300 m and WL-U8 at ~100 m (Figure 3.6). The record relevant to this thesis lies between 151 and 104 mbsf, directly below WL-U8, within sedimentary package WL-S8.

Figure 3.6 Seismic reflection profile of the Jussieu Canyon system and U1361. This profile is approximately parallel to the levee crest. Coloured lines represent seismic uniformities shown in **Table 3.1**. This thesis focuses on the interval directly below the interpreted depth of WL-U8 (blue line). From Escutia *et al.* (2011)

Seismic interpretation of Unit 4 (Late Miocene to Pleistocene) describes the development of thick, prograding shelf-edge wedges on the continental shelf and slope of the Wilkes Land margin, which suggests that from the Late Miocene the ice margin regularly reached the shelf edge (Donda *et al.* 2007). This differs from Unit 3 in that the bulk of the sediment remained within the prograding wedge and was sequestered on the continental shelf, rather than being delivered to the continental rise. This decreased sediment supply to the continental rise reduced the building of the channel-levee system on the continental rise, however, turbidity currents and down-slope current activity continued, as confirmed by sedimentary studies (Escutia *et al.* 2011), and were probably related to occasional slope-failures and bottom water formation related to the Mertz Polynya. Seismic interpretation indicates bottom currents are not a dominant process at this site during the Late Miocene to Pliocene, with well-developed sediment waves recognised in the older upper Oligocene

to Miocene sequences on the continental rise, and at sites higher on the continental rise (Presti *et al.* 2003; Escutia *et al.* 2011).

Table 3.1 Wilkes Land seismic stratigraphy. Unconformities are shown in **Error! Reference source not found.**. WL = Wilkes Land, S = Sequence = S, U = unconformity. From Donda *et al.* (2007).

Unconformity	Seismic Sequence	Seismic Unit	Deposition
WL-U8	WL-S9	Unit 4	Late Miocene to Pleistocene: Decreased sediment supply to the continental rise, sediment collected on the shelf, a more stable and less erosive (colder) ice sheet.
	WL-S8		
WL-U7	WL-S7	Unit 3	Early Miocene to Late Miocene: Maximum sediment supply through turbidity currents to the continental slope and rise, development of channel-levee complexes and formation of ridges and drifts. Dynamic and erosive ice sheet.
WL-U6	WL-S6		
WL-U5	WL-S5		
WL-U4	WL-S4		
WL-U3	WL-S3	Unit 1	Palaeocene to Oligocene: Early post-rift period with thermal subsidence of the continental margin and no grounded ice sheets on the margin.
	WL-S2		
WL-U1	WL-S1		Syn-rift sequences.

On the Wilkes Land margin, Caburlotto *et al.* (2010) described distinct sedimentation within the levee systems related to the distance from the channel axis to the crest ridges. Proximal to the crest, turbidite sequences reflect high energy depositional processes with thick and well-defined laminations with no IBRD and sharp to irregular bases. On distal channel settings, turbidites are low-density, fine-grained, cm-thick laminated sequences. Site U1361 lies on the lowermost continental rise, with the position of the core site allowing for the preservation of low-density turbidite deposition, as the fine-grained material creates a non-erosive ‘overspill’ drape from inter and overflow processes (Patterson 2014), while erosion occurs mostly within the canyon system itself and on ridges higher up in the canyon system (Escutia *et al.* 2005).

Patterson (2014) refined the existing lithofacies scheme and depositional model of continental rise settings around the Antarctic margin and found sediment deposition at

U1361A closely aligns with established models of non-erosive overbank turbidite deposition in a lower continental rise ‘distal’ levee system (Lucchi *et al.* 2002), with two dominant lithofacies related to glacial–interglacial variability (Table 3.2).

Table 3.2 Refined facies scheme relating to glacial-interglacial processes for Wilkes Land margin from Patterson (2014).

Climatic Interpretation	Lithofacies
Glacials	Massive and laminated mud, frequent sharp-based thin silty laminae, and decrease in biogenic material, sparse IBRD.
Deglaciations	Similar to interglacials but with peaks in IBRD.
Interglacials	Diatom-rich (/bearing) mud dominated by biogenic (pelagic) sedimentation, sparse IBRD, and homogenisation of sediment by bioturbation, increasing mean silt component. Increase in biogenic opal weight percent and Ba/Al.

CHAPTER 4: METHODOLOGY

4.1 Core Recovery

Cores 1–16 (0–151.1 m) at Site U1361 were recovered using the advance piston coring (APC) system and the rest drilled using the extended core barrel system, with 87% recovery. Cores 12H–16H were collected using APC between 104 and 151.5 mbsf (CSF-B). Core depths were corrected using IODP nomenclature, where CSF-A is non-corrected (e.g. depths may overlap) and CSF-B is corrected to absolute depth, whereby cores with greater than 100% recovery (due to expansion) are linearly compressed to the same length as the core run (usually 9.5 m) (Table 4.). Each core run is separated into six sections (labeled 1W– 6W) and a core catcher (CC). Naming convention of IODP cores is Expedition-Hole-Core-Section-Half-Interval, where core describes drilling type ('H' for hydraulic in this section), 'W' is working half, and interval is in centimetres below the top of the section.

Table 4.1 Core recovery from U1361 between 104 and 151.7 m.

Core	Top Depth CSF-A (m)	Bottom Depth CSF-A (m)	Advance (m)	Recovered Length (m)	Curated Length (m)	Top Depth CSF-B (m)	Bottom Depth CSF-B (m)	Recovery (%)
12	104.0	113.5	9.5	7.925	7.925	104.0	111.32	77
13	113.5	123.0	9.5	9.07	9.07	113.5	122.57	95
14	123.0	132.5	9.5	9.57	9.57	123.0	132.57	101
15	132.5	142.0	9.5	9.26	9.26	132.5	141.76	97
16	142.0	151.5	9.5	9.7	9.7	142.0	151.7	102

Recovery of the studied interval section of U1361A is reasonably good, with an average of 97% recovery, excluding core 12H which missed ~1.6 m. Cores 12H–16H 1W contained slight drilling disturbance throughout the core, with extreme disturbance in only core 12H section 1–2W (104–106.24 m). A total of 444 samples were requested from the IODP repository in College Station, Texas (USA). 423 samples were processed for XRF and IBRD analysis at an average sample spacing of 10 cm. A total of 294 samples were processed for grain size analysis using a Laser Particle Sizer Analyser (LPSA).

4.2 Age model

The original age model for this core was constructed using biostratigraphy and magnetostratigraphy with relatively good confidence in the stratigraphic position of six paleomagnetic reversals within, or just outside of, the studied interval (Table 4.; Tauxe *et*

al. 2012). This age model produces fairly uniform sedimentation rates between 12 and 16H decreasing from 33 m/myr to ~27 m/myr in the Late Pliocene.

Table 4.2 Age and stratigraphic depth of paleomagnetic tie points from Tauxe *et al.* (2012).

Occurrence	Age (Ma)	Top Section	Bottom Section	Top (m)	Bottom (m)
C3n.1n (y) (Cochiti)	4.187	11H-3, 82	11H-4, 99	98.32	98.32
C3n.2n (o) (Nunivak)	4.631	12H-6, 40	12H-6, 40	109.69	109.69
C3n.4n (y) (Thvera)	4.997	13H, 80	13H, 85	120.3	120.35
C3n.4n (o) Thvera)	5.235	14H-1, 65	14H-1, 70	123.65	123.7
C3An.1n (y)	6.033	16H-2, 30	16H-2, 35	143.8	143.85
C3An.1n (o)	6.252	17X-2, 80	17X-2, 80	153.8	153.8

The age model for this thesis was reassessed independently of the existing age model using a constrained optimization (CONOP) model of biostratigraphic constraints of first appearance and last appearance dates of radiolarians and diatoms (Cody *et al.* 2012). Following the method of Crampton *et al.* (2016), two methods — a relaxed and a strict hybrid model were ran (independent of this project; see Crampton *et al.* (2016) for details of these models) to give an envelope of possible biostratigraphic ages for any given depth. This differs from the ‘average’ CONOP range model of Cody *et al.* (2008), used by Tauxe *et al.* (2012) to construct the original age model. Notably, numerous biostratigraphic constraints used by Tauxe *et al.* (2012) cannot be directly reconciled by the magnetostratigraphy. However, using the hybrid CONOP model, as recommended by Cody *et al.* (2012) and Crampton *et al.* (2016) for Antarctic proximal sites, the revised age model used in this thesis demonstrates a much better fit between biostratigraphy and paleomagnetic constraints. A line of best fit could then be constructed within the CONOP error envelope to identify tie points of magnetostratigraphy to paleomagnetic reversals (Figure 4.1). While perturbations to the age model are possible due to the sparse nature of biostratigraphic sampling in this interval, the age model in this thesis follows that of previous studies (Tauxe *et al.* 2012; Cook *et al.* 2013). Higher resolution biostratigraphic sampling could be undertaken to assess if remaining mismatches between the magnetostratigraphy and the CONOP model is a consequence of sample spacing for the biostratigraphy, but this is beyond the scope of this thesis.

Figure 4.1 Age model for Site U1361 between cores 1 and 16 following Tauxe *et al.* (2012). Blue shading denotes CONOP error envelope. Tie points within this study are shown in orange. Pink sections of magnetostratigraphy are sections with poor paleomagnetic data recovery.

4.3 Lithofacies analysis

A lithofacies scheme was constructed for the cores using shipboard core descriptions that were developed using visual core descriptions and smear slides to determine microfossil classification and abundance (Escutia *et al.* 2011). The lithological name of the sediment was based on the relative abundances of biogenic and terrigenous material, with the

principle name of sediment with <50% biogenic grains based on the grain-size characteristics of the terrigenous clastic fraction (after Moncrieff 1989). If the sediment contains no gravel, then the name is based on the abundance of sand, silt, and clay (Figure 4.2). If the sediment contains gravel, the principle name is determined by the relative proportion of gravel and the sand/mud ratio of the terrigenous matrix (Figure 4.3).

Figure 4.2 Lithological name of sediment was based on the relative percentage of terrigenous to biogenic material.

Percent gravel (>2 mm) in whole rock estimated from core					
	Trace-<1	1-5	5-30	30-80	>80
0	Clay/Silt with dispersed clasts	Clay/Silt with common clasts	Clay/Silt with abundant clasts	Clayey/Silty conglomerate/breccia	Gravel/ Conglomerate/ Breccia
25	Sandy clay/silt with dispersed clasts	Clast-poor muddy diamict	Clast-rich muddy diamict	Sandy muddy conglomerate/breccia	
50	Clayey/Silty sand with dispersed clasts	Clast-poor sandy diamict	Clast-rich sandy diamict	Muddy sandy conglomerate/breccia	
75	Clayey/Silty sand with dispersed clasts	Sand with common clasts	Sand with abundant clasts	Sandy conglomerate/breccia	
100					

Figure 4.3 Classification scheme for terrigenous clastic sediments containing gravel. After Moncrieff (1989).

All sediment within the core sections were dominated by terrigenous material. If the sediment contained a biogenic proportion, modifications were applied if the components were between 25 and 50% indicated by the suffix '-rich' and between 10 and 25% indicated by '-bearing' (Figure 4.2; Escutia *et al.* 2011). Refinement of lithofacies were developed using XRF measurements, examining core photos, and manual counting of laminae using functions in Matlab, in order to create a higher-resolution stratigraphic column than the generalised stratigraphies that were developed during shipboard descriptions. To refine shipboard descriptions, as smear slides were only performed every

1 m, physical properties such as bulk density (GRA) and Natural Gamma Ray (NGR) were used to estimate regions of biogenic silica. Assuming a bulk density less than 1.6 cm/g^2 and low NGR (<30) signified regions of increased diatom content, as diatoms are much less dense than terrigenous grains.

4.4 Sample pre-processing

423 10cc core plug samples were processed for XRF and IBRD analysis. The samples were first sprayed with deionised (DI) water prior to freezing and then freeze-dried for at least 24 hours. Increasing the moisture content of the sample meant the process of freeze drying disaggregated the sample without force. Representative subsamples were taken for XRF ($\sim 2\text{--}3 \text{ g}$) and grain size ($\sim 1 \text{ g}$). The XRF sample was returned to the bulk sample after measurements were made and prior to processing for IBRD.

4.5 X-Ray Fluorescence

A bulk $2\text{--}3 \text{ g}$ subsample was placed in 5 ml plastic vials 15 mm in diameter, with any large ($>2 \text{ mm}$) clasts removed as they may skew the results of the bulk composition of the sediment. The vials were covered in $4 \mu\text{m}$ polypropylene and measured using a handheld Olympus Vanta M-series XRF scanner connected to a workstation that allowed remote operation.

Each sample was measured using two methods (Geochem and Soil), calibrated by Olympus to yield quantitative values for uncompressed powders. Each method varies the energy of the primary beam, allowing the measurement of different elements. The Geochem method utilises two beams of 40 kV and 10 kV, and were set to irradiate the sample for 30 and 90 seconds respectively. The long integration time for the 10 kV beam was used in order to improve detection and quantification of light elements (e.g. Al, Ca, Fe, K, Mg, Si). The Soil method runs three beams at 50 kV, 40 kV, and 15 kV for 30 seconds each and is used in order to measure and improve the accuracy of Ba and Ti.

The nominal calibration provided by Olympus was not used in this study. Instead, seven rock powder standards from US Geological Survey and Geological Survey of Japan were measured at the beginning, after ~ 3 hours, and at the end of the day to calibrate sample data to absolute concentrations. Corrections were calculated using linear least squares

between the average of each standard and the known measurement of each element. The slope coefficient is then applied to each measurement of corrected elements. This data can be found in Appendix D.

4.6 Iceberg rafted debris

4.6.1 Iceberg rafted debris mass accumulation rates

A total of 423 samples were sieved in order to calculate iceberg rafted debris mass accumulation rates (IBRD MAR) using the weight percent of the 250 μm –2 mm fraction of coarse sand, following the method established by Krissek (1995). This method was used in order to compare our findings to results from other sections of U1361 (Patterson 2014). As found by Albot (2016), different methods of establishing IBRD MAR are not comparable (see section 2.3.5.1).

Approximately 10 cm^3 of dry sample was first weighed, then wet-sieved at 250 μm and the coarse fraction was placed underneath a lamp until dry. The weight of >250 μm was recorded before any >2 mm clasts were removed and then reweighed. Clasts >2mm may skew IBRD MAR and are therefore removed. All 250 μm –2 mm material were checked under a binocular microscope and the volume percent of biogenic material was recorded, as well as the presence of any volcanic or authigenic minerals. No samples contained substantial authigenic or volcanic material, although 5 samples contained a small amount of pumice, and these samples were removed from IBRD calculations.

Once terrigenous material was isolated, mass accumulation rates could be estimated using:

$$IBRD\ MAR = CS\% * DBD * LSR$$

Where IBRD MAR is the mass accumulation rate ($\text{g}/\text{cm}^2/\text{ky}$), CS% is the weight percent of 250 μm –2 mm fraction, DBD is the dry-bulk density of the nearest value (g/cm^3), and LSR is the average linear sedimentation rate (cm/ky).

4.6.2 Removal of biogenic material

Samples with an abundant (>~20% volume) biogenic component were treated to remove the siliceous material to ensure IBRD is an entirely terrigenous signal. The siliceous (diatoms and radiolarians) material was removed by softly crushing with gentle finger pressure the >250 μm fraction between paper, then resieved at 250 μm and reweighed. The samples were then examined again for biogenic volume percent. Following this treatment, the reduction in biogenic volume varied between 10 and 55%, with an associated weight difference of 8–46% (average 19.5%). The difference in total weight of IBRD (e.g. relative to the bulk sample weight) varied between 2.6 and 27% (average 9.7%). This indicates that despite the volume percent of biogenic material within the 250 μm –2 mm fraction being up to 70%, the associated weight of biogenic material is small, and thus processing for samples with <20% biogenic silica results in minimal errors (e.g. <3% of the IBRD weight percent value). However, for methods using volume estimates of IBRD using LPSA measurements (e.g. Passchier *et al.* 2011), this error is likely to be more significant unless all biogenic material is removed.

4.6.3 Sample reproducibility

Ten 20 cm^3 (2 x 10cc core plugs) samples were split and processed to test the reproducibility of IBRD data. Each core plug was treated separately and the samples were not homogenised and then separated before processing. The difference between sample A and B ranged from 0.0001 to 0.009%, with an average 0.0032% difference (Figure 4.4), and analysis found a moderate relationship between samples, indicating reasonably good reproducibility ($R^2 = 0.53$; Figure 4.5). When combined into individual downcore data series, both datasets show the same pattern, and in general, despite a couple of outlier samples there is a good correlation – e.g. the removal of a single outlier value at 109.5 mbsf demonstrates an R^2 value of 0.75.

Figure 4.4 Reproducibility of IBRD repeats between 107 and 111 m show relatively good reproducibility with all major peaks and trends represented. Average weight percent difference between sampled repeats 0.003%.

Figure 4.5 Cross-correlation plot between repeated samples shows a moderate correlation ($R^2 = 0.533$).

4.7 Bulk Grain Size Analysis (<1500 μm)

The ~1 g subsample was first dry sieved at -0.5 phi (1500 μm) to remove the coarse material that the LPSA machine cannot analyse accurately. The bulk sample was then subsampled based on the methodology of Lewis & McConchie (1994) of coning and

quartering. The disaggregated sample was shaken and dumped onto paper and carefully quartered with a metal spatula, on the principle that the way the grains are distributed when dumped into a cone, each quarter has an equal probability of containing grains of any particular size, shape, and specific gravity (Lewis & McConchie 1994). Initially, three subsamples were taken using this method (labelled A, B, C) in a 50 ml centrifuge tube in order to test the reproducibility of the data set.

4.7.1 Chemical treatment for Laser Particle Size Analysis

Each subsample was then digested in 10% hydrogen peroxide (H_2O_2) at room temperature for 24 hours, stirring every ~30 minutes, to chemically remove all organic matter. After 24 hours, the pH was tested, and if the fluid was still acidic, it was likely all the organic matter had been removed. The samples were then heated in a 70°C water bath and monitored for any activity for 2 hours to ensure the reaction was complete. Because the samples contained very low amounts of organic matter, this process was sufficient to remove all organic matter. The samples were then washed with ~40 ml of DI H_2O and centrifuged at 4700 rpm for 10 minutes, three times (until neutral pH 7), to remove any remaining H_2O_2 and reaction product. The calcium carbonate (CaCO_3) content of this section of the core was nil, so the removal of carbonate was deemed unnecessary.

The removal of biogenic silica was necessary for all samples as most of the core was described as diatom-rich to diatom-bearing. Initially, eight samples were treated with 10 ml of 1M sodium hydroxide (NaOH) and heated in a 70°C water bath (Lewis & McConchie 1994). Smear slides were made every 24 hours to check remaining biogenic silica content, but after 72 hours, sponge spicules were still present indicating this method did not remove all biogenic silica. Following the method of Patterson *et al.* (2014), the rest of the samples were treated with 2M NaOH for 48 hours in a 70°C water bath and stirred occasionally. The samples were then washed with DI H_2O and centrifuged at 4700 rpm for 10 minutes until neutral (3–4 times).

4.7.2 Laser Particle Size Analysis

Once all non-terrigenous material had been removed, samples were analysed using a Beckman Coulter LS 13 320 LPSA with the aqueous liquid module to obtain grain size

distributions. The samples were dispersed in ~80 ml of 0.5 g/L calgon ($\text{Na}_6\text{P}_6\text{O}_{18}$) and stirred in an ultrasonic bath for at least 20 minutes. Between the ultrasonic bath and the LPSA, the samples were stirred by hand to keep sediment in suspension and fine grains dispersed. The laser requires a specific range of obscuration (6–14%) to ensure accurate measurement of grain size, with higher mud content increasing obscuration.

The laser was realigned and the offset measured every 30 minutes, the background measured before every sample, and the unit automatically rinsed for 2 minutes before and after each run. The measurements were converted to grain size using the QuartzNatural.rf780d optimal model, which assumed all material has a refractive index of quartz suspended in water. Glass bead standards were run at the beginning and end of every day to ensure the machine was measuring accurately (Figure 4.6). Sample statistics were then processed using GRADISTAT, a Microsoft Excel macro package (Blott & Pye 2001).

Figure 4.6 Grain size distribution of 63 µm glass bead standards repeats show consistent estimation of grain size with the applied model.

4.7.3 Grain size reproducibility

In order to test the reproducibility of the results, 66 samples were subsampled twice and 12 samples were subsampled three times, and then processed and measured on the LPSA. All lithofacies show reasonable coherence between measurements and show high reproducibility in the clay to silt fractions (Figure 4.7). Samples with a high sand percent show less coherence in terms of volume percent, but in general, show the same

distribution. The lack of coherence in the sand fraction can be explained by the presence of distinct modes in the sand fraction, likely attributed to IBRD or coarse turbidite sand. As the LSPA relies on a volume-based estimate of the distribution and volume of sediment being measured in mud-rich samples, a few large grains can provide a disproportionate signal and difficult to reproduce between separate sample runs (Albot 2016).

Figure 4.7 Grain size distributions of repeated samples to test for reproducibility. In general, repeats show the same trend, with the exception of samples with high coarse sand. **a.** mud, **b.** diatom-bearing, **c.** diatom rich.

4.8 Spectral Analysis

Frequency analysis of datasets was performed using the Astrochron package for R (Meyers 2014). Spectral analysis allows the quantitative identification of cycles within a dataset that may underpin the behaviour of the system. Spectral analysis was performed on the IBRD MAR dataset at different time-steps to assess the relative periodicities at different stages in the dataset. It was also performed on key elemental ratios from XRF data. Datasets were linearly interpolated and detrended prior to analysis. These functions are used to approximate values to insert into the dataset to maintain even sample spacing, and to remove any trends from the mean of the dataset. The two main functions that were used to test the datasets for astronomical cycles were multitaper method and evolutionary harmonic analysis.

4.8.1 Multitaper Method

The mtmML96 function within astrochron is based on the original multitaper (MTM) algorithm of (Thomson 1982), where a series of special data windows are used to taper the time series and allows an increase in degrees of freedom without the rate of widening bandwidth (Weedon 2003). MTM reduces estimation biases by obtaining multiple independent estimates from the same sample, and the spectrum is then based on the average generated from each tapered series (Meyers 2012). Astronomical cycles are

identified by testing for significance relative to the null hypothesis of a red-noise background (the AR1 coefficient). Paleoclimate records often have a large red noise component, and the AR1 coefficient estimates this natural red noise component of the dataset, and only spectra more significant than this level is considered a signal. The original Mann & Lees (1996) approach applies a truncation of the median smoothing window to include fewer frequencies near the edges of the spectrum; while truncation is required, the original method often results in an 'edge effect' that can produce false positives at low frequencies (Meyers 2012). Instead, the mtmML96 method implements a new robust red noise Tukeys 'robust end-point rule' (Tukey 1977) to help overcome undesirable edge effects. This produces uniform false positive rates across the spectrum, so while there are still edge effects they are substantially reduced.

4.8.2 Evolutive Harmonic Analysis

To assess periodicities in the time domain, the Evolutive Harmonic Analysis (EHA) function is used which is based on moving window adaptations of MTM. The size of the moving window is a compromise between the maximum resolution of sedimentation rates and the maximum frequency resolution — this is the minimum window size that will still statistically capture significant orbital components (Meyers *et al.* 2001). A window size of 600 was used in the analysis, as for an average sedimentation rate of ~3 cm/kyr, 100 kyr eccentricity cycles should be captured. The significance of the periodicities is presented by a colour scale with warmer colours (red) reflecting more significant periods.

CHAPTER 5: RESULTS

5.1 Age Model

The age model for this thesis was constructed independently from the established age model from Tauxe *et al.* (2012), and although the use of the Hybrid Relaxed and Strict CONOP biostratigraphic age models has reconciled offsets with magnetostratigraphy, the same paleomagnetic tie points that were identified by Tauxe *et al.* (2012) are used here. This new age adds confidence to possible further constraint on the start of Chron C3n.3n (Sudufjall) and Chron C2n.1n (Cochiti), which were previously uncertain due to initial poor paleomagnetic data at the end of cores 12H and start of 13H (Table 5.1) — although determining the exact depth of these transitions remains difficult. Our refined age model fits cores 12H–16H between ~6.2 Ma and 4.4 Ma, beginning in the early Gilbert (Chron C3n.1r) and ending in Chron C3An.1n, with the base of the Pliocene in the uppermost Chron C3r at ~126 m (Figure 5.1). Although these age model constraints provide a non-unique solution and uncertainty, this uncertainty is greatest within the areas of poor core recovery between cores 11H and 12H, and at the base of 12H, where the four subchrons of Gilbert (Chron C2n.1 to C3n.4n) are not well expressed in the magnetostratigraphy. However, long-term sedimentation rates derived by the CONOP-based age envelope uncertainties support this as being the best-fit age model, while the long duration reversed interval in cores 14H and 15H is more reconcilable with the long duration Chron C3r.

Table 5.1 Magnetostratigraphic tie points used to construct the age model for this thesis. Bold intervals are additional tie points from Tauxe *et al.* (2012), but are the most uncertain. PM = paleomagnetic tie points.

Type	Occurrence	Age (Ma)	Top	Bottom	Top (m)	Bottom (m)
PM	C3n.1n (y) (Cochiti)	4.187	11H-3, 82	11H-4, 99	98.32	
PM	C3n.1n(o) (Cochiti)	4.300	11H-4, 27	11H-4, 27	100.18	
PM	C3n.2n (o) (Nunivak)	4.631	12H-6, 40	12H-6, 40	109.69	109.69
PM	C3n.3n (o) (Sidufjall)	4.896	13H-1, 55	13-1, 55	114.05	
PM	C3n.4n (y) (Thvera)	4.997	13H-5, 80	13H-5, 85	120.3	120.35
PM	C3n.4n (o) (Thvera)	5.235	14H-1, 65	14H-1, 70	123.65	123.7
PM	C3An.1n (y)	6.033	16H-2, 30	16H-2, 35	143.8	143.85
PM	C3An.1n (o)	6.252	17X-2, 80	17X-2, 80	153.8	153.8

Sedimentation rates range from 1.4 cm/kyr to 6.1 cm/kyr, with an average 3.7 cm/kyr. The highest sedimentation rates occur between 114 and 120 m, with 6.25 m of sediment deposited within the 101 kyr C3n.3r (Figure 5.1). Mud lithofacies have a slightly higher average sedimentation rate of 3.7 cm/kyr, vs 3.1 cm/kyr of diatom-rich (/bearing) lithofacies.

Figure 5.1 Magnetostratigraphic tie points within the timeframe of this thesis and linear sedimentation rates between tie points. Dashed line denotes average LSR (~3.1 cm/kyr).

5.2 Lithofacies analysis

Site U1361 lithofacies follows the shipboard descriptions (Escutia et al. 2011) with two major lithofacies identified; diatom rich/bearing mud and massive and laminated mud (Figure 5.2), directly applicable to 'Unit 2a', as defined by Escutia et al. (2011). Sixteen lithological cycles are identified based on the sequence of mud to diatom-bearing/rich facies (Figure 5.10). Most cycles contain a facies sequence of mud to diatom-bearing to diatom-rich spanning at least 3–4 m, except in the middle section (cycle 3–10) where cycles are shorter in thickness (1–2 m). Although this is the results section, facies are commonly associated with depositional processes, and an interpretation immediately follows each facies description for reader convenience.

Figure 5.2 Core photographs of intervals that are representative of the two primary facies types within the Late Miocene to Early Pliocene section of U1361.

Massive and laminated mud

The massive and laminated muds are equivalent to Facies E from the initial Expedition 318 Site Summary (Escutia et al. 2011). This facies is dark-grey to dark-greenish grey with

absent-to-moderate bioturbation, although the intensity of bioturbation decreases above 130 mbsf. IBRD is dispersed to rare within this facies and contains low Ba/Al and biogenic silica. Mud facies have common wispy to parallel silty laminae to dm-scale sandy silty beds with sharp bases that fine to massive muds. The tops of laminae and silty beds are often bioturbated. Laminae and silt beds contain no IBRD or biogenic material and are normally graded. In total, 337 laminations were identified and occurred almost exclusively within mud facies. In some areas (e.g. cycle 1, 3, and 8), laminae are concentrated in the beginning and end of the cycle, but in most cases, the distribution of laminae is relatively sporadic. On average the laminae are 5 mm thick, with only 49 silt beds (14%) exceeding 1 cm, with the largest reaching 3.6 cm (Figure 5.3). Silt beds and laminae are most frequent in cores between ~131 and 110 m.

The structure of laminae and silt beds within the mud facies are consistent with levee over-spill of low-density turbidity flows, and show similarities to previous facies schemes from the Antarctic margin (e.g. Lucchi *et al.* 2002; Patterson 2014). This indicates the laminae represent times when the ice sheet was expanded onto the continental shelf during glacials, during which sediment was delivered straight to the continental slope. Low Ba/Al and biogenic silica indicate marine productivity was reduced, possibly as a result of increased sea ice and/or upwelling of nutrients.

Diatom-rich (bearing) silty clays

The diatom-rich (bearing) silty clays are equivalent to Facies D from the initial IODP Exp. 318 Site U1361 (Escutia *et al.* 2011). This facies is light-olive green to dark-olive grey with no to moderate bioturbation and contains >25 wt% diatoms (rich; >15 wt% bearing). This facies is relatively structureless, apart from rare and discontinuous silt laminae/lenses and common/occasional IBRD and dispersed clasts. Diatom-rich sections contain high Ba/Al and biogenic silica.

Diatom-rich intervals are consistent with glacial minima, with a lack of laminae indicating a reduction in turbidite activity as the ice sheet retreats. High biogenic content indicates increased marine productivity. IBRD peaks generally occur at the transition between facies which is evidence for accelerated calving at the Wilkes Land margin and EAIS outlet

glaciers from the Ross Sea as the ice sheet retreats (Patterson 2014). Diatom-rich (/bearing) facies are interpreted to represent decreased turbidite activity, however, turbidites can still be triggered by slope failure on the continental slope due to isostatic rebound of the deglaciated margin (Anderson *et al.* 1979; Escutia *et al.* 2005), or cascading dense water forming on the continental shelf flow down the continental rise canyon as HSSW production is increased following deglaciation (Caburlotto *et al.* 2010). Regardless of sediment delivery processes, primary turbidite structures are no longer present/obscured in this facies, due to extensive reworking by bioturbation (due to oxygenated waters) and possible enhanced bottom current activity.

Figure 5.3 Distribution and thickness of laminae and silt beds identified within the core photos associated with turbidity flows. Laminae / silt beds were defined as a layer over half the width of the core. Dashed line denotes transition from laminae to bed at 1 cm.

5.2.1 Estimation of biogenic silica

Biogenic silica (BSi) was semi-quantitatively estimated using bulk density measurements. As porosity within the core does not vary significantly ($66\% \pm 5$), it is assumed that the changes in bulk density are related to changes in grain density. Assuming a porosity of 0.66, grain density was estimated using:

$$G\rho = B\rho - (1.024 * \Phi)$$

Where $G\rho$ is grain density, $B\rho$ is bulk density (GRA), 1.024 is the density of seawater, and Φ is porosity. Once grain density was calculated for every sample, biogenic silica was

estimated by calculating the relative contribution of lithogenic grains and biogenic silica using:

$$BSi = 1 - \left(\frac{G\rho - 2}{L\rho - 2} \right) * 100$$

Where it is assumed a lithogenic grain ($L\rho$) has a density of at least 2.7 g/cm^3 and biogenic silica has a density of 2.0 g/cm^3 . This equation results in samples over 2.7 g/cm^3 to have a negative value, which have been removed from figure 5.4. It is unlikely these sections have 0% biogenic silica, but this method was used to highlight regions where biogenic silica is high. It is acknowledged this is not a method to estimate exact biogenic silica values, and percentage values should be treated as semi-quantitative. However, BSi results compared to the stratigraphic log, XRF data (e.g. Si/Al and Ba/Al), and previously published data (e.g. diatom concentrations, and low-resolution opal percent from Cook *et al.* (2013)) show a moderate correlation to areas of estimated high productivity and diatom-rich facies (grey boxes; Figure 5.4). This method also suggests that Si/Al is not a wholly biogenic signal, however, peaks of Si/Al (>7) show a good relationship to BSi and Ba/Al.

Figure 5.4 Estimation of biogenic silica (blue) compared to Si/Al and diatom valve concentration (black). Highly productive periods ($Ba/Al = >0.019$) are shaded grey. High biogenic silica is coincident with Si/Al over ~ 7 . Diatom valve concentration (diatom valves per gram) from Cook *et al.* (2013).

5.3 Grain size

Figure 5.5 presents a ternary diagram of the ratio of sand, silt, and clay content between facies. These results use an $8 \mu\text{m}$ clay–silt boundary due to the issues with LPSA method underestimating mud percentages in highly cohesive sediments (Sperazza *et al.* 2004;

McCave *et al.* 2006). The majority of samples plot as clayey silts or silty clays, with mud diatom-rich (/bearing) facies containing a slightly higher silt and sand percent.

Figure 5.5 Ternary diagram displaying textural variability between massive and laminated mudstone facies and diatom-rich(/bearing) mudstone facies.

5.3.1 Grain size distribution

All samples have a bimodal distribution with a dominant silt mode and a smaller sand mode. Diatom-rich facies contain a silt mode that centres around 10 μm and a sand mode beginning around 120 μm and centred on $\sim 150 \mu\text{m}$ (Figure 5.6a). Diatom-rich sediments contain higher amounts of sand and coarse-to-medium silt with less fine silt and clay. Whereas, in the massive and laminated mud facies, the silt mode is around 5 μm , sand around $\sim 180 \mu\text{m}$, and contains a higher amount of fine silt and clay (Figure 5.6b). The diatom-rich samples are more uniform in distribution than mud facies, probably as a result of increased bioturbation within diatom-rich sediments homogenising any preexisting laminae. While the texture of mud facies is more variable, this is a reflection of massive muds and silty laminae and heterogeneity associated with graded turbidite beds. Laminae have a much broader distribution, with no distinctive mode except in sandy silt beds (Figure 5.7).

Figure 5.6 Grain size distributions of a. diatom-rich mud b. massive and laminated mud. Diatom-rich muds are slightly coarser with higher amounts of IBRD. Mud facies are slightly finer but are less consistent due to the prevalence of mottled silt.

Figure 5.7 Grain size distribution of laminae and silt beds demonstrate that the laminae are much coarser than massive mud.

Thick sand beds (>2 cm, e.g. larger than the sample width) are distinctive from the surrounding sediment and are characterised by a well sorted, unimodal distribution of fine sand (~70 μm) with little to no silt or clay (Figure 5.8). Internally massive sand beds with sharp bases that fine to massive mud indicate relatively rapid deposition and are indicative of turbidite deposits (Figure 5.8). Bioturbated muds with persistent IBRD between laminae indicate periods between numerous turbidity flows over long periods of time (Figure 5.8). Large broad sand peaks within the sand beds demonstrate that sands larger than 250 μm may potentially be deposited by down-slope processes (Figure 5.8),

but grain sizes larger than 250 μm are unlikely to represent turbidite sands in any diatom-rich/bearing or massive mud samples.

Figure 5.8 A transect of grain size distributions across a thick sand turbidite bed indicating these sand beds are distinctive from the surrounding sediment. Grain size distribution of a turbidite bed characterised by very coarse silt to fine sand.

5.3.2 Down-core grain size trends

A summary of grain size results is shown in figure 5.9 with IBRD MAR and facies. All samples were poorly to very poorly sorted ($2 - 4\sigma$; Figure 5.9), except for one moderately sorted fine sand bed. Mean grain size ranges from 1.6 to 98 μm with an average of fine to medium silt (9 μm). Grain size data indicate there are no overall trends in sediment size or texture within this section of the core.

Figure 5.9 Stratigraphic log with lithology, facies, lithostratigraphic cycles, and degree of bioturbation with IBRD and summary grain size data.

5.4 Iceberg rafted debris

Iceberg rafted debris mass accumulation rates (IBRD MAR) were calculated following Patterson *et al.* (2014) using the sieved coarse-sand (250 μm –2 mm) weight percent of each sample. Once biogenic material was removed, analogous IBRD peaks were inspected visually in core photos to ensure the coarse sand was not a result of a thick sand bed. Only two peaks were possibly associated with a sand bed, and these samples were removed from the IBRD data set, as it is assumed turbidite deposits are deposited instantaneously and contain no real IBRD signal (Figure 5.10a). High sand percent (>10%) from grain size data were also checked in core photos for the same reason, and only one sample corresponded to a potential turbidite (Figure 5.10b). In contrast, real IBRD peaks are distinctive from background sedimentation (Figure 5.10c), and were often in proximity to larger >2 mm IBRD clasts, confirming an iceberg origin of the finer IBRD sediment.

Figure 5.10 Core photographs of high sand percent as a consequence of sand beds, sandy laminae, and real IBRD events. **a.** Core 16H 1 25–27, false IBRD peak associated with a sand bed. **b.** Core 13H 4 65– 67 false IBRD peak associated with sand-rich laminae. **c.** Core 12H 5 showing real IBRD in bioturbated diatom-rich mud.

After the removal of data points biased by non-IBRD grains, this IBRD record contains 420 samples (Figure 5.11; Appendix A). Coarse sand percent (cs%) did not exceed 10.1% and averaged much lower at 0.7%. IBRD MAR ranged from 0 to 0.241 g/cm²/kyr, averaging 0.018 g/cm²/kyr, with only 11 samples containing no IBRD. Generally, IBRD is clustered in a slightly skewed normal distribution, increasing up core. IBRD MAR increases above ~126

m with thicker packages of IBRD compared to the frequent, smaller events below, increasing from an average of 0.013 g/cm²/kyr to 0.025 g/cm²/kyr. Visual observations of the IBRD trends are highly cyclic, with a persistent low frequency ~4 m wavelength cycles throughout the record, which is especially evident in the bottom 10 m and top 20 m. Between ~123 m and 142 m, the long-scale cycles are overprinted by smaller ~1 m cycles. This is explored quantitatively in the spectral analysis section of the results. IBRD is higher within diatom-rich facies, although in general peaks of IBRD occur at transitions of facies.

Previous studies have used a sorting parameter, as defined by Folk & Ward (1957), of the terrigenous fraction in order to differentiate IBRD and enrichments in coarse sand caused by current winnowing fine material (Passchier 2011; Patterson *et al.* 2014). It has been suggested that a peak in IBRD coinciding with well-sorted material is likely to be condensed sections of coarse material caused by high current activity, rather than delivery of coarse sand by IBRD (Passchier 2011); whereas IBRD correlating with poorly to very poorly sorted material reflect actual IBRD events superimposed on background hemipelagic sedimentation. This parameter may also identify potential hiatuses between magnetic tie points related to enhanced current winnowing (Patterson 2014). Only one sample was moderately sorted, with the rest of the samples poorly to very poorly sorted. The moderately sorted sample was associated with the bottom of a turbidite, and for the same reason, samples with high (>10%) sand were visually checked to ensure they were real IBRD peaks and not coarse material from turbidites.

Figure 5.11 Stratigraphic log with IBRD, sorting and on-board estimations of gravel (clasts >2mm) content. IBRD that occurs in poor to very poorly sediment argues for no current winnowing and therefore a true signal. IBRD and gravel estimations show relatively similar patterns, where increased IBRD occur in regions with high gravel content.

5.4.1 Comparison of iceberg rafted debris mass accumulation rate methodologies

5.4.1.1 Sieved iceberg rafted debris versus on-board measurements

Figure 5.12 compares the sieved IBRD record (250 μm –2 mm and >2 mm) to the average clast abundance over 10 cm as described by the initial shipboard science team. A noted day/night cyclicity was presented in the shipboard counting of clasts in the core face (as clasts were often smeared over by soft sediment during core splitting), where one shift consistently observed lower clast numbers (McKay, pers. comm). Despite this issue, sieved >2 mm gravel and clast abundance show roughly consistent patterns, except clast abundance is higher in places (Figure 5.12a). IBRD (250 μm –2 mm) and gravel clasts show broadly consistent patterns (Figure 5.12b). Comparison of this record with other established records gives confidence that despite the coarser sampling resolution, the observed patterns of IBRD are consistent.

Figure 5.12 Sieved IBRD (250 μm –2 mm and >2 mm) records compared to shipboard count of clast abundance of core face. **a.** Sieved >2 mm IBRD gravel and clast abundance. **b.** 250 μm –2 mm IBRD and clast abundance.

5.4.1.2 Sieved iceberg rafted debris versus volume percent estimates

An alternative method of IBRD calculation has been proposed using bulk particle size analysis (such as LPSA) to estimate the volume percent (v%) of coarse sand (Passchier 2011), and one of these records overlaps with the same time period of this study at nearby site U1359. Site U1359 is situated on the same levee, but at shallower water depths, and where the channel levee system is higher relief (Hansen *et al.* 2015). LPSA estimates of >250 μm compared to the sieved weight (wt%) of IBRD showed no correlation ($R^2 = 0.0002$; Figure 5.13a), and the volume percent underestimated IBRD compared to the sieved weight (0.23% vs 0.79%). This may be related to the amount of sample used for

LPSA (<1 g) compared to sieved weights (8–12 g), resulting in a higher margin of error with LPSA whereby a single large sand grain can bias a volume percent calculation, and thus is not statistically significant in a sample dominated by mud grain sized.

Volume percent of sand over 125 μm has been previously used as an IBRD proxy (e.g. Passchier 2011), however in this U1361 IBRD record, the wt% of 250 μm –2 mm IBRD and v% of >125 μm from LPSA estimates are poorly correlated ($R^2 = 0.002$; Figure 5.13b). Down-core comparisons of the three IBRD records of U1361 shows there is no consistent pattern with what >250 μm wt% >250 μm v% peaks it mimics, i.e. not all big >250 μm wt% peaks are captured in the LPSA data (Figure 5.14).

Figure 5.13 Cross-correlation of sieved 250 μm –2 mm IBRD and LPSA v% estimations. **a.** >250 v% and 250 μm –2 mm wt% show a poor correlation ($R^2 = 0.0002$). **b.** >125 μm v% and 250 μm –2 mm wt% shows a poor correlation ($R^2 = 0.002$).

Figure 5.14 Down-core measurements of weight percent (wt%) and volume (v%) estimations of IBRD. In general, there is poor agreement between the two methods.

Similar comparisons between methodology to Albot (2016) were produced for U1361 in this record. Raw v% and wt% of the same samples saw a moderate relationship between >250 μm fraction of both measurements (Figure 5.14), however the actual values of the volume % were 0 in 54% of samples (compared to 3% of sieved samples), or much higher (average 0.2% vs 0.007%). The relationship between >125 μm LPSA and 250 μm –2 mm

sieved data is poor, with the LPSA measuring far higher estimates of IBRD than the sieved method (1.96% vs 0.007%), highlighting the likelihood of >125 µm LPSA capturing the coarse silty sand of down-slope sedimentary processes rather than IBRD deposition. Grain size measurements through LPSA found relatively poor reproducibility in samples with high IBRD (Figure 4.7). Following this, Albot (2016) also concluded that it is difficult to make comparisons between methodologies of estimating IBRD, so as this record is a continuation of Patterson *et al.* (2014), the sieve analysis was used to be consistent.

5.5 X-Ray Fluorescence

5.5.1 Single-element trends

All elements show dm to m-scale cyclicity at variable amplitudes throughout, with high amplitude variations occurring at around ~120 mbsf (Figure 5.15). Metals show roughly similar trends with peaks at the same depth. Fe and K are co-varying. Al shows low amplitude variability with cyclicity of ~5 m wavelength. Ba is highly cyclic, with the bottom 10 m showing broad ~4 m highs, increasing to ~1 m frequency between ~140 m and 124 m. The upper section of the core shows broad 2 m cyclicity, with long sustained highs, especially in the top 8 m. In comparison, Si has no obvious trends except for low amplitude, low frequency cyclicity. Calcium (Ca) content is very low (<0.6%) throughout the record, consistent with shipboard measurements of near-zero CaCO₃ values (Escutia *et al.* 2011).

Figure 5.16 Single element downcore trends of key terrigenous, authigenic, and (potentially) biogenic elements (Al, Ti, Fe, K, Mn, Ba, Si, Ca).

5.5.2 Cross-correlation matrix

Once calibrated to ppm, selected data were examined for affinity between elements using a correlation matrix. Correlation matrix calculations show how closely related elements are to one another (Table 5.2). Single element correlations show moderate-to-high (0.52–0.80) correlations between ‘terrestrial’ elements (Al, Fe, potassium (K), rubidium (Rb)). The only strongly negative correlation is between Si and Al (-0.65). No correlation between iron (Fe) and titanium (Ti) could indicate diagenetic alteration of Fe (-0.025; Croudace & Rothwell 2015).

Table 5.2 Correlation matrix for element counts. Strong positive correlations (>0.6) are highlighted in orange. Strong negative correlations (>0.6) are highlighted in blue.

	Mg	Al	Si	Fe	Ca	Ba	Mn	K	Ti	Sr	Rb	Zr
Mg	1.0											
Al	0.66	1.0										
Si	-0.47	-0.65	1.0									
Fe	0.76	0.63	-0.36	1.0								
Ca	-0.16	-0.11	0.17	-0.12	1.0							
Ba	-0.01	-0.2	0.41	-0.06	0.2	1.0						
Mn	-0.02	0.05	-0.11	-0.01	0.14	-0.09	1.00					
K	0.64	0.76	-0.49	0.8	-0.14	-0.24	0.05	1.0				
Ti	0.02	-0.01	0.26	-0.03	0.2	0.72	-0.09	-0.19	1.0			
Sr	0.1	-0.08	0.52	0.1	0.28	0.27	-0.04	-0.01	0.17	1.0		
Rb	0.62	0.78	-0.59	0.75	-0.26	-0.37	0.08	0.88	-0.24	-0.19	1.0	
Zr	-0.36	-0.25	0.32	-0.38	0.15	0.23	-0.09	-0.31	0.18	0.3	-0.4	1.0

5.5.3 Down-core ratio trends

Elemental ratios are often used in preference to concentration as a tool for normalizing elemental abundance to a common factor. For example, Ba is hosted in lithogenic minerals and forms as barite (BaSO_4) in proportion to plankton productivity. Assuming the Ba/Al in lithogenic minerals is fixed, down-core Ba/Al therefore shows intervals of ‘excess’ Ba, attributable to enhanced productivity. In this case, the down-core Ba/Al pattern is similar to the Ba concentration curves, with high cyclicity and fourteen intervals of ‘high productivity’, designated as values above 0.019 (see section 5.5.5 below for details). Below 140 m, these intervals are thick (>2.5 m), compared to between 125 m and 140 m where the cycles are <1 m. Si/Al, a potential indicator of ‘excess’ biogenic silica, loosely co-varies with Ba/Al, with peaks in Si/Al coinciding with highly productive intervals, but the ratio is noisier and occurs as ~5 m wavelength cycles. Between 142 and 120 mbsf, the

cycles in all ratios are shorter in wavelength and appear superimposed on longer duration cycles of ~3 m. Ca/Al, a potential indicator of biogenic carbonate, is relatively constant throughout the core (Fig. 5.16).

K/Ti also show long wavelength cycles, but these are anti-phased with Si/Al especially, e.g. where Ki/Al is high, K/Ti ratios are low. Similarly, eight ~5 m cycles can be seen in the dataset, with four ~2 m cycles between 132 m and 120 m. Highly productive intervals correspond to shifts in K/Ti. Al/Ti ratios are often used as indicators of provenance of lithogenic grains, since both elements have similar weathering characteristics and are not incorporated in biogenic minerals. Al/Ti is highly cyclic ranging from 8 to 20, with well-defined, short duration cycles between 143 and 120 m. Al/Ti closely matches K/Ti and is anti-phased with zirconium/rubidium (Zr/Rb). The range and cyclic manner of Al/Ti suggests a change in provenance of terrigenous material and/or systematic variations in grain size throughout the core (Fig. 5.16).

5.5.4 X-Ray Fluorescence and facies

Differences in elemental composition between facies can be observed in figure 5.16. Sustained peaks of Ba/Al occur within diatom-rich intervals where the average ratio is much higher than found in massive and laminated muds (0.018 vs 0.009 respectively). Diatom-rich facies also have a higher Si/Al ratio (6.8 vs 5.5), which is consistent with Si/Al being used as a proxy for diatom content. Both K/Ti and Al/Ti are lower within diatom-rich facies, where large amplitude shifts in the ratios occur at the transition between mud and diatom-rich facies. A similar pattern is observed in Zr/Rb where the ratio is higher within diatom-rich facies, likely reflecting an increase of coarse material (see section 5.5.6). No consistent trends or patterns are observed in Ca/Al.

Figure 5.1 Lithologic log and XRF ratios presented in this thesis. All ratios except Ca/Al show cyclic, co-varying relationships related to facies. Grey shading denotes 'highly productive' intervals as defined by accumulations of Ba/Al above 0.019.

5.5.5 Barium and a proxy for primary productivity

Ba/Al ratios show a strongly bimodal distribution (Figure 5.17), with a larger mode in the lower accumulations between 0.005 and 0.015, which possibly corresponds to the terrigenous portion of the ratio and low accumulations of biogenic barium. There is a second, smaller and broad normally distributed second mode between 0.015 and 0.025. The shift in accumulations at 0.019 is interpreted in this thesis to represent sediments that contain increased marine productivity and is distinct from low marine productivity or terrigenous barium.

Figure 5.17 Histogram of Ba/Al displaying a bimodal distribution. Ratios exceeding 0.019 are considered to represent enhanced marine productivity.

5.5.6 X-Ray Fluorescence and grain size

NGR is the amount of natural gamma-rays emitted from the core and is related to grain size due to the concentration of radiogenic isotopes (^{40}K , ^{238}U , and ^{232}Th) within clay minerals (Blum 1997). This allows estimation of grain size independent of the LPSA grain size measurements. XRF measurements of Zr/Rb can also reflect grain size as Zr is preferentially concentrated in coarser sediment, while Rb is concentrated within the fine fraction (Wang *et al.* 2011; Rothwell & Croudace 2015; Penkrot *et al.* 2018). Trends of NGR and Zr/Rb are in agreement, where peaks in both measurements indicate a higher percentage of clay within the sediment (Figure 5.18). The Zr/Rb record provides an independent, high-resolution grain size record of the core, and is used as a geochemical proxy for grain size trends.

Figure 5.2 NGR and Zr/Rb show consistent trends, where higher values in the measurements indicate higher clay percent.

5.6 Spectral analysis

Spectral analysis was performed on IBRD MAR, Ba/Al, benthic $\delta^{18}\text{O}$ records, and eccentricity tilt precession (ETP) insolation datasets to reveal significant frequencies associated with orbital pacing within the time series. The results of MTM power spectrum for all datasets indicate frequencies associated with either 100 kyr eccentricity, 40 kyr obliquity, or 20 kyr precession, but with varying degrees of timing, strength, or confidence. Spectral analysis was run on the full length of all the datasets, despite the missing core section between 111.26 and 113.56 mbsf which requires interpolation, and introduces false low frequency signal within the spectral analysis. However, EHA requires a minimum length to evaluate orbital signals, and thus a longer record is preferred. EHA results show persistent, low-frequency cycles in all datasets, and despite the record showing intervals of significant orbital pacing, some intervals are noisy, and this may be associated with variations in sedimentation rates or unidentified hiatuses in the core (Meyers & Hinnov 2010).

5.6.1 Iceberg rafted debris

Three double-peaked frequencies cross the 90% confidence interval in MTM spectral analysis results which correspond to ~ 100 kyr, ~ 23 kyr, and ~ 19 kyr (Figure 5.19). The alignment of these frequencies in this untuned data suggests IBRD is directly influenced by eccentricity-modulation of precession cycles. In addition, 40 kyr obliquity is the next most significant frequency but is only significant at 80%. The EHA results show a similar low-frequency band throughout the record which sharpens at ~ 5500 ka. Prior to 5500 ka, frequencies between 0 and 0.03 ($> \sim 35$ ka) show low-power significance with sections of strong obliquity ($1/40 = 0.025$) at 5700 ka. At 5500 ka, a truncation of the signal may potentially correspond to a hiatus in the record (Fig. 5.20). Above 5500 ka eccentricity

($1/100 = 0.01$) is the most significant frequency, although minor obliquity and precession forcing is observed. It is important to note the EHA assigns more power to the lower frequencies where there is a red-noise component to the dataset, which is common in paleoclimate datasets. The MTM analysis models this background red-noise component, and can identify the significance of the power at lower frequency (Meyers 2012). The comparison of the EHA results (Figure 5.20) with the MTM analysis (Figure 5.19) indicates eccentricity and precession have equal power throughout the length of the dataset, further supporting eccentricity modulation of precession cycles in the IBRD data. The gap at 4900 ka is associated with the core gap and is not a real signal.

Figure 5.19 MTM results of the complete IBRD record highlights eccentricity and precession at, or above, 90% confidence level. Obliquity pacing is identified but is only significant at 80%. AR1 confidence level (red line) is assigned by the red-noise model; values over this value are statistically significant.

Figure 5.20 EHA of IBRD showing the relative power of frequencies of time series where red = stronger power. Eccentricity is in the frequency band $\sim 0.08\text{--}0.01$, obliquity ~ 0.025 , precession 0.052 and 0.043.

5.6.2 Miocene versus Pliocene iceberg rafted debris

The record of IBRD in the depth domain show apparent changes in accumulation rates between the Miocene to the Pliocene (Figure 5.11; ~ 125 m depth). This was confirmed by whole record spectral analysis (Figure 5.20), which identified eccentricity, obliquity, and precession in the older section of the record, but switched to a clearer signal at ~ 5500 ka, but only displayed strong eccentricity forcing. To test potential differences in orbital pacing between the two time periods indicated in the EHA results, two sections of the record were run separately within the Miocene between 5.3 and 6.2 Ma, and Pliocene between 4.2 and 5.3 Ma (Figure 5.21). MTM results from the Miocene show no significant signal in eccentricity frequency, but does show a stronger, albeit broad, 40 kyr frequency. In contrast, MTM results for the Pliocene shows a clear change and coherence in pacing

of IBRD with only eccentricity and precession bands above the significance levels. It also shows 70 and 30 kyr frequency bands, which are typical combination tones of 100 and 41 kyr frequencies in untuned data (Huybers & Wunsch 2004).

Figure 5.21 MTM spectral analysis results of **a.** Early Pliocene and **b.** Late Miocene IBRD records run separately highlights the drop-off of obliquity within the Early Pliocene, replaced by strong eccentricity and precession. AR1 confidence level (red line) is assigned by the red-noise model; values over this value are statistically significant.

5.6.3 Inorganic geochemical spectral analysis

Several frequencies are significant in MTM analysis of the Ba/Al record (Figure 5.22), but the orbital signal is not as clear as the IBRD record due to a large red-noise component centred on a period of 200 kyrs ($0.005 \text{ cycles kyr}^{-1}$). This is similar to observations of Patterson *et al.* (2014) for the Late Pliocene to Pleistocene of the core, with only two frequencies reaching 95% (200 kyr and 19 kyr), but frequencies of 40 kyr and 23 kyr are above >80% (Figure 5.22), and may increase in strength if the red-noise component was removed. EHA results show the 200 kyr frequency is persistent throughout most of the record, although shifts into the 100 kyr eccentricity band between 5500 and 5000 kyr (Figure 5.23).

Figure 5.22 MTM and EHA spectral analysis results of Ba/Al. AR1 confidence level (red line) is assigned by the red-noise model; values over this value are statistically significant.

Figure 5.23 EHA spectral analysis results of Ba/Al.

Similar results are identified in EHA of K/Ti, Al/Ti, and Zr/Rb (Figure 5.24). All results show persistent 200 kyr cyclicity except between 5600 ka and 5000 ka, where the significant power is centred in eccentricity and obliquity between 5700 and 5200 ka in K/Ti and Al/Ti, and up to 4800 ka in Zr/Rb.

Figure 5.24 EHA spectral analysis results of **a.** K/Ti, **b.** Al/Ti, and **c.** Zr/Rb. All records contain ~100 kyr eccentricity, with secondary obliquity between ~5.6 Ma and ~5 Ma.

CHAPTER 6: DISCUSSION

The objective of this chapter is to integrate the U1361 datasets to understand local changes in sediment delivery processes relating to ice sheet dynamics on the continental shelf and productivity in surface waters at the Wilkes Land margin. Comparisons are then made to ‘far-field’ proxies of ice volume and temperature changes, in particular, the deep-sea foraminifera $\delta^{18}\text{O}$ record. The results are discussed in the context of existing hypotheses used to explain shifts in the orbital pacing of ice sheet variability over the past ~6 Ma.

6.1 Controls on sediment deposition

6.1.1 Facies analysis

Sixteen sedimentary cycles were identified between 6.4 and 4.4 Ma, with each cycle defined as a single package of the two primary facies of massive/laminated mud passing up into diatom-rich (/bearing) sediment (see section 5.2) (Figure 5.11). In general, each facies package follows a consistent pattern of diatom content increasing up-section, which may indicate a gradual, and in some cases, step-wised increase of biological activity, or alternatively a decrease in terrigenous input.

The base of each lithostratigraphic cycle is defined by massive muds that contain intervals of laminae and silt beds, and massive internals with occasional evidence of relict bedding or mottling texture indicative of enhanced bioturbation. This facies contains relatively low IBRD and low biogenic content (low Ba/Al, Si/Al, and higher grain density). The graded nature of the silt beds with sharp bases are consistent with the deposition of non-erosive spill-over of low-density turbidity currents (Stow & Piper 1984; Lucchi *et al.* 2002; Caburlotto *et al.* 2010). Thicker silt beds are generally overlain by progressively thinner laminae, and then bioturbated/massive muds, which is consistent with Stow & Piper (1984) mud turbidite facies scheme ‘T3’ to ‘T7’ (Figure 2.7). Such ‘base-cut-out’ turbidite beds are consistent with deposition occurring on a deep-sea levee system, whereby fine-grained over-flow reaches the crest of the levee and settles, but the erosion and deposition of coarser grained material is restricted to the channel. This facies is interpreted to be related to glacial advance, whereby sediment is delivered at an ice sheet grounding line located near the continental shelf edge. Turbidity currents result from

sediment loading and slope failures at the shelf edge, although slope failure and turbidity currents could also be triggered by isostatic rebound during glacial retreat, or from cascading AABW waters (Escutia *et al.* 2011; Patterson *et al.* 2014).

The top of each cycle is characterised by pervasively bioturbated diatom-rich mud intervals with high IBRD and biogenic silica content. Any primary sedimentary structures originally present have been destroyed by bioturbation. These features are consistent with glacial minima with warmer surface oceans promoting enhanced primary productivity. A lack of laminae combined with increased bioturbation indicates a reduction or lull in turbidite activity as the ice sheet retreats. However, as noted above, isostatic rebound and AABW formation processes could still result in some turbidite activity. There is no evidence for significant winnowing during these intervals, and there is no significant shift in grain size between facies — although, there is a slight decrease in clay content (Figures 5.6; Rebesco *et al.* 2014). IBRD peaks generally occur at the transition between facies which may reflect accelerating calving at both the Wilkes Land margin and EAIS outlet glaciers from the Ross Sea as the ice sheet undergoes rapid retreat.

The sediment facies pattern discussed above in the Late Miocene to Early Pliocene interval of U1361 conforms with the Patterson (2014) model described for the Late Pliocene to Early Pleistocene. It also suggests the same environmental drivers on sedimentation and glacial dynamics, as discussed above are similar and allows for a direct comparison to that later record.

6.1.2 Provenance of terrigenous material

Down-core patterns of Al/Ti within this interval of U1361 show cyclic variability (ratios of between 8 and 20) that coincides with facies shifts (Figure 6.1). This may indicate a change in the composition of terrigenous material feeding the site between glacial and interglacial periods. The large variability in this ratio adds complexities to interpreting geochemical data because a normalising factor must be used to remove terrigenous signals within the data (Calvert & Pedersen 2007). As a result, variations within proxy records such as biological productivity could be interpreted as a consequence of another unrelated process — such as a shift in grain size or terrigenous sediment source.

Therefore, it is important to understand the drivers of the variation in terrigenous material to identify real signals in the data. A systematic decrease in Al/Ti may be a consequence of two different processes — an increased contribution of mafic minerals within the source material, or an increase in grain size as minerals containing Ti are commonly concentrated in coarser fractions (Calvert & Pedersen 2007).

Figure 6.1 Lithologic log with Al/Ti and Zr/Rb shows increases in Al/Ti in glacials corresponds to an increase in Zr/Rb which implies a coarsening of terrigenous sediment.

The relationship between Al/Ti and mean grain size from the LSPA data is poor (Figure 6.2a, $R^2 = 0.017$). However, relationships between other geochemical ratios commonly associated with grain size (e.g. Zr/Rb) and Al/Ti show a stronger relationship (Figure 6.2b, $R^2 = 0.26$; Figure 6.2). Low values of Zr/Rb coincides with diatom-rich facies, whereas Zr/Rb decreases in mud facies which implies an increase in grain size in mud facies as Rb is more prevalent in clays (Wang et al. 2011). This is consistent with grain size data of a reduction in clay content in diatom-rich facies (section 5.3). Consequently, the correlation between Al/Ti and Zr/Rb implies there is a grain size component to the Al/Ti record. However, the large residual in the linear relationship suggest it is probably a minor contributor to the variability of the ratio and that shifts in provenance within the terrigenous sediment also influence Al/Ti.

Figure 6.2a Cross-correlation plot of Al/Ti and LPSA mean grain size shows no relationship ($R^2 = 0.017$). **b.** Al/Ti vs Zr/Rb shows a moderate relationship ($R^2 = 0.26$), suggesting some of the Al/Ti variability is a consequence of grain size.

The relationship between the facies cycles in the Al/Ti data is also seen in K/Ti (Figure 6.3). K/Ti has previously been used as a provenance indicator in Wilkes Land where low K/Ti indicates a more basaltic source such as FLIP or Beacon Supergroup (Bertram *et al.* 2018). The Al/Ti from this record compared against local and Ross Sea bedrock lithologies does not show a clear association with any one source (Figure 6.3a), although these comparisons are limited by the availability of geological data and the likelihood of mixed geochemical signals in marine sediments. It is likely, given the complex geology of the region and the complexities of this dataset, that it is not possible to trace the source of the terrigenous material with these data alone. However, the distinction between sources (e.g. mafic versus felsic lithologies) in K/Ti can still be seen as shifts within this ratio with changes in the lithofacies (Figure 6.3b) Terrigenous material in diatom-rich muds display a more mafic composition, which is consistent with the findings of previous provenance studies that identified Ferrar-sourced sediment from the inner regions of the WSB as the ice sheet retreated during interglacials (Cook *et al.* 2013; Bertram *et al.* 2018). Glacial sediments are more felsic in composition and are likely sourced from local Paleozoic granitoids (e.g. Bowers Terrane) that outcrop at the coastal margin of the modern ice sheet at the Wilkes Land margin (Figure 6.4).

Further isotope work (e.g. Nd/Sr), as undertaken by Cook *et al.* (2013) and Bertram *et al.* (2018), could better-constrain provenance of terrigenous material. These XRF data provide a useful tool for identifying where the largest provenance shifts occur, and may assist in a sampling plan for more time-intensive and expensive isotope analysis. An interval of interest that is largely unanalysed for isotope work is the apparent increase in

mafic sources within interglacial facies below 120 mbsf, which suggest a major change in sediment erosion sources or delivery near the Miocene–Pliocene boundary (Figure 6.3).

Figure 6.3a Downcore Al/Ti ratios and Al/Ti ratios of potential local and Ross Sea source rocks. **b.** Downcore K/Ti ratios and K/Ti ratios of potential local and Ross Sea source rocks. Dolerites denoted on Al/Ti graph have similar K/Ti ratios as Ferrar diabase (~2) so have not been plotted. Granulite does not plot on the axis (0.8). There is a distinction in both Al/Ti and K/Ti ratios between diatom rich and mud facies, but source lithologies do not provide a unique solution regarding a Wilkes Land vs Ross Sea influence. Geological data from Roser & Pyne (1989), Henjes-Kunst & Schüssle (2003), and Goodge & Fanning (2010).

Figure 6.4 Geological map of Wilkes Land and Ross Sea. The proximity of the core site to a variety of bedrock lithologies and major sediment transport pathways makes identification of source region with XRF data alone difficult. Grey arrows indicate approximate modern-day transect of icebergs (Tournadre *et al.* 2016). Modified from Cook *et al.* (2017).

6.1.3 Evidence for biological productivity

Proxies of marine primary productivity, such as Ba/Al and Si/Al, can highlight intervals of enhanced biological productivity. Despite the uncertainties regarding the links between Ba/Al and primary productivity (section 2.6.6), studies from the Wilkes land margin have related elevated Ba/Al and Si/Al to other proxies of enhanced biological productivity during interglacials (e.g. sediment density (section 5.2.2), biogenic opal and diatom valve concentration; Cook *et al.* 2013; Hansen & Passchier 2017; Khim *et al.* 2017), suggesting this proxy is valid at this site. As outlined above, there are issues with isolating contributions from terrigenous and biogenic components due to varying terrigenous sediment source changes affecting the trend in aluminium. It is acknowledged that this signal cannot be removed completely, however, consistencies between independent datasets imply biogenic silica primary productivity changes are recorded at this site.

In particular, the covariance between an estimate of diatom content in sedimentary facies, biogenic silica estimates (developed in section 5.2.2), low-resolution opal measurements (Figure 5.4), Ba/Al, and Si/Al strongly suggests these records are a true biological productivity signal, and intervals with high (>0.019) Ba/Al are indicative of enhanced marine productivity (Figure 6.5). This indicates the signal above background levels (e.g. Si/Al above >7 ; Figure 6.5) is a reliable indicator of biogenic Si dominating over Si supplied from terrigenous sources. The relationship between increased Si/Al and Ba/Al is consistent with siliceous diatom plankton making up the dominant proportion of microfossils within U1361 (Escutia *et al.* 2011; Cook *et al.* 2013; Patterson *et al.* 2014). CaCO_3 content between 152 to 104 m is negligible, and Sr/Ca ratios are consistently well-above those expected for nannofossils (<5 ; Appendix B), indicating the Ca content within the core is entirely terrigenous. This is consistent with shipboard descriptions and measurements of CaCO_3 in this interval (Escutia *et al.* 2011).

Figure 6.5 Comparison of Ba/Al and Si/Al with an estimated biogenic silica% from GRA measurements. Intervals of high Ba/Al and Si/Al, in general, coincide with areas of high BSi.

In the Southern Ocean, the zone of maximum primary productivity occurs south of the APF (Pondaven *et al.* 2000; Nelson *et al.* 2002), and corresponds to the region of maximum supply of dissolved nutrients by upwelling of nutrient-rich deep water masses (Trull *et al.* 2001). The supply of these nutrients (e.g. iron, nitrate, phosphate, silica) are a limiting factor on phytoplankton blooms (Bathmann *et al.* 1997; Hoppe *et al.* 2017), and changes in production at millennial scale are related to the intensity of upwelling of deep, nutrient-rich waters (Anderson *et al.* 2009; Tang *et al.* 2016; Billups *et al.* 2018). Poleward shifts in the strength and position of the westerly winds lead to latitudinal variation in the

invigoration of the APF (Toggweiler *et al.* 2006) — intensifying upwelling, and in turn increasing primary productivity (Bonn *et al.* 1998; Jaccard *et al.* 2013; Wu *et al.* 2017; Billups *et al.* 2018). This process may be occurring at U1361 during the Late Miocene to Early Pliocene, where diatom-rich intervals during interglacials are a consequence of enhanced upwelling south of the APF over the core site, whereby warmer waters and decreased sea ice drives the increase in primary productivity by increasing light availability. Enhanced deposition of biogenic opal is seen in other records south of the APF during interglacials (Falkowski *et al.* 1998; Billups *et al.* 2013). A decrease in marine productivity during glacials are a result of decreased nutrient supply due to weakened upwelling (Horn *et al.* 2011; Studer *et al.* 2015). This reduces the amount of heat in surface waters, which enhances sea ice formation. Interestingly, nutrient-rich deep waters upwelling during the last glacial termination also brought high concentrations of dissolved carbon to the surface ocean that was subsequently vented into the atmosphere (e.g. Anderson *et al.* 2009). Amplification of the 100 kyr eccentricity cycle through exchange of carbon between ocean and atmosphere is one mechanism proposed to account for the high amplitude of climate change at this frequency in spite of low insolation forcing (Imbrie *et al.* 1992; Shackleton 2000). Whilst not conclusive in the absence of highly resolved and precise atmospheric CO₂ measurements, the 100 kyr cyclicity in diatom production seen here is consistent with ocean–atmosphere exchange at this period. It therefore remains possible the Southern Ocean may have played a key role in Earth’s climate system since at least the Miocene.

6.1.4 Iceberg rafted debris as a proxy for ice dynamics

The accumulation of IBRD at site U1361 has previously been interpreted as reflecting the flux of ice discharge from the EAIS from the Wilkes Land margin and the Ross Sea (section 5.4; Patterson *et al.* 2013; Wilson *et al.* 2018). The consistency of the sedimentation model in this studied interval, along with these earlier studies, suggests the Late Miocene to Early Pliocene IBRD record is also a measure of changes in iceberg flux. Down-core trends of IBRD between 6.2 and 4.4 Ma displays three distinct stages which are closely linked to the duration of sedimentary cycles (Figure 6.6). That is, when IBRD variance occurs as high frequency events, the facies are also highly cyclic and of short duration. This relationship between facies and IBRD suggests the response of the whole sedimentary and biological

system to the climatic state is directly tied to changes in ice dynamics of the ice sheet on the continental margin. Stage 1 (6.2–6.03 Ma) contains high frequency events superimposed on two broad cycles of accumulation within prolonged interglacials. Stage 2 (6–5.2 Ma) contains high frequency, low accumulation cycles with peaks concentrated within interglacials. A baseline shift of IBRD occurs in stage 3 (5.2–4.4 Ma), with an increase in IBRD MAR even in glacial periods. Testing different modelled sedimentation rates confirm these trends are not a consequence of changing sedimentation rates at paleomagnetic tie points, which can often exaggerate the contrast in mass accumulation rate values between periods of short duration paleomagnetic chrons (Figure 6.7). IBRD peaks occur at the transition between facies, which is consistent with an increased flux of ice during deglaciations. Pulses of IBRD at ~5.2, ~4.9, and 4.6 Ma occur during intervals identified by Cook *et al.* (2013) when the ice sheet retreated into the WSB central basin.

Figure 6.6 Lithologic log and IBRD MAR against paleomagnetic tie points. IBRD accumulations occur broadly in three stages.

Figure 6.7 IBRD MAR as calculated with LSR from the age model (blue) and with a constant LSR derived from the upper and lowermost tie points only (orange). Although there are areas where mass accumulation rates are different, this confirms trends within the data are not a consequence of changing sedimentation rates at paleomagnetic tie points.

6.2 Assessing leads and lags between ice sheet dynamics and Southern Ocean response

6.2.1 Integrated depositional model

Measurements of different sediment components on the same sample allow the relative timing and relationships between ice sheet dynamics and biological productivity to be assessed independent of age model uncertainty. Most sedimentary cycles from U1361 contain a distinct pattern within the IBRD record at glacial to interglacial transitions. There is a gradual initial increase of IBRD that occurs at the end of a glacial period just below the facies transition from massive/laminated muds into diatom-bearing muds (Figure 6.8). This is followed by a second, often larger, peak at the transition from diatom-bearing to diatom-rich sediments, indicating an intensification of ice-rafting levels leading into peak interglacial conditions (Figure 6.8). The increase in IBRD precedes maxima in primary productivity indicators (Ba/Al, biogenic opal), which indicates deglaciation of the ice margin occurs prior to the primary response of the biological system. The systematic pattern of facies in each sedimentary cycle may indicate the diatom-bearing facies is a transitional facies between glacial and full-interglacial conditions (Figure 6.8).

Figure 6.8 The pattern of deglaciation within the IBRD record shows a gradual increase of IBRD from mud to diatom-bearing facies, and the largest IBRD accumulation at the boundary of diatom-bearing to diatom-rich facies.

The poleward shift in southern hemisphere westerly and polar easterly wind fields, which enhances upwelling at the Antarctic divergence and promotes warm water flux on the continental shelf, is inferred to be the primary process driving past marine-based ice sheet retreat (Naish *et al.* 2009; Hillenbrand *et al.* 2017), but is also an important process in promoting increased marine productivity (Falkowski *et al.* 1998; Billups *et al.* 2013). The largest flux of IBRD occurs within the interval of highest diatom/biogenic opal content (e.g. diatom-rich muds), but then IBRD decreases while diatom content remains high

(Figure 6.8; 6.9). This is interpreted as representing the continuing enhancement of upwelling, resulting from a more southern westerly wind field as the ice margin retreated. The resultant ice sheet melt increases light availability as warmer surface water associated with CDW is upwelling during active phases of ice sheet melting can act to restrict sea ice growth (Figure 6.10b; Jourdain *et al.* 2017).

The repeated advance and retreat of the ice margin, as indicated by facies cycles and IBRD, also influences K/Ti ratios, indicating a probable switch between two lithological end members as the ice sheet oscillates (section 6.1.2). The largest shifts in sediment provenance occur rapidly at glacial-to-interglacial transitions, coinciding with the onset of enhanced marine productivity and the onset of peak IBRD accumulation (Figure 6.9). The pattern of K/Ti displays a ‘see-saw’ pattern over glacial-to-interglacial cycles throughout the record, where the ratio gradually increases throughout a glacial period, until sharply decreasing (towards mafic values) at the transition into the interglacial facies (Figure 6.9b). During interglacials, the ratio remains low until an abrupt shift at the beginning of a glacial. This pattern is more clear at mud to diatom-rich transitions.

Figure 6.9 The pattern of K/Ti, IBRD, and highly productive intervals. IBRD leads the biological response and provenance shifts during deglaciations, but Ba/Al and K/Ti remains stable while IBRD decreases during the interglacial.

This pattern may represent the slow growth of an ice sheet across the Wilkes Land continental shelf, into more granitic lithologies (Figure 6.4; 6.10a). During retreat, there is a rapid transition toward mafic compositions, as the erosive ice sheet margin retreats towards FLIP lithologies that underlie parts of the EAIS within the inland section of the

WSB (Figure 6.10c). This would also lead to enhanced nutrient delivery as the retreat of the ice sheet exposes the continental shelf seafloor. The winnowing by currents may also resuspend glaciomarine sediment — leading to a release in nutrients, particularly Fe, to the ocean which in turn promotes enhanced productivity in the coastal margin of Antarctica (Arrigo *et al.* 2015).

Figure 6.10 Schematic diagram of **a.** glacial, **b.** glacial–interglacial transition, and **c.** interglacial sedimentary processes occurring at Site U1361. During interglacials, the poleward invigoration of the westerly winds increases upwelling of warm, nutrient-rich waters, increasing marine productivity and CDW (red arrow) influx onto the continental shelf. The Antarctic Coastal Current is denoted by a circle with a black cross, indicating flow direction is into the diagram. The southern boundary of the ACC is represented by a circle with a black dot, with flow direction coming out of the diagram.

6.3 Orbital pacing of Wilkes Subglacial Basin ice dynamics

6.3.1 Late Miocene to Pleistocene iceberg rafted debris record from U1361

Time-series analysis of ice discharge and marine productivity from the Late Miocene to Early Pliocene from U1361 contain statistically significant frequencies corresponding to Milankovitch cycles (Section 5.6). This indicates the cyclic nature of these records is not a random process and is paced by variations in orbital parameters. In order to assess the response of the WSB to orbital pacing throughout the Plio–Pleistocene, the record established in this thesis has been combined with previously published IBRD data from the Late Pliocene to Early Pleistocene to provide a complete 4 million-year long record of ice rafting between 6.2 and 2.2 Ma (Patterson *et al.* 2014; Figure 6.11).

Figure 6.11 EHA results of U1361 IBRD with $\delta^{18}\text{O}$ and ETP records from Late Miocene to Pleistocene. **a.** combined IBRD MAR record. **b.** LR04 $\delta^{18}\text{O}$ (Lisiecki & Raymo 2004). **c.** Site 846 $\delta^{18}\text{O}$ (Shackleton 1995). **d.** ETP (Laskar *et al.* 2004).

The original dataset of Patterson *et al.* (2014) indicated iceberg discharge at the Wilkes Land margin was highly sensitive to obliquity forcing during inferred high atmospheric CO_2 conditions between 4.2 and 3.5 Ma, but transitions to precession at eccentricity pacing at

3.5 Ma (Figure 6.11a). It was hypothesised by Patterson *et al.* (2014) that under a warm climate state and high radiative forcing (e.g. high CO₂) the duration and intensity of austral summer warmth would be increased. This would result in the reduced extent or duration of the summer sea ice belt, allowing for greater influence of ocean processes in regulating ice sheet mass balance (Patterson *et al.* 2014). Obliquity forced shifts in the latitudinal temperature gradient resulted in obliquity paced shifts in the southern hemisphere zonal wind fields. As a result, the upwelling of warm subsurface waters onto the continental shelves forced marine ice sheet retreat (Patterson *et al.* 2014). However, a lowering in radiative forcing raised the melt threshold required to melt seasonal sea ice, thus extending the seasonal duration of sea ice. Consequently, the melt season of the sea ice is reduced, and the insolation threshold required to melt sea ice become skewed towards the peak intensity of summer insolation, which is a precession dominated signal (Patterson *et al.* 2014).

However, the extension of the IBRD dataset back to 6.2 Ma appears to contradict this hypothesis (Figure 6.11a). Strong obliquity does not appear to continue into the Early Pliocene, contrary to what is expected in a warmer-than-present climate (Huybers & Tziperman 2008; Patterson *et al.* 2014; Levy *et al.* 2019). Instead, this record displays significant eccentricity throughout the majority of the warmest intervals of the Early Pliocene, contrary to the dominant obliquity that is observed in the $\delta^{18}\text{O}$ record between ~4.2 to 3.5 Ma (Figure 6.11bc). It appears that the obliquity-dominated signal seen by Patterson *et al.* (2014) is a relatively short-lived interval within the context of the Late Miocene to Early Pleistocene IBRD from this site. While obliquity is present throughout much of the IBRD record (Figure 6.11a), it is subordinate to eccentricity which is persistent throughout the duration of the 6.2 to 4.2 Ma interval. In the LR04 $\delta^{18}\text{O}$ stack (Figure 6.10b), obliquity is the dominant signal, and while eccentricity is present throughout, it is subordinate. The LR04 $\delta^{18}\text{O}$ stack only extends back to 5.3 Ma (Fig. 6.11b), but comparison with ODP Site 846 (equatorial Pacific Ocean) to 6 Ma indicates obliquity dominance continues back to ~5.2Ma, and prior to this time eccentricity dominates this deep-sea signal.

The LA04 ETP orbital solution shows a quasi-linear relationship to IBRD MAR, where shifts in the 400/100 kyr eccentricity and 40 kyr obliquity bands can be observed in both records

(Laskar *et al.* 2004; Figure 6.11d). A drop in obliquity frequencies in the IBRD and ETP EHA occurs between 4.6 and 4.2 Ma, yet power in the obliquity band remains in the $\delta^{18}\text{O}$ (Figure 6.11). The loss of obliquity strength in both IBRD and LA04 at ~5.6 Ma is consistent with the drop in obliquity sensitivity observed by Levy *et al.* (2019) (Figure 2.1).

Significant peaks in IBRD and marine productivity can be visually correlated to records of $\delta^{18}\text{O}$ from South Pacific ODP Site 846 (Figure 6.12). The IBRD record is compared against this site as LR04 does not extend into the Late Miocene, and noise in the glacial–interglacial temperature changes are likely to be muted at this site due to the thermal inertia of the deep Pacific, and thus should be a record of Antarctic ice volume and deep-water outflow (Shackleton 1995). It is acknowledged these correlations do not represent a unique solution, but are consistent with the pacing inferred by spectral analysis results. Intervals of high marine productivity can be correlated to warm interglacials values of ~2.8‰ that occur at every ~100 kyr. Long duration intervals of marine productivity occur during prolonged interglacials. At ~5.2 Ma, a baseline shift in IBRD accumulation coincides with proxy records obtained from several Antarctic sites indicating warming conditions and a significant shift after 5.2 Ma in both the LR04 and Site 846 $\delta^{18}\text{O}$; with increased biogenic opal (Grützner *et al.* 2005), reduced sea ice (Whitehead *et al.* 2005), and a retreat of the Wilkes Land margin (Cook *et al.* 2013). The first evidence of open water conditions in the Ross Sea occurred at ~4.9 Ma is not recorded at U1361 (missing core section), however, subsequent periods of retreated WAIS occur during highly productive intervals at WSB.

Figure 6.12 Sedimentary and marine productivity record with global proxy records. a. IBRD MAR (this study) b. Ba/Al (this study). c. neodymium isotope values — low values of Nd = FLIP indicates retreat of WSB (Cook et al. 2013). d) LR04 $\delta^{18}\text{O}$ (Lisiecki & Raymo 2005). e. Site 846 $\delta^{18}\text{O}$ — raw data (grey) and filtered data (black) (Shackleton 1995). f. Light purple = SST > 5°C in Ross Sea (McKay et al. 2012); purple = SST > 4°C at Prydz Bay (Whitehead & Bohaty 2003); green = high productivity (Cook et al. 2013); pink = WAIS retreat (Naish et al. 2009); blue = Wilkes Land continental shelf open marine conditions (Reinardy et al. 2015).

The same tie points correlated to the $\delta^{18}\text{O}$ stack were extended to insolation parameters (Figure 6.13). Initial peaks in IBRD occur at the transition point of the peak in mean annual insolation, while highly productive periods generally occur at ~100 kyr cyclicity, consistent

with spectral analysis results (Figure 5.23), but appear to correlate to peaks in obliquity. Long duration productivity intervals span multiple obliquity cycles and occur at peaks in eccentricity. Tie points of IBRD also occur at every second to third obliquity cycle, which suggests an influence of the 40 kyr cycle does play a role in ice sheet variability. This may indicate a combined role of eccentricity and obliquity, where perennial sea ice was modulated by eccentricity such that the extent of sea ice was reduced every ~100 kyr, allowing obliquity-paced intrusions of warm CDW onto the continental shelf increasing ice discharge which results in pulses of IBRD (Toggweiler *et al.* 2006). This is consistent with the Levy *et al.* (2019) theory of the role of sea ice and oceanic forcing on ice sheet dynamics. Enhanced upwelling due to a decrease in sea ice during these highly productive intervals may have increased outgassing of CO₂, as seen at the termination of the Last Glacial Maximum (Anderson *et al.* 2009), and led to prolonged interglacial conditions. It is possible that the lack of obliquity at the top of the record is the result of conditions that are too warm to regrow the ice sheet, and thus long duration interglacials that are consistent with eccentricity pacing. This occurs at the same time as a retreated WAIS, open marine conditions on the Wilkes Land shelf as a consequence of warm SST, and reduced winter sea ice around the continent (Levy *et al.* 2012; McKay *et al.* 2012). The AND-1B drillcore records a series of obliquity-paced WAIS advances on the Ross Sea continental shelf between 4.9 Ma and 4.7 Ma. This interval in U1361 is characterised by a missing section between the base of Sidufjall (4.9 Ma) and base of Nunivak (4.63 Ma) subchrons, and thus the orbital pacing of IBRD MAR in this interval is difficult to constrain. It is likely there is missing section in this interval, perhaps associated with this short-lived phase of high amplitude glacial cycles noted in the LR04 benthic stack within the Sidufjall subchron (Figure 6.11).

Figure 6.12 Ice discharge and marine productivity between 6.2 and 4.4 Ma with insolation and orbital parameters. **a.** IBRD MAR. **b.** Ba/Al. **c.** total integrated summer insolation energy. **d.** mean annual insolation. **e.** eccentricity. **f.** obliquity.

This IBRD record from U1361 indicates there may be complications to the climate state dependence of orbital theory, and/or it is difficult to apply Northern Hemisphere models of orbital ice sheet theory to the AIS. The assumption of a 40 kyr dominated world during the Early Pliocene is not supported by this thesis. However, it remains consistent with theories that propose surface melting in Antarctica was not a control on the mass balance of the ice sheet after ~14 Ma (Lewis *et al.* 2007; Levy *et al.* 2019). Instead, oceanic forcing is the primary driver in ice sheet melt at this time. It may be that only during the Middle

Pliocene was the background climate state warm enough (e.g. increased atmospheric CO₂) to lower the melt threshold, such that increased surface melt and reduced sea ice resulted in a sustained 40 kyr response of the ice sheet (Huybers & Tziperman 2008; Patterson *et al.* 2014; Levy *et al.* 2019). This implies the prominent obliquity within the $\delta^{18}\text{O}$ stack may be a Northern Hemisphere ice sheet response. However, the lack of a comprehensive CO₂ record in this understudied period means it is difficult to link the background climate state to the response of the AIS. This may be remedied by a SST temperature record or diatom studies to better constrain background climate states at the site which may help to link insolation forcing to ice dynamics.

CHAPTER 7: CONCLUSIONS AND FUTURE WORK

7.1 Conclusions

This study examined the glacial-to-interglacial characteristics of sedimentation and marine productivity at Site U1361 between 6.2 to 4.4 Ma. This work extends a record of ice sheet dynamics from the Wilkes Land margin conducted by Patterson *et al.* (2014) and produces a relatively continuous high-resolution dataset of IBRD and XRF proxies across the Miocene to Pliocene boundary, where records of EAIS variability are scarce. These records assess the relationships between marine based EAIS and biological productivity in this sector of the Southern Ocean and confirms a dynamic system responding to orbital and oceanic forcing. The findings within this thesis are consistent with previous studies that suggested significant variability of the ice sheet during warmer-than-present Early Pliocene. The key findings are summarised below:

- The age model for this core was refined with a CONOP model using improved biostratigraphic and paleomagnetic constraints, producing a well-constrained age model with six paleomagnetic tie points, and a sample resolution of ~4 kyr with an average sedimentation rate of 3.7 cm/kyr.
- Facies analysis identified two primary facies — diatom-rich (/bearing) mud and massive muds — relating to the relative contribution of terrigenous and biogenic sedimentation, ultimately controlled by glacial–interglacial conditions. Sixteen glacial to interglacial cycles were identified, with sedimentary cycles containing a package of mud into diatom-bearing muds and then diatom-rich muds, which indicates a systematic increase in biological activity or decrease in terrigenous sedimentation through the transition from glacials to interglacials. Geochemical and physical properties strongly co-vary with facies.
 - Glacial maxima are characterised by massive muds containing low IBRD and low biogenic content indicating a relatively stable ice sheet margin and low marine productivity due to colder conditions and increased sea ice. Laminated muds contain no IBRD or biogenic material and are consistent with non-erosive spill-over of low-density turbidity currents, relating to increased terrigenous sediment

delivered to the continental shelf by glacial advance, triggering turbidity currents, to deliver sediment to the continental rise.

- Glacial minima are characterised by bioturbated diatom-bearing/rich facies containing increased IBRD, especially at the glacial termination. This facies contains enhanced marine productivity as reflected by consistently high values of estimated biogenic silica and Ba/Al ratios, which may indicate enhanced upwelling of CDW and nutrients. Al/Ti and K/Ti values are low, indicating an increased basaltic provenance of the terrigenous material. This is consistent with Cook *et al.* (2013) isotope work that suggests retreat of the ice margin into the basin interior during Early Pliocene interglacials (Cook *et al.* 2013).
- Primary drivers of terrigenous sedimentation were assessed using LPSA grain size data, geochemical grain size proxy (Zr/Rb), and geochemical provenance proxies (K/Ti and Al/Ti). Relationships between these data indicate not all the variability seen in the terrigenous sediment was a grain size signal, as only minor differences in grain size was seen in the LPSA data, suggesting a change in terrigenous source also occurred between facies.
- Identifying specific terrigenous sources was not possible with the data available due to the complex geology of the region and depositional complexities at the site. However, shifts within provenance ratios between facies were consistent with previous studies showing significant retreat into more mafic lithologies during interglacials. Slow increase in K/Ti during glacials may indicate the slow growth of the ice sheet, with sharp shifts at the glacial terminations occurring with IBRD peaks reflecting fast destabilization of the ice margin.
- Shifts in sediment provenance were paralleled by increases in marine productivity, preceded by a flux in IBRD. These changes may be the result of upwelled warm CDW water triggering retreat of the ice margin, releasing nutrients which in turn promote enhanced productivity.

- IBRD and marine productivity peaks could be correlated with the $\delta^{18}\text{O}$ record, with intervals of high productivity tied to $\delta^{18}\text{O}$ values below $\sim 2.9\%$, occurring at a frequency of ~ 100 kyr.
- Spectral analysis results confirm the observed presence of ~ 100 kyr cyclicity throughout the IBRD record, with secondary obliquity present in the Late Miocene. Where obliquity is significant, IBRD and facies packages are shorted in duration, and peaks in IBRD can be tied to approximately every second to third obliquity cycle. This suggests the oceanic influence on WSB may be modulated by eccentricity cycles controlling sea ice extent.
- A compilation record of Plio–Pleistocene IBRD from U1361 suggests that the strong obliquity forcing Patterson *et al.* (2014) observed between 4 and 3.5 Ma was anomalous to the predominant eccentricity forcing throughout the rest of the record. This is contrary to what is predicted in a warmer climate state, but further validation of Antarctic and global climate state at this time is required.

7.2 Future Work

Work is currently being completed on the Early Pleistocene record of U1361 which will complete the Plio–Pleistocene record of WSB ice sheet variability. More detailed provenance analysis through neodymium and strontium isotope compositions of fine-grained terrigenous sediment and identifying shifts in lithology of IBRD may help in establishing the timing and shifts of ice sheet retreat and assess the contribution of certain sub-basins or regions of the Ross Sea and Wilkes Land margin.

Redox minerals and chemistry identified through XRF were not explored in the scope of this thesis, but further investigation into the available data may highlight the timing of the change in water masses as the site.

This research highlights that the lack of comprehensive Late Miocene and Early Pliocene records inhibit interpretations of the response of the ice sheet to the background climate state. The understanding of ice sheet behaviour in response to orbital forcing could be improved by the development of high-resolution atmospheric CO_2 through this interval, alongside a high-resolution SST record using lipid biomarker analysis, in combination with

diatom and radiolarian assemblages, that will provide local SST variability during key climatic events, especially in the Earliest Pliocene and Late Miocene.

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APPENDICES

Appendix A: Iceberg rafted debris mass accumulation rate calculations.

Appendix B: Geochemical XRF ratios.

The following appendices are attached on Disk 1:

Appendix C: Grain size data and statistics.

Appendix D: Raw and corrected XRF data.

IODP Site U1361 IBRD MAR and facies code.

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
12	H	1	5	7	104.06	4432.071	1.154	0.750	2.938	0.0254	DR
12	H	1	15	17	104.16	4435.474	1.228	0.750	2.938	0.0271	DR
12	H	1	25	27	104.26	4438.878	0.681	0.750	2.938	0.0150	DR
12	H	1	35	37	104.36	4442.282	1.327	0.750	2.938	0.0292	DR
12	H	1	45	47	104.46	4445.686	1.308	0.750	2.938	0.0288	DR
12	H	1	55	57	104.56	4449.090	0.910	0.750	2.938	0.0201	DR
12	H	1	65	67	104.66	4452.494	1.095	0.750	2.938	0.0241	DR
12	H	1	75	77	104.76	4455.898	1.091	0.976	2.938	0.0313	DR
12	H	1	85	87	104.86	4459.302	1.496	0.976	2.938	0.0429	DR
12	H	1	97	99	104.98	4463.386	1.500	0.976	2.938	0.0430	DR
12	H	1	105	107	105.06	4466.109	2.022	0.976	2.938	0.0580	DR
12	H	1	115	117	105.16	4469.513	1.364	0.976	2.938	0.0391	DR
12	H	1	125	127	105.26	4472.917	1.000	0.976	2.938	0.0287	DR
12	H	1	135	137	105.36	4476.321	1.700	0.976	2.938	0.0487	DR
12	H	1	145	147	105.46	4479.725	1.514	0.976	2.938	0.0434	DR
12	H	2	5	7	105.56	4483.129	1.691	0.976	2.938	0.0485	DR
12	H	2	15	17	105.66	4486.533	1.318	0.976	2.938	0.0378	DR
12	H	2	25	27	105.76	4489.937	1.422	0.976	2.938	0.0408	DR
12	H	2	35	37	105.86	4493.360	1.079	0.976	2.921	0.0307	DR
12	H	3	5	7	105.94	4496.243	1.178	0.976	2.775	0.0319	DR
12	H	3	15	17	106.04	4499.846	0.850	0.976	2.775	0.0230	DR
12	H	3	25	27	106.14	4503.449	1.600	0.976	2.775	0.0433	DR
12	H	3	35	37	106.24	4507.052	1.047	0.976	2.775	0.0284	DR
12	H	4	8	10	106.36	4511.376	1.493	0.976	2.775	0.0404	DR
12	H	4	15	17	106.43	4513.898	1.159	0.976	2.775	0.0314	M
12	H	4	25	27	106.53	4517.501	0.752	0.976	2.775	0.0204	M
12	H	4	35	37	106.63	4521.104	0.459	0.976	2.775	0.0124	M
12	H	4	45	47	106.73	4524.708	0.805	0.976	2.775	0.0218	M
12	H	4	53	55	106.81	4527.590	1.508	0.976	2.775	0.0408	M
12	H	4	65	67	106.93	4531.914	0.167	0.976	2.775	0.0045	M
12	H	4	75	77	107.03	4535.517	0.387	0.976	2.775	0.0105	M
12	H	4	85	87	107.13	4539.120	0.295	0.976	2.775	0.0080	M
12	H	4	95	97	107.23	4542.723	0.934	0.976	2.775	0.0253	M
12	H	4	105	107	107.33	4546.326	0.300	0.976	2.775	0.0081	M
12	H	4	115	117	107.43	4549.930	0.367	0.976	2.775	0.0099	M
12	H	4	135	137	107.63	4557.136	0.084	0.997	2.775	0.0023	M
12	H	4	145	147	107.73	4560.739	0.243	0.997	2.775	0.0067	M
12	H	5	5	7	107.84	4564.702	0.383	0.997	2.775	0.0106	M
12	H	5	15	17	107.94	4568.305	0.048	0.997	2.775	0.0013	M
12	H	5	25	27	108.04	4571.909	0.377	0.997	2.775	0.0104	M
12	H	5	35	37	108.14	4575.512	0.101	0.997	2.775	0.0028	M
12	H	5	45	47	108.24	4579.115	0.171	0.997	2.775	0.0047	M

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
12	H	5	53	55	108.32	4581.997	0.543	0.997	2.775	0.0150	M
12	H	5	65	67	108.44	4586.321	0.077	0.997	2.775	0.0021	M
12	H	5	75	77	108.54	4589.924	0.159	0.997	2.775	0.0044	M
12	H	5	85	87	108.64	4593.527	1.270	0.997	2.775	0.0352	M
12	H	5	125	127	109.04	4607.940	2.052	0.736	2.775	0.0419	DR
12	H	5	135	137	109.14	4611.543	3.544	0.736	2.775	0.0724	DR
12	H	5	145	147	109.24	4615.146	0.982	0.736	2.775	0.0201	DR
12	H	6	5	7	109.35	4619.110	1.359	0.736	2.775	0.0278	DR
12	H	6	17	19	109.47	4623.433	0.578	0.736	2.775	0.0118	DR
12	H	6	25	27	109.55	4626.316	0.761	0.736	2.775	0.0155	DR
12	H	6	35	37	109.65	4629.919	1.413	0.736	2.775	0.0289	DR
12	H	6	47	49	109.77	4636.458	1.497	0.736	1.835	0.0202	DR
12	H	6	55	57	109.85	4641.309	1.879	0.736	1.649	0.0228	DR
12	H	6	77	79	110.07	4654.650	0.775	0.736	1.649	0.0094	DR
12	H	6	87	89	110.17	4660.714	0.565	0.736	1.649	0.0069	DR
12	H	6	97	99	110.27	4666.778	1.088	0.736	1.649	0.0132	DR
12	H	6	107	109	110.37	4672.842	0.998	0.852	1.649	0.0140	DR
12	H	6	117	119	110.47	4678.906	0.519	0.852	1.649	0.0073	DR
12	H	7	5	7	110.56	4684.364	0.011	0.852	1.649	0.0002	DR
12	H	7	6	8	110.57	4684.970	0.544	0.852	1.649	0.0076	DR
12	H	7	25	27	110.76	4696.492	1.880	0.852	1.649	0.0264	DR
12	H	7	35	37	110.86	4702.556	0.113	0.852	1.649	0.0016	DR
12	H	7	45	47	110.96	4708.620	0.050	0.852	1.649	0.0007	DR
12	H	7	55	57	111.06	4714.684	0.161	0.852	1.649	0.0023	DR
12	H	CC	5	7	111.16	4720.748	0.750	0.852	1.649	0.0105	DR
12	H	CC	15	17	111.26	4726.812	0.094	0.852	1.649	0.0013	DR
13	H	1	5	7	113.56	4866.286	2.036	0.893	1.649	0.0300	DR
13	H	1	15	17	113.66	4872.350	0.898	0.893	1.649	0.0132	DR
13	H	1	25	27	113.76	4878.414	0.596	0.893	1.649	0.0088	DR
13	H	1	35	37	113.86	4884.478	0.582	0.893	1.649	0.0086	DR
13	H	1	45	47	113.96	4890.542	0.221	0.893	1.649	0.0033	DR
13	H	1	55	57	114.06	4896.162	0.290	0.893	1.780	0.0046	DR
13	H	1	65	67	114.16	4897.778	0.608	0.893	6.188	0.0336	M
13	H	1	75	77	114.26	4899.394	0.401	0.893	6.188	0.0221	M
13	H	1	85	87	114.36	4901.010	0.331	0.893	6.188	0.0183	M
13	H	1	95	97	114.46	4902.626	0.130	0.893	6.188	0.0072	M
13	H	1	105	107	114.56	4904.242	0.556	0.893	6.188	0.0308	M
13	H	1	115	117	114.66	4905.858	1.146	0.893	6.188	0.0633	M
13	H	1	125	127	114.76	4907.474	0.303	0.893	6.188	0.0167	M
13	H	1	135	137	114.86	4909.090	0.059	0.893	6.188	0.0033	M
13	H	1	145	147	114.96	4910.706	2.254	0.893	6.188	0.1245	M
13	H	2	5	7	115.06	4912.322	0.445	0.945	6.188	0.0260	DR
13	H	2	35	37	115.36	4917.170	0.284	0.945	6.188	0.0166	DR
13	H	2	43	45	115.44	4918.462	2.930	0.945	6.188	0.1713	DB

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
13	H	2	55	57	115.56	4920.402	2.723	0.945	6.188	0.1593	DB
13	H	2	65	67	115.66	4922.018	1.328	0.945	6.188	0.0777	DB
13	H	2	75	77	115.76	4923.634	0.250	0.945	6.188	0.0146	M
13	H	2	85	87	115.86	4925.250	0.312	0.945	6.188	0.0182	M
13	H	2	95	97	115.96	4926.866	0.774	0.945	6.188	0.0452	M
13	H	2	105	107	116.06	4928.482	0.109	0.945	6.188	0.0064	M
13	H	2	115	117	116.16	4930.098	0.157	0.945	6.188	0.0092	M
13	H	2	125	127	116.26	4931.714	0.055	0.945	6.188	0.0032	M
13	H	2	135	137	116.36	4933.330	0.220	0.945	6.188	0.0128	M
13	H	2	145	147	116.46	4934.946	0.025	0.945	6.188	0.0015	M
13	H	3	5	7	116.56	4936.562	1.000	1.02	6.188	0.0631	DR
13	H	3	16	18	116.67	4938.339	1.180	1.020	6.188	0.0745	DR
13	H	3	25	27	116.76	4939.794	0.725	1.020	6.188	0.0458	DB
13	H	3	35	37	116.86	4941.410	0.183	1.020	6.188	0.0115	DB
13	H	3	45	47	116.96	4943.026	0.243	1.020	6.188	0.0154	DB
13	H	3	55	57	117.06	4944.642	0.056	1.020	6.188	0.0036	M
13	H	3	65	67	117.16	4946.258	0.120	1.020	6.188	0.0076	M
13	H	3	75	77	117.26	4947.874	0.000	1.020	6.188	0.0000	M
13	H	3	85	87	117.36	4949.490	0.025	1.02	6.188	0.0016	M
13	H	3	95	97	117.46	4951.106	0.084	1.020	6.188	0.0053	M
13	H	3	105	107	117.56	4952.722	0.041	1.020	6.188	0.0026	M
13	H	3	115	117	117.66	4954.338	0.393	1.020	6.188	0.0248	M
13	H	3	125	127	117.76	4955.954	0.062	1.020	6.188	0.0039	M
13	H	3	135	137	117.86	4957.570	0.567	1.020	6.188	0.0358	M
13	H	3	145	147	117.96	4959.186	0.167	1.020	6.188	0.0105	M
13	H	4	6	8	118.07	4960.963	0.198	0.997	6.188	0.0122	M
13	H	4	14	16	118.15	4962.256	0.013	0.997	6.188	0.0008	M
13	H	4	25	27	118.26	4964.034	0.104	0.997	6.188	0.0064	M
13	H	4	35	37	118.36	4965.650	0.323	0.997	6.188	0.0199	M
13	H	4	45	47	118.46	4967.266	0.072	0.997	6.188	0.0044	M
13	H	4	55	57	118.56	4968.882	0.618	0.997	6.188	0.0381	M
13	H	4	65	67	118.66	4970.498	0.099	0.997	6.188	0.0061	M
13	H	4	75	77	118.76	4972.114	0.078	0.997	6.188	0.0048	M
13	H	4	85	87	118.86	4973.730	0.131	0.997	6.188	0.0081	M
13	H	4	95	97	118.96	4975.346	0.402	0.997	6.188	0.0248	M
13	H	4	105	107	119.06	4976.962	0.010	0.997	6.188	0.0006	M
13	H	4	115	117	119.16	4978.578	0.025	0.997	6.188	0.0016	M
13	H	4	135	137	119.36	4981.810	0.044	0.997	6.188	0.0027	M
13	H	4	145	147	119.46	4983.426	0.505	0.997	6.188	0.0312	M
13	H	5	5	7	119.56	4985.042	4.933	0.791	6.188	0.2414	M
13	H	5	15	17	119.66	4986.658	0.413	0.791	6.188	0.0202	M
13	H	5	25	27	119.76	4988.274	1.277	0.791	6.188	0.0625	M
13	H	5	35	37	119.86	4989.890	1.820	0.791	6.188	0.0891	M
13	H	5	45	47	119.96	4991.506	0.019	0.791	6.188	0.0009	M

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
13	H	5	55	57	120.06	4993.122	3.071	0.791	6.188	0.1503	M
13	H	5	65	67	120.16	4994.738	1.482	0.791	6.188	0.0726	DR
13	H	5	85	87	120.36	5001.263	1.961	0.791	3.065	0.0476	DR
13	H	5	95	97	120.46	5008.367	0.950	0.791	1.408	0.0106	DR
13	H	5	105	107	120.56	5015.472	3.736	0.791	1.408	0.0416	DR
13	H	5	115	117	120.66	5022.576	3.079	0.791	1.408	0.0343	DR
13	H	5	125	127	120.76	5029.681	2.392	0.791	1.408	0.0266	DR
13	H	5	135	137	120.86	5036.785	1.557	0.791	1.408	0.0173	DR
13	H	5	145	147	120.96	5043.890	1.926	0.791	1.408	0.0214	DR
13	H	6	6	8	121.07	5051.704	0.589	0.805	1.408	0.0067	DR
13	H	6	15	17	121.16	5058.099	0.515	0.805	1.408	0.0058	DR
13	H	6	25	27	121.26	5065.203	0.706	0.805	1.408	0.0080	DR
13	H	6	35	37	121.36	5072.307	2.113	0.805	1.408	0.0239	DR
13	H	6	45	47	121.46	5079.412	1.027	0.805	1.408	0.0116	DR
13	H	6	55	57	121.56	5086.516	0.661	0.805	1.408	0.0075	DR
13	H	6	65	67	121.66	5093.621	0.422	0.805	1.408	0.0048	DR
13	H	6	85	87	121.86	5107.830	0.673	0.805	1.408	0.0076	DR
13	H	6	95	97	121.96	5114.934	0.442	0.805	1.408	0.0050	DR
13	H	6	105	107	122.06	5122.039	1.424	0.805	1.408	0.0161	DR
13	H	6	115	117	122.16	5129.143	0.723	0.805	1.408	0.0082	DR
13	H	6	125	127	122.26	5136.248	0.721	0.805	1.408	0.0082	DR
13	H	6	132	134	122.33	5141.221	2.343	0.805	1.408	0.0265	DR
13	H	CC	5	7	122.40	5146.194	0.838	0.805	1.408	0.0095	DR
13	H	CC	15	17	122.50	5153.299	0.381	0.805	1.408	0.0043	DR
14	H	1	5	7	123.06	5193.084	0.157	0.800	1.408	0.0018	DR
14	H	1	15	17	123.16	5200.188	0.589	0.800	1.408	0.0066	DR
14	H	1	25	27	123.26	5207.293	2.579	0.800	1.408	0.0290	DR
14	H	1	35	37	123.36	5214.397	0.914	0.800	1.408	0.0103	DR
14	H	1	45	47	123.46	5221.501	7.577	0.800	1.408	0.0853	DR
14	H	1	55	57	123.56	5228.606	0.782	0.800	1.408	0.0088	DR
14	H	1	85	87	123.86	5243.317	10.116	0.800	2.039	0.1650	DR
14	H	1	95	97	123.96	5247.277	3.813	0.800	2.525	0.0770	DR
14	H	1	105	107	124.06	5251.237	2.644	0.800	2.525	0.0534	DR
14	H	1	115	117	124.16	5255.198	0.519	0.800	2.525	0.0105	DR
14	H	1	124	126	124.25	5258.762	0.144	0.800	2.525	0.0029	DR
14	H	1	135	137	124.36	5263.118	0.665	0.800	2.525	0.0134	M
14	H	1	145	147	124.46	5267.078	0.174	0.800	2.525	0.0035	M
14	H	2	5	7	124.56	5271.039	0.487	0.930	2.525	0.0114	M
14	H	2	15	17	124.66	5274.999	0.207	0.930	2.525	0.0049	M
14	H	2	25	27	124.76	5278.959	0.201	0.930	2.525	0.0047	M
14	H	2	35	37	124.86	5282.920	0.622	0.930	2.525	0.0146	M
14	H	2	45	47	124.96	5286.880	0.487	0.930	2.525	0.0114	M
14	H	2	57	59	125.08	5291.632	0.057	0.930	2.525	0.0013	M
14	H	2	65	67	125.16	5294.800	0.386	0.93	2.525	0.0091	M

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
14	H	2	75	77	125.26	5298.761	0.180	0.930	2.525	0.0042	M
14	H	2	95	97	125.46	5306.681	0.119	0.930	2.525	0.0028	M
14	H	2	105	107	125.56	5310.642	0.135	0.930	2.525	0.0032	M
14	H	2	115	117	125.66	5314.602	0.000	0.930	2.525	0.0000	M
14	H	2	125	127	125.76	5318.562	1.776	0.930	2.525	0.0417	M
14	H	2	135	137	125.86	5322.523	3.165	0.930	2.525	0.0743	DR
14	H	2	145	147	125.96	5326.483	0.702	0.930	2.525	0.0165	DR
14	H	3	6	8	126.07	5330.839	1.575	0.951	2.525	0.0378	DR
14	H	3	15	17	126.16	5334.403	0.671	0.951	2.525	0.0161	DR
14	H	3	25	27	126.26	5338.364	0.369	0.951	2.525	0.0089	DR
14	H	3	35	37	126.36	5342.324	0.142	0.951	2.525	0.0034	DR
14	H	3	45	47	126.46	5346.284	0.000	0.951	2.525	0.0000	DR
14	H	3	55	57	126.56	5350.245	0.103	0.951	2.525	0.0025	M
14	H	3	64	66	126.65	5353.809	0.099	0.951	2.525	0.0024	M
14	H	3	75	77	126.76	5358.165	0.279	0.951	2.525	0.0067	M
14	H	3	85	87	126.86	5362.126	0.127	0.951	2.525	0.0031	M
14	H	3	95	97	126.96	5366.086	0.161	0.951	2.525	0.0039	M
14	H	3	105	107	127.06	5370.046	0.408	0.951	2.525	0.0098	DR
14	H	3	115	117	127.16	5374.006	0.780	0.951	2.525	0.0187	DR
14	H	3	125	127	127.26	5377.967	1.754	0.951	2.525	0.0421	DR
14	H	3	135	137	127.36	5381.927	0.503	0.951	2.525	0.0121	DR
14	H	3	145	147	127.46	5385.887	0.343	1.002	2.525	0.0087	DR
14	H	4	5	7	127.56	5389.848	0.226	1.002	2.525	0.0057	DB
14	H	4	15	17	127.66	5393.808	0.374	1.002	2.525	0.0095	DB
14	H	4	25	27	127.76	5397.768	0.093	1.002	2.525	0.0024	DB
14	H	4	35	37	127.86	5401.729	0.112	1.002	2.525	0.0028	M
14	H	4	45	47	127.96	5405.689	0.020	1.002	2.525	0.0005	M
14	H	4	55	57	128.06	5409.649	0.143	1.002	2.525	0.0036	M
14	H	4	65	67	128.16	5413.609	0.306	1.002	2.525	0.0077	M
14	H	4	75	77	128.26	5417.570	0.047	1.002	2.525	0.0012	M
14	H	4	85	87	128.36	5421.530	0.071	1.002	2.525	0.0018	M
14	H	4	95	97	128.46	5425.490	0.152	1.002	2.525	0.0038	M
14	H	4	105	107	128.56	5429.451	0.462	1.002	2.525	0.0117	M
14	H	4	115	117	128.66	5433.411	0.348	1.002	2.525	0.0088	M
14	H	4	125	127	128.76	5437.371	0.613	1.002	2.525	0.0155	M
14	H	4	135	137	128.86	5441.332	0.478	1.002	2.525	0.0121	DR
14	H	4	145	147	128.96	5445.292	0.288	1.002	2.525	0.0073	DR
14	H	5	5	7	129.06	5449.252	0.437	1.002	2.525	0.0111	DB
14	H	5	15	17	129.16	5453.212	0.142	0.841	2.525	0.0030	DB
14	H	5	25	27	129.26	5457.173	0.484	0.841	2.525	0.0103	DB
14	H	5	35	37	129.36	5461.133	0.394	0.841	2.525	0.0084	M
14	H	5	45	47	129.46	5465.093	0.013	0.841	2.525	0.0003	M
14	H	5	55	57	129.56	5469.054	0.000	0.841	2.525	0.0000	M
14	H	5	65	67	129.66	5473.014	0.043	0.841	2.525	0.0009	M

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
14	H	5	75	77	129.76	5476.974	0.645	0.841	2.525	0.0137	M
14	H	5	85	87	129.86	5480.934	1.145	0.841	2.525	0.0243	DR
14	H	5	95	97	129.96	5484.895	1.584	0.841	2.525	0.0336	DR
14	H	5	105	107	130.06	5488.855	1.281	0.841	2.525	0.0272	DR
14	H	5	115	117	130.16	5492.815	1.041	0.841	2.525	0.0221	DB
14	H	5	125	127	130.26	5496.776	0.925	0.841	2.525	0.0196	DB
14	H	5	135	137	130.36	5500.736	1.005	0.841	2.525	0.0213	DR
14	H	5	145	147	130.46	5504.696	0.026	0.841	2.525	0.0005	DR
14	H	6	5	7	130.56	5508.657	0.257	1.070	2.525	0.0069	DB
14	H	6	15	17	130.66	5512.617	0.487	1.07	2.525	0.0132	DB
14	H	6	25	27	130.76	5516.577	0.020	1.070	2.525	0.0005	M
14	H	6	35	37	130.86	5520.537	0.000	1.070	2.525	0.0000	M
14	H	6	45	47	130.96	5524.498	0.000	1.070	2.525	0.0000	M
14	H	6	55	57	131.06	5528.458	0.070	1.070	2.525	0.0019	M
14	H	6	65	67	131.16	5532.418	0.000	1.070	2.525	0.0000	M
14	H	6	75	77	131.26	5536.379	0.259	1.070	2.525	0.0070	M
14	H	6	85	87	131.36	5540.339	0.167	1.070	2.525	0.0045	M
14	H	6	95	97	131.46	5544.299	0.070	1.070	2.525	0.0019	M
14	H	6	105	107	131.56	5548.260	0.109	0.666	2.525	0.0018	M
14	H	6	115	117	131.66	5552.220	0.046	0.666	2.525	0.0008	M
14	H	6	125	127	131.76	5556.180	1.174	0.666	2.525	0.0197	DB
14	H	7	5	7	131.86	5560.140	3.056	0.666	2.525	0.0514	DB
14	H	7	15	17	131.96	5564.101	1.035	0.666	2.525	0.0174	DB
14	H	7	25	27	132.06	5568.061	1.195	0.666	2.525	0.0201	DB
14	H	7	35	37	132.16	5572.021	1.735	0.666	2.525	0.0292	DB
14	H	7	45	47	132.26	5575.982	1.464	0.666	2.525	0.0246	DB
14	H	7	55	57	132.36	5579.942	1.801	0.666	2.525	0.0303	DR
15	H	1	15	17	132.66	5591.823	0.096	0.981	2.525	0.0024	DR
15	H	1	25	27	132.76	5595.783	0.219	0.981	2.525	0.0054	DB
15	H	1	35	37	132.86	5599.743	0.253	0.981	2.525	0.0063	DB
15	H	1	45	47	132.96	5603.704	0.221	0.981	2.525	0.0055	DB
15	H	1	55	57	133.06	5607.664	0.356	0.981	2.525	0.0088	DB
15	H	1	65	67	133.16	5611.624	0.138	0.981	2.525	0.0034	DB
15	H	1	75	77	133.26	5615.585	0.475	0.981	2.525	0.0118	DB
15	H	1	85	87	133.36	5619.545	0.039	0.981	2.525	0.0010	M
15	H	1	95	97	133.46	5623.505	0.071	0.981	2.525	0.0018	M
15	H	1	105	107	133.56	5627.466	0.494	0.981	2.525	0.0122	M
15	H	1	115	117	133.66	5631.426	0.812	0.981	2.525	0.0201	DR
15	H	1	125	127	133.76	5635.386	0.654	0.981	2.525	0.0162	DR
15	H	1	135	137	133.86	5639.346	0.416	0.981	2.525	0.0103	DR
15	H	1	145	147	133.96	5643.307	0.955	0.981	2.525	0.0237	DR
15	H	2	15	17	134.16	5651.227	0.264	0.926	2.525	0.0062	DR
15	H	2	25	27	134.26	5655.188	0.350	0.926	2.525	0.0082	DR
15	H	2	35	37	134.36	5659.148	0.640	0.926	2.525	0.0150	DR

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
15	H	2	45	47	134.46	5663.108	0.384	0.926	2.525	0.0090	DR
15	H	2	55	57	134.56	5667.068	0.518	0.926	2.525	0.0121	DR
15	H	2	65	67	134.66	5671.029	2.021	0.926	2.525	0.0473	DR
15	H	2	75	77	134.76	5674.989	0.732	0.926	2.525	0.0171	DB
15	H	2	85	87	134.86	5678.949	0.506	0.926	2.525	0.0118	DB
15	H	2	105	107	135.06	5686.870	0.760	0.926	2.525	0.0178	DB
15	H	2	115	117	135.16	5690.830	0.240	0.926	2.525	0.0056	DB
15	H	2	123	125	135.24	5693.999	0.322	0.926	2.525	0.0075	DB
15	H	2	135	137	135.36	5698.751	1.259	0.926	2.525	0.0294	M
15	H	2	145	147	135.46	5702.711	1.139	0.926	2.525	0.0266	M
15	H	3	5	7	135.56	5706.671	3.759	1.041	2.525	0.0988	M
15	H	3	15	17	135.66	5710.632	0.055	1.041	2.525	0.0014	M
15	H	3	25	27	135.76	5714.592	0.114	1.041	2.525	0.0030	M
15	H	3	35	37	135.86	5718.552	0.095	1.041	2.525	0.0025	M
15	H	3	45	47	135.96	5722.513	0.143	1.041	2.525	0.0038	M
15	H	3	55	57	136.06	5726.473	0.468	1.041	2.525	0.0123	M
15	H	3	65	67	136.16	5730.433	0.121	1.041	2.525	0.0032	M
15	H	3	75	77	136.26	5734.394	0.465	1.041	2.525	0.0122	M
15	H	3	85	87	136.36	5738.354	0.027	1.041	2.525	0.0007	M
15	H	3	95	97	136.46	5742.314	0.823	1.041	2.525	0.0216	M
15	H	3	105	107	136.56	5746.274	0.158	1.041	2.525	0.0042	M
15	H	3	115	117	136.66	5750.235	0.241	1.041	2.525	0.0063	M
15	H	3	125	127	136.76	5754.195	0.216	1.041	2.525	0.0057	M
15	H	3	137	139	136.88	5758.947	0.599	1.041	2.525	0.0157	M
15	H	3	145	147	136.96	5762.116	0.305	1.041	2.525	0.0080	M
15	H	4	5	7	137.06	5766.076	0.099	0.779	2.525	0.0019	M
15	H	4	15	17	137.16	5770.036	0.026	0.779	2.525	0.0005	M
15	H	4	25	27	137.26	5773.997	0.076	0.779	2.525	0.0015	M
15	H	4	35	37	137.36	5777.957	0.476	0.779	2.525	0.0094	M
15	H	4	45	47	137.46	5781.917	0.119	0.779	2.525	0.0023	M
15	H	4	55	57	137.56	5785.877	0.245	0.779	2.525	0.0048	M
15	H	4	65	67	137.66	5789.838	0.519	0.779	2.525	0.0102	DR
15	H	4	85	87	137.86	5797.758	0.371	0.779	2.525	0.0073	DR
15	H	4	95	97	137.96	5801.719	1.291	0.779	2.525	0.0254	DB
15	H	4	105	107	138.06	5805.679	0.731	0.779	2.525	0.0144	DB
15	H	4	115	117	138.16	5809.639	0.712	0.779	2.525	0.0140	DB
15	H	4	125	127	138.26	5813.600	0.031	0.779	2.525	0.0006	DB
15	H	4	135	137	138.36	5817.560	0.264	0.779	2.525	0.0052	M
15	H	4	145	147	138.46	5821.520	0.239	0.779	2.525	0.0047	M
15	H	5	5	7	138.56	5825.480	0.083	0.840	2.525	0.0018	M
15	H	5	15	17	138.66	5829.441	0.000	0.840	2.525	0.0000	M
15	H	5	25	27	138.76	5833.401	0.332	0.840	2.525	0.0070	DB
15	H	5	35	37	138.86	5837.361	0.337	0.840	2.525	0.0072	DB
15	H	5	45	47	138.96	5841.322	1.604	0.840	2.525	0.0340	DB

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
15	H	5	55	57	139.06	5845.282	1.775	0.840	2.525	0.0376	DR
15	H	5	65	67	139.16	5849.242	0.702	0.840	2.525	0.0149	DR
15	H	5	75	77	139.26	5853.202	2.061	0.840	2.525	0.0437	DR
15	H	5	85	87	139.36	5857.163	1.158	0.840	2.525	0.0246	DR
15	H	5	95	97	139.46	5861.123	0.760	0.840	2.525	0.0161	DR
15	H	5	105	107	139.56	5865.083	1.329	0.84	2.525	0.0282	DR
15	H	5	115	117	139.66	5869.044	0.142	0.840	2.525	0.0030	DB
15	H	5	125	127	139.76	5873.004	0.265	0.840	2.525	0.0056	DB
15	H	5	135	137	139.86	5876.964	0.353	0.840	2.525	0.0075	M
15	H	5	145	147	139.96	5880.925	0.234	0.840	2.525	0.0050	M
15	H	6	15	17	140.16	5888.845	1.075	0.996	2.525	0.0270	M
15	H	6	25	27	140.26	5892.805	0.009	0.996	2.525	0.0002	M
15	H	6	35	37	140.36	5896.766	0.759	0.996	2.525	0.0191	M
15	H	6	45	47	140.46	5900.726	0.065	0.996	2.525	0.0016	M
15	H	6	55	57	140.56	5904.686	0.098	0.996	2.525	0.0025	M
15	H	6	65	67	140.66	5908.647	0.095	0.996	2.525	0.0024	M
15	H	6	75	77	140.76	5912.607	0.073	0.996	2.525	0.0018	M
15	H	6	85	87	140.86	5916.567	0.000	0.996	2.525	0.0000	M
15	H	6	95	97	140.96	5920.528	0.236	0.996	2.525	0.0059	M
15	H	6	105	107	141.06	5924.488	0.178	0.996	2.525	0.0045	M
15	H	6	115	117	141.16	5928.448	0.086	0.996	2.525	0.0022	M
15	H	6	125	127	141.26	5932.408	0.211	0.996	2.525	0.0053	M
15	H	7	5	7	141.36	5936.369	0.000	0.996	2.525	0.0000	DR
15	H	7	15	17	141.46	5940.329	0.024	0.996	2.525	0.0006	DR
15	H	7	25	27	141.56	5944.289	0.286	0.996	2.525	0.0072	DR
15	H	7	35	37	141.66	5948.250	0.049	0.996	2.525	0.0012	DR
16	H	1	5	7	142.06	5964.091	1.686	0.583	2.525	0.0248	DR
16	H	1	15	17	142.16	5968.051	1.718	0.583	2.525	0.0253	DR
16	H	1	25	27	142.26	5972.011	11.367	0.583	2.525	0.1673	DR
16	H	1	35	37	142.36	5975.972	1.184	0.583	2.525	0.0174	DR
16	H	1	45	47	142.46	5979.932	0.819	0.583	2.525	0.0121	DR
16	H	1	65	67	142.66	5987.853	0.333	0.583	2.525	0.0049	DR
16	H	1	75	77	142.76	5991.813	0.403	0.583	2.525	0.0059	DR
16	H	1	85	87	142.86	5995.773	0.573	0.583	2.525	0.0084	DR
16	H	1	95	97	142.96	5999.733	0.788	0.583	2.525	0.0116	DR
16	H	1	105	107	143.06	6003.694	0.677	0.583	2.525	0.0100	DR
16	H	1	115	117	143.16	6007.654	0.526	0.583	2.525	0.0077	DR
16	H	1	125	127	143.26	6011.614	2.941	0.583	2.525	0.0433	DR
16	H	1	145	147	143.46	6019.535	0.444	0.583	2.525	0.0065	DR
16	H	2	5	7	143.56	6023.495	0.206	0.666	2.525	0.0035	DR
16	H	2	15	17	143.66	6027.456	1.079	0.666	2.525	0.0181	DR
16	H	2	25	27	143.76	6031.416	0.953	0.666	2.525	0.0160	DR
16	H	2	35	37	143.86	6034.314	0.570	0.666	3.451	0.0131	DR
16	H	2	45	47	143.96	6036.504	1.428	0.666	4.566	0.0434	DR

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
16	H	2	55	57	144.06	6038.694	1.598	0.666	4.566	0.0486	DR
16	H	2	65	67	144.16	6040.884	1.216	0.666	4.566	0.0370	DR
16	H	2	75	77	144.26	6043.074	0.701	0.666	4.566	0.0213	DR
16	H	2	85	87	144.36	6045.264	0.397	0.666	4.566	0.0121	DR
16	H	2	95	97	144.46	6047.454	0.958	0.666	4.566	0.0291	DR
16	H	2	105	107	144.56	6049.644	1.802	0.666	4.566	0.0548	DR
16	H	2	115	117	144.66	6051.834	0.331	0.666	4.566	0.0101	DR
16	H	2	125	127	144.76	6054.024	0.035	0.666	4.566	0.0011	DR
16	H	2	135	137	144.86	6056.214	0.934	0.666	4.566	0.0284	DR
16	H	2	145	147	144.96	6058.404	0.446	0.666	4.566	0.0136	DR
16	H	3	5	7	145.06	6060.594	0.424	0.949	4.566	0.0184	DR
16	H	3	15	17	145.16	6062.784	0.342	0.949	4.566	0.0148	DR
16	H	3	25	27	145.26	6064.974	1.504	0.949	4.566	0.0652	M
16	H	3	35	37	145.36	6067.164	0.189	0.949	4.566	0.0082	M
16	H	3	45	47	145.46	6069.354	0.156	0.949	4.566	0.0068	M
16	H	3	55	57	145.56	6071.544	0.447	0.949	4.566	0.0194	M
16	H	3	65	67	145.66	6073.734	1.170	0.949	4.566	0.0507	M
16	H	3	85	87	145.86	6078.114	0.055	0.949	4.566	0.0024	M
16	H	3	95	97	145.96	6080.304	0.564	0.949	4.566	0.0245	M
16	H	3	105	107	146.06	6082.494	0.076	0.949	4.566	0.0033	M
16	H	3	115	117	146.16	6084.684	0.104	0.949	4.566	0.0045	M
16	H	3	125	127	146.26	6086.874	0.086	0.949	4.566	0.0037	M
16	H	3	135	137	146.36	6089.064	0.151	0.949	4.566	0.0066	M
16	H	3	145	147	146.46	6091.254	0.237	0.949	4.566	0.0103	M
16	H	4	5	7	146.56	6093.444	0.114	0.961	4.566	0.0050	M
16	H	4	15	17	146.66	6095.634	0.463	0.961	4.566	0.0203	M
16	H	4	25	27	146.76	6097.824	0.239	0.961	4.566	0.0105	M
16	H	4	35	37	146.86	6100.014	0.194	0.961	4.566	0.0085	M
16	H	4	45	47	146.96	6102.204	0.225	0.961	4.566	0.0099	DR
16	H	4	55	57	147.06	6104.394	0.133	0.961	4.566	0.0058	DR
16	H	4	65	67	147.16	6106.584	0.381	0.961	4.566	0.0167	DR
16	H	4	75	77	147.26	6108.774	1.916	0.961	4.566	0.0841	DR
16	H	4	85	87	147.36	6110.964	0.215	0.961	4.566	0.0094	DR
16	H	4	95	97	147.46	6113.154	0.298	0.961	4.566	0.0131	DR
16	H	4	105	107	147.56	6115.344	0.455	0.961	4.566	0.0200	DR
16	H	4	115	117	147.66	6117.534	0.116	0.961	4.566	0.0051	DR
16	H	4	125	127	147.76	6119.724	0.067	0.961	4.566	0.0029	DR
16	H	4	135	137	147.86	6121.914	0.021	0.961	4.566	0.0009	DR
16	H	4	145	147	147.96	6124.104	0.930	0.961	4.566	0.0408	DR
16	H	5	5	7	148.06	6126.294	0.713	0.809	4.566	0.0263	DR
16	H	5	15	17	148.16	6128.484	0.619	0.809	4.566	0.0229	DR
16	H	5	25	27	148.26	6130.674	0.868	0.809	4.566	0.0321	DR
16	H	5	35	37	148.36	6132.864	0.832	0.809	4.566	0.0307	DR
16	H	5	45	47	148.46	6135.054	0.299	0.809	4.566	0.0110	DR

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	LSR (cm/k.y)	CS (%)	DBD (g/cm ³)	IBRD MAR (g.cm ² .kyr)	Facies Code
16	H	5	55	57	148.56	6137.244	0.388	0.809	4.566	0.0143	DR
16	H	5	65	67	148.66	6139.434	0.633	0.809	4.566	0.0234	DR
16	H	5	75	77	148.76	6141.624	0.326	0.809	4.566	0.0120	DR
16	H	5	85	87	148.86	6143.814	0.033	0.809	4.566	0.0012	DR
16	H	5	95	97	148.96	6146.004	1.029	0.809	4.566	0.0380	DR
16	H	5	105	107	149.06	6148.194	0.294	0.809	4.566	0.0109	DR
16	H	5	115	117	149.16	6150.384	0.220	0.809	4.566	0.0081	DR
16	H	5	125	127	149.26	6152.574	0.514	0.809	4.566	0.0190	DR
16	H	5	135	137	149.36	6154.764	0.459	0.809	4.566	0.0169	DR
16	H	5	145	147	149.46	6156.954	0.361	0.809	4.566	0.0133	DR
16	H	6	5	7	149.56	6159.144	0.334	0.720	4.566	0.0110	DR
16	H	6	15	17	149.66	6161.334	0.780	0.720	4.566	0.0257	DR
16	H	6	25	27	149.76	6163.524	0.634	0.720	4.566	0.0208	DR
16	H	6	55	57	150.06	6170.094	0.299	0.720	4.566	0.0098	DR
16	H	6	65	67	150.16	6172.284	0.220	0.720	4.566	0.0072	DR
16	H	6	75	77	150.26	6174.474	0.323	0.720	4.566	0.0106	DR
16	H	6	85	87	150.36	6176.664	0.395	0.720	4.566	0.0130	DR
16	H	6	95	97	150.46	6178.854	0.620	0.720	4.566	0.0204	DB
16	H	6	106	108	150.57	6181.263	0.918	0.720	4.566	0.0302	DB
16	H	7	5	7	150.68	6183.672	0.669	0.720	4.566	0.0220	DB
16	H	7	15	17	150.78	6185.862	0.114	0.720	4.566	0.0037	DB
16	H	7	25	27	150.88	6188.052	0.264	1.125	4.566	0.0136	M
16	H	7	35	37	150.98	6190.242	0.087	1.125	4.566	0.0045	M
16	H	7	45	47	151.08	6192.432	0.000	1.125	4.566	0.0000	M
16	H	7	55	57	151.18	6194.622	0.146	1.125	4.566	0.0075	M
16	H	7	65	67	151.28	6196.812	0.061	1.125	4.566	0.0031	M
16	H	7	75	77	151.38	6199.002	0.046	1.125	4.566	0.0024	M
16	H	7	84	86	151.47	6200.973	0.089	1.125	4.566	0.0046	M
16	H	CC	5	7	151.54	6202.506	0.171	1.125	4.566	0.0088	M
16	H	CC	15	17	151.64	6204.696	0.143	1.125	4.566	0.0073	M

IODP SITE U1361 XRF ratios

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
12	H	1	5	7	104.06	4432.071	DR	0.0217	5.488	13.190	6.064	1.542	104
12	H	1	15	17	104.16	4435.474	DR	0.0218	6.045	12.567	5.888	1.781	104
12	H	1	25	27	104.26	4438.878	DR	0.0097	5.203	15.211	7.037	1.372	94
12	H	1	35	37	104.36	4442.282	DR	0.0196	5.827	12.595	5.978	1.691	107
12	H	1	45	47	104.46	4445.686	DR	0.0225	6.177	12.146	5.799	1.691	108
12	H	1	55	57	104.56	4449.090	DR	0.0222	6.468	12.237	5.602	1.851	111
12	H	1	65	67	104.66	4452.494	DR	0.0220	5.735	12.767	5.992	1.599	112
12	H	1	75	77	104.76	4455.898	DR	0.0221	6.497	11.953	5.479	1.850	104
12	H	1	85	87	104.86	4459.302	DR	0.0232	5.898	11.982	5.864	1.812	111
12	H	1	97	99	104.98	4463.386	DR	0.0200	5.391	13.310	6.293	1.507	108
12	H	1	105	107	105.06	4466.109	DR	0.0233	5.826	12.107	5.782	1.818	112
12	H	1	115	117	105.16	4469.513	DR	0.0231	5.889	12.464	5.797	1.879	108
12	H	1	125	127	105.26	4472.917	DR	0.0213	5.804	12.487	5.861	1.789	107
12	H	1	135	137	105.36	4476.321	DR	0.0221	5.803	12.102	5.819	1.751	109
12	H	1	145	147	105.46	4479.725	DR	0.0205	5.815	12.406	6.076	1.614	108
12	H	2	5	7	105.56	4483.129	DR	0.0231	5.975	12.292	5.698	1.876	112
12	H	2	15	17	105.66	4486.533	DR	0.0232	5.784	12.158	5.673	1.940	104
12	H	2	25	27	105.76	4489.937	DR	0.0228	5.958	12.216	5.828	1.978	110
12	H	2	35	37	105.86	4493.360	DR	0.0209	5.764	12.651	5.972	1.689	110
12	H	3	5	7	105.94	4496.243	DR	0.0215	5.529	12.741	6.001	1.824	108
12	H	3	15	17	106.04	4499.846	DR	0.0215	5.596	12.400	5.789	1.713	108
12	H	3	25	27	106.14	4503.449	DR	0.0216	5.294	12.532	5.976	1.579	116
12	H	3	35	37	106.24	4507.052	DR	0.0206	5.261	12.581	5.981	1.545	107
12	H	4	8	10	106.36	4511.376	DR	0.0200	5.309	12.677	5.909	1.553	112
12	H	4	15	17	106.43	4513.898	M	0.0184	5.003	12.432	6.388	1.328	107
12	H	4	25	27	106.53	4517.501	M	0.0090	4.219	14.799	7.968	0.968	99
12	H	4	35	37	106.63	4521.104	M	0.0085	5.140	15.685	7.677	1.102	94
12	H	4	45	47	106.73	4524.708	M	0.0098	3.833	15.604	8.906	0.754	92
12	H	4	53	55	106.81	4527.590	M	0.0100	4.789	15.181	7.604	1.107	102
12	H	4	65	67	106.93	4531.914	M	0.0076	4.946	15.171	7.822	0.883	91
12	H	4	75	77	107.03	4535.517	M	0.0074	5.448	14.763	8.145	0.967	123
12	H	4	85	87	107.13	4539.120	M	0.0081	5.147	15.953	7.345	0.994	129
12	H	4	95	97	107.23	4542.723	M	0.0092	4.141	17.286	7.524	0.859	124
12	H	4	105	107	107.33	4546.326	M	0.0094	5.219	15.404	7.842	0.985	109
12	H	4	115	117	107.43	4549.930	M	0.0115	5.649	16.960	8.116	1.402	145
12	H	4	135	137	107.63	4557.136	M	0.0082	6.156	14.141	5.969	1.126	125
12	H	4	145	147	107.73	4560.739	M	0.0085	4.728	15.197	7.152	1.149	130
12	H	5	5	7	107.84	4564.702	M	0.0096	7.025	16.285	7.285	1.457	137
12	H	5	15	17	107.94	4568.305	M	0.0080	5.903	15.476	7.296	1.093	122
12	H	5	25	27	108.04	4571.909	M	0.0087	4.312	16.575	6.547	1.186	141
12	H	5	35	37	108.14	4575.512	M	0.0082	4.322	16.681	7.867	1.223	129
12	H	5	45	47	108.24	4579.115	M	0.0084	5.095	16.595	7.479	1.060	119
12	H	5	53	55	108.32	4581.997	M	0.0171	5.409	13.595	5.748	1.507	143

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
12	H	5	65	67	108.44	4586.321	M	0.0105	6.483	14.257	5.606	1.329	139
12	H	5	75	77	108.54	4589.924	M	0.0087	5.248	14.309	6.788	1.369	143
12	H	5	85	87	108.64	4593.527	M	0.0196	5.544	13.228	5.786	1.534	152
12	H	5	125	127	109.04	4607.940	DR	0.0226	6.975	12.310	5.591	1.870	133
12	H	5	135	137	109.14	4611.543	DR	0.0228	8.157	10.482	4.572	1.882	137
12	H	5	145	147	109.24	4615.146	DR	0.0282	9.369	9.070	3.913	2.043	156
12	H	6	5	7	109.35	4619.110	DR	0.0248	7.684	10.771	4.741	1.845	141
12	H	6	17	19	109.47	4623.433	DR	0.0185	6.787	12.611	6.175	1.462	115
12	H	6	25	27	109.55	4626.316	DR	0.0180	7.126	11.969	5.749	1.465	119
12	H	6	35	37	109.65	4629.919	DR	0.0220	7.559	11.215	4.980	1.765	132
12	H	6	47	49	109.77	4636.458	DR	0.0217	7.150	12.276	5.143	1.874	128
12	H	6	55	57	109.85	4641.309	DR	0.0241	6.859	11.939	5.324	1.750	149
12	H	6	77	79	110.07	4654.650	DR	0.0215	8.483	10.898	4.732	1.955	144
12	H	6	87	89	110.17	4660.714	DR	0.0236	7.591	11.535	5.030	1.740	134
12	H	6	97	99	110.27	4666.778	DR	0.0229	7.081	12.590	5.679	1.484	143
12	H	6	107	109	110.37	4672.842	DR	0.0228	7.783	10.610	4.859	2.032	159
12	H	6	117	119	110.47	4678.906	DR	0.0235	7.754	10.762	4.727	1.954	144
12	H	7	5	7	110.56	4684.364	DR	0.0201	7.216	11.651	5.291	1.778	142
12	H	7	6	8	110.57	4684.970	DR	0.0186	6.419	9.497	3.304	1.702	10
12	H	7	25	27	110.76	4696.492	DR	0.0142	5.750	13.840	7.032	1.331	130
12	H	7	35	37	110.86	4702.556	DR	0.0105	4.920	15.889	7.194	0.941	119
12	H	7	45	47	110.96	4708.620	DR	0.0096	5.329	15.885	8.421	0.984	118
12	H	7	55	57	111.06	4714.684	DR	0.0122	5.455	16.051	8.736	1.050	116
12	H	C	5	7	111.16	4720.748	DR	0.0097	4.987	16.006	8.359	0.960	87
12	H	C	15	17	111.26	4726.812	DR	0.0083	3.870	17.895	10.87	0.705	111
13	H	1	5	7	113.56	4866.286	DR	0.0131	6.104	14.075	6.434	1.423	134
13	H	1	15	17	113.66	4872.350	DR	0.0111	4.501	15.436	8.047	1.277	132
13	H	1	25	27	113.76	4878.414	DR	0.0139	4.725	10.922	4.061	1.167	9
13	H	1	35	37	113.86	4884.478	DR	0.0139	5.112	14.362	7.017	0.944	134
13	H	1	45	47	113.96	4890.542	DR	0.0154	5.922	13.637	6.329	1.359	127
13	H	1	55	57	114.06	4896.162	DR	0.0109	5.177	14.708	7.479	1.178	117
13	H	1	65	67	114.16	4897.778	M	0.0111	6.374	13.669	6.795	1.791	114
13	H	1	75	77	114.26	4899.394	M	0.0109	7.048	14.211	8.008	1.151	106
13	H	1	85	87	114.36	4901.010	M	0.0117	6.301	13.318	6.993	1.250	166
13	H	1	95	97	114.46	4902.626	M	0.0123	5.478	11.515	4.164	1.374	9
13	H	1	105	107	114.56	4904.242	M	0.0113	6.181	13.792	6.769	1.309	111
13	H	1	115	117	114.66	4905.858	M	0.0142	6.784	13.067	5.999	1.918	128
13	H	1	125	127	114.76	4907.474	M	0.0082	6.646	15.543	7.851	1.004	138
13	H	1	135	137	114.86	4909.090	M	0.0107	7.060	12.321	5.194	1.657	127
13	H	1	145	147	114.96	4910.706	M	0.0087	6.720	13.795	5.395	1.961	138
13	H	2	5	7	115.06	4912.322	DR	0.0123	5.986	11.318	4.025	1.524	9
13	H	2	35	37	115.36	4917.170	DR	0.0194	8.934	9.948	4.920	1.980	140
13	H	2	43	45	115.44	4918.462	DB	0.0197	6.643	12.127	5.277	2.476	114
13	H	2	55	57	115.56	4920.402	DB	0.0220	7.253	11.059	4.566	2.714	166
13	H	2	65	67	115.66	4922.018	DB	0.0115	5.819	12.028	4.411	1.645	10
13	H	2	75	77	115.76	4923.634	M	0.0113	6.607	14.883	7.171	1.360	117
13	H	2	85	87	115.86	4925.250	M	0.0104	5.587	16.428	8.907	1.140	105

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
13	H	2	95	97	115.96	4926.866	M	0.0117	6.108	13.428	6.311	1.256	129
13	H	2	105	107	116.06	4928.482	M	0.0111	7.663	13.759	6.107	2.280	137
13	H	2	115	117	116.16	4930.098	M	0.0121	6.632	14.935	7.261	1.886	138
13	H	2	125	127	116.26	4931.714	M	0.0110	7.534	14.457	5.887	2.515	101
13	H	2	135	137	116.36	4933.330	M	0.0117	6.349	11.477	4.290	1.502	9
13	H	2	145	147	116.46	4934.946	M	0.0083	7.854	14.044	6.725	1.311	98
13	H	3	5	7	116.56	4936.562	DR	0.0147	5.209	11.880	5.593	1.427	136
13	H	3	16	18	116.67	4938.339	DR	0.0200	6.857	10.325	4.306	2.055	158
13	H	3	25	27	116.76	4939.794	DB	0.0191	5.951	13.577	5.901	1.525	112
13	H	3	35	37	116.86	4941.410	DB	0.0159	5.454	14.324	6.683	1.294	128
13	H	3	45	47	116.96	4943.026	DB	0.0127	4.786	15.147	7.531	1.022	127
13	H	3	55	57	117.06	4944.642	M	0.0102	4.475	12.602	4.677	0.966	7
13	H	3	65	67	117.16	4946.258	M	0.0105	4.892	17.317	5.619	1.062	161
13	H	3	75	77	117.26	4947.874	M	0.0087	5.967	15.794	8.016	1.643	114
13	H	3	85	87	117.36	4949.490	M	0.0127	8.444	12.014	8.226	1.183	127
13	H	3	95	97	117.46	4951.106	M	0.0085	4.290	16.584	9.016	0.832	120
13	H	3	105	107	117.56	4952.722	M	0.0095	8.244	17.180	12.33	1.101	172
13	H	3	115	117	117.66	4954.338	M	0.0078	8.443	15.491	7.539	1.179	94
13	H	3	125	127	117.76	4955.954	M	0.0085	8.686	12.983	4.705	1.662	9
13	H	3	135	137	117.86	4957.570	M	0.0083	3.890	19.321	10.19	0.858	105
13	H	3	145	147	117.96	4959.186	M	0.0078	4.391	17.611	8.885	0.920	144
13	H	4	6	8	118.07	4960.963	M	0.0090	4.666	17.359	8.829	1.027	113
13	H	4	14	16	118.15	4962.256	M	0.0108	7.490	13.262	6.632	1.647	107
13	H	4	25	27	118.26	4964.034	M	0.0085	4.005	16.855	8.829	1.058	110
13	H	4	35	37	118.36	4965.650	M	0.0088	3.951	17.801	9.354	0.805	131
13	H	4	45	47	118.46	4967.266	M	0.0080	3.858	15.337	5.849	0.799	7
13	H	4	55	57	118.56	4968.882	M	0.0093	7.666	12.753	5.714	2.135	114
13	H	4	65	67	118.66	4970.498	M	0.0074	6.891	14.765	7.043	1.037	101
13	H	4	75	77	118.76	4972.114	M	0.0085	4.409	17.000	6.858	0.861	121
13	H	4	85	87	118.86	4973.730	M	0.0084	4.232	18.083	9.487	0.307	114
13	H	4	95	97	118.96	4975.346	M	0.0086	4.464	17.366	9.049	0.921	108
13	H	4	105	107	119.06	4976.962	M	0.0103	6.318	14.501	5.344	1.271	10
13	H	4	115	117	119.16	4978.578	M	0.0074	3.815	17.582	9.688	0.795	108
13	H	4	135	137	119.36	4981.810	M	0.0069	3.911	17.598	9.375	0.990	116
13	H	4	145	147	119.46	4983.426	M	0.0077	4.455	17.811	9.534	0.798	103
13	H	5	5	7	119.56	4985.042	M	0.0085	6.031	15.827	7.246	1.248	145
13	H	5	15	17	119.66	4986.658	M	0.0077	6.283	15.216	7.069	1.131	112
13	H	5	25	27	119.76	4988.274	M	0.0105	4.965	14.700	7.215	1.125	124
13	H	5	35	37	119.86	4989.890	M	0.0102	5.021	14.292	7.309	0.358	129
13	H	5	45	47	119.96	4991.506	M	0.0097	6.256	14.747	7.207	1.147	151
13	H	5	55	57	120.06	4993.122	M	0.0172	5.614	9.502	3.329	1.437	11
13	H	5	65	67	120.16	4994.738	DR	0.0226	7.239	9.374	3.908	1.966	202
13	H	5	85	87	120.36	5001.263	DR	0.0251	7.258	10.188	4.430	1.847	119
13	H	5	95	97	120.46	5008.367	DR	0.0278	8.416	8.167	3.491	0.820	197
13	H	5	105	107	120.56	5015.472	DR	0.0224	8.006	9.981	3.821	1.875	161
13	H	5	115	117	120.66	5022.576	DR	0.0180	7.056	7.826	2.507	2.726	14
13	H	5	125	127	120.76	5029.681	DR	0.0240	7.358	11.335	4.776	2.455	112
13	H	5	135	137	120.86	5036.785	DR	0.0202	6.362	11.231	5.114	1.760	156

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
13	H	5	145	147	120.96	5043.890	DR	0.0206	5.595	10.939	5.062	0.799	155
13	H	6	6	8	121.07	5051.704	DR	0.0176	5.553	12.152	5.470	1.492	139
13	H	6	15	17	121.16	5058.099	DR	0.0153	5.414	13.098	5.767	1.266	149
13	H	6	25	27	121.26	5065.203	DR	0.0178	5.373	12.776	5.879	1.200	136
13	H	6	35	37	121.36	5072.307	DR	0.0229	6.682	7.454	2.638	1.975	13
13	H	6	45	47	121.46	5079.412	DR	0.0205	6.074	9.870	4.353	0.732	192
13	H	6	55	57	121.56	5086.516	DR	0.0216	6.417	10.975	4.856	1.882	163
13	H	6	65	67	121.66	5093.621	DR	0.0206	6.876	10.856	4.682	2.123	172
13	H	6	85	87	121.86	5107.830	DR	0.0190	5.781	12.526	5.921	1.432	135
13	H	6	95	97	121.96	5114.934	DR	0.0148	5.066	14.760	6.797	1.208	127
13	H	6	105	107	122.06	5122.039	DR	0.0174	5.739	13.223	6.115	0.829	138
13	H	6	115	117	122.16	5129.143	DR	0.0178	5.593	11.455	4.125	1.613	9
13	H	6	125	127	122.26	5136.248	DR	0.0070	4.035	14.409	5.005	1.859	148
13	H	6	132	134	122.33	5141.221	DR	0.0205	6.169	11.796	5.358	1.693	142
13	H	c	5	7	122.40	5146.194	DR	0.0178	5.882	13.294	6.190	1.316	141
13	H	c	15	17	122.50	5153.299	DR	0.0093	4.216	17.218	9.236	0.491	116
14	H	1	5	7	123.06	5193.084	DR	0.0087	4.703	15.416	8.056	0.854	149
14	H	1	15	17	123.16	5200.188	DR	0.0083	4.868	16.395	7.770	0.874	109
14	H	1	25	27	123.26	5207.293	DR	0.0143	4.790	11.724	3.850	1.187	10
14	H	1	35	37	123.36	5214.397	DR	0.0200	6.147	11.852	5.192	1.987	144
14	H	1	45	47	123.46	5221.501	DR	0.0211	6.967	11.387	4.735	1.772	184
14	H	1	55	57	123.56	5228.606	DR	0.0193	6.406	11.761	5.057	1.623	149
14	H	1	85	87	123.86	5243.317	DR	0.0240	6.836	9.223	3.795	2.283	244
14	H	1	95	97	123.96	5247.277	DR	0.0161	5.619	9.882	3.135	1.807	11
14	H	1	105	107	124.06	5251.237	DR	0.0209	6.737	9.768	4.193	2.149	182
14	H	1	115	117	124.16	5255.198	DR	0.0131	5.040	15.521	7.616	1.174	121
14	H	1	124	126	124.25	5258.762	DR	0.0092	5.290	14.306	7.148	1.113	152
14	H	1	135	137	124.36	5263.118	M	0.0126	4.615	16.734	8.590	1.347	131
14	H	1	145	147	124.46	5267.078	M	0.0083	3.928	18.973	10.97	0.827	110
14	H	2	5	7	124.56	5271.039	M	0.0104	4.168	16.896	7.943	0.871	120
14	H	2	15	17	124.66	5274.999	M	0.0090	5.013	15.956	8.471	0.876	136
14	H	2	25	27	124.76	5278.959	M	0.0091	3.811	14.729	5.547	0.708	8
14	H	2	35	37	124.86	5282.920	M	0.0076	5.991	15.979	8.747	0.947	103
14	H	2	45	47	124.96	5286.880	M	0.0080	5.052	16.057	7.926	0.801	105
14	H	2	57	59	125.08	5291.632	M	0.0078	5.936	16.220	8.319	1.005	110
14	H	2	65	67	125.16	5294.800	M	0.0059	3.708	20.766	9.276	0.996	103
14	H	2	75	77	125.26	5298.761	M	0.0082	5.965	14.130	7.239	1.121	142
14	H	2	95	97	125.46	5306.681	M	0.0079	5.527	15.178	6.853	1.010	111
14	H	2	105	107	125.56	5310.642	M	0.0079	5.573	12.049	4.337	0.991	8
14	H	2	115	117	125.66	5314.602	M	0.0078	6.792	14.584	6.343	1.350	121
14	H	2	125	127	125.76	5318.562	M	0.0138	6.546	11.201	4.473	1.621	187
14	H	2	135	137	125.86	5322.523	DR	0.0245	8.603	9.644	3.789	2.376	156
14	H	2	145	147	125.96	5326.483	DR	0.0163	6.823	11.048	4.529	1.711	143
14	H	3	6	8	126.07	5330.839	DR	0.0215	6.625	10.346	4.260	2.201	168
14	H	3	15	17	126.16	5334.403	DR	0.0206	5.895	11.399	4.747	1.526	198
14	H	3	25	27	126.26	5338.364	DR	0.0205	6.010	9.405	3.138	1.712	11
14	H	3	35	37	126.36	5342.324	DR	0.0231	8.427	9.414	4.022	2.107	158
14	H	3	45	47	126.46	5346.284	DR	0.0096	6.795	14.324	6.350	2.331	123

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
14	H	3	55	57	126.56	5350.245	M	0.0078	6.019	13.989	6.836	1.123	133
14	H	3	64	66	126.65	5353.809	M	0.0094	6.922	13.748	6.623	1.471	108
14	H	3	75	77	126.76	5358.165	M	0.0096	4.941	15.631	7.313	1.211	119
14	H	3	85	87	126.86	5362.126	M	0.0090	4.894	12.005	4.046	1.336	10
14	H	3	95	97	126.96	5366.086	M	0.0099	4.399	14.879	7.213	1.171	163
14	H	3	105	107	127.06	5370.046	DR	0.0124	5.351	12.973	5.549	1.586	145
14	H	3	115	117	127.16	5374.006	DR	0.0190	6.833	11.261	4.941	2.008	156
14	H	3	125	127	127.26	5377.967	DR	0.0058	3.858	15.077	4.488	2.006	158
14	H	3	135	137	127.36	5381.927	DR	0.0235	8.644	9.480	4.067	2.443	204
14	H	3	145	147	127.46	5385.887	DR	0.0214	8.801	10.709	4.888	1.829	145
14	H	4	5	7	127.56	5389.848	DB	0.0187	8.004	11.837	5.063	2.743	146
14	H	4	15	17	127.66	5393.808	DB	0.0129	6.334	10.962	3.787	2.393	11
14	H	4	25	27	127.76	5397.768	DB	0.0115	5.856	14.082	6.076	1.638	176
14	H	4	35	37	127.86	5401.729	M	0.0099	5.319	14.593	7.242	1.532	116
14	H	4	45	47	127.96	5405.689	M	0.0088	7.335	15.196	7.945	1.194	106
14	H	4	55	57	128.06	5409.649	M	0.0083	6.414	15.224	7.000	1.231	100
14	H	4	65	67	128.16	5413.609	M	0.0117	5.145	13.558	5.827	1.295	161
14	H	4	75	77	128.26	5417.570	M	0.0100	8.180	12.902	5.576	1.743	114
14	H	4	85	87	128.36	5421.530	M	0.0096	6.119	11.301	3.994	1.703	10
14	H	4	95	97	128.46	5425.490	M	0.0119	5.848	15.286	6.683	1.800	151
14	H	4	105	107	128.56	5429.451	M	0.0121	6.211	14.656	6.077	2.098	197
14	H	4	115	117	128.66	5433.411	M	0.0086	6.095	15.371	6.880	1.442	113
14	H	4	125	127	128.76	5437.371	M	0.0103	4.935	16.629	7.699	2.578	128
14	H	4	135	137	128.86	5441.332	DR	0.0087	5.864	12.723	5.424	1.669	134
14	H	4	145	147	128.96	5445.292	DR	0.0151	8.034	10.740	4.351	2.525	188
14	H	5	5	7	129.06	5449.252	DB	0.0144	6.905	8.966	3.064	2.366	10
14	H	5	15	17	129.16	5453.212	DB	0.0099	5.910	14.599	6.074	2.060	154
14	H	5	25	27	129.26	5457.173	DB	0.0111	6.330	14.899	6.910	1.513	125
14	H	5	35	37	129.36	5461.133	M	0.0121	6.111	12.956	6.178	1.666	184
14	H	5	45	47	129.46	5465.093	M	0.0070	5.390	15.319	7.923	1.453	119
14	H	5	55	57	129.56	5469.054	M	0.0065	5.151	15.742	8.265	1.008	112
14	H	5	65	67	129.66	5473.014	M	0.0090	5.936	13.532	5.488	1.495	134
14	H	5	75	77	129.76	5476.974	M	0.0121	5.108	12.104	5.263	1.505	193
14	H	5	85	87	129.86	5480.934	DR	0.0112	5.611	10.496	3.419	1.549	9
14	H	5	95	97	129.96	5484.895	DR	0.0208	6.612	9.746	4.006	2.356	191
14	H	5	105	107	130.06	5488.855	DR	0.0167	6.627	11.780	5.043	2.042	167
14	H	5	115	117	130.16	5492.815	DB	0.0161	6.356	11.073	4.472	2.142	212
14	H	5	125	127	130.26	5496.776	DB	0.0164	6.078	12.590	5.241	1.844	151
14	H	5	135	137	130.36	5500.736	DR	0.0145	6.414	11.146	5.037	2.309	157
14	H	5	145	147	130.46	5504.696	DR	0.0096	6.146	12.056	4.139	2.078	11
14	H	6	5	7	130.56	5508.657	DB	0.0133	6.029	11.919	5.588	1.602	187
14	H	6	15	17	130.66	5512.617	DB	0.0062	4.266	18.233	6.918	1.350	116
14	H	6	25	27	130.76	5516.577	M	0.0086	6.546	14.874	7.266	1.215	99
14	H	6	35	37	130.86	5520.537	M	0.0096	7.743	14.681	6.964	1.633	112
14	H	6	45	47	130.96	5524.498	M	0.0096	7.577	13.570	5.757	1.693	149
14	H	6	55	57	131.06	5528.458	M	0.0132	6.292	13.657	5.585	2.441	167
14	H	6	65	67	131.16	5532.418	M	0.0092	5.447	11.351	4.047	1.234	9
14	H	6	75	77	131.26	5536.379	M	0.0135	6.347	13.096	5.379	2.532	152

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
14	H	6	85	87	131.36	5540.339	M	0.0121	5.165	13.411	5.996	1.410	186
14	H	6	95	97	131.46	5544.299	M	0.0093	6.345	13.144	5.881	1.187	149
14	H	6	105	107	131.56	5548.260	M	0.0088	5.531	14.657	7.242	1.244	123
14	H	6	115	117	131.66	5552.220	M	0.0076	5.554	15.458	7.774	1.120	127
14	H	6	125	127	131.76	5556.180	DB	0.0183	8.181	10.326	4.567	2.074	203
14	H	7	5	7	132.06	5560.140	DB	0.0259	7.181	8.216	3.333	2.212	182
14	H	7	15	17	132.06	5564.101	DB	0.0236	6.650	9.490	3.880	2.039	175
14	H	7	25	27	132.16	5568.061	DB	0.0242	7.377	8.076	2.634	2.091	13
14	H	7	35	37	132.36	5572.021	DB	0.0240	7.106	8.706	3.398	2.120	239
14	H	7	45	47	132.46	5575.982	DB	0.0259	8.815	9.005	3.613	2.770	163
14	H	7	55	57	132.56	5579.942	DR	0.0248	7.995	9.048	3.796	2.226	167
15	H	1	15	17	132.66	5591.823	DR	0.0130	7.866	15.322	5.891	3.196	174
15	H	1	25	27	132.76	5595.783	DB	0.0112	6.274	14.207	6.475	1.822	137
15	H	1	35	37	132.86	5599.743	DB	0.0095	6.271	13.789	5.780	1.870	134
15	H	1	45	47	132.98	5603.704	DB	0.0117	5.590	14.258	5.881	1.682	139
15	H	1	55	57	133.06	5607.664	DB	0.0115	5.999	13.066	5.355	1.728	130
15	H	1	65	67	133.16	5611.624	DB	0.0112	6.045	13.735	7.205	1.473	124
15	H	1	75	77	133.26	5615.585	DB	0.0113	6.075	13.705	6.374	1.552	124
15	H	1	85	87	133.36	5619.545	M	0.0066	4.116	18.603	7.331	1.443	106
15	H	1	95	97	133.46	5623.505	M	0.0093	5.441	13.964	5.829	1.224	125
15	H	1	105	107	133.56	5627.466	M	0.0097	6.488	15.367	6.485	1.449	101
15	H	1	115	117	133.66	5631.426	DR	0.0145	8.692	11.462	4.624	1.787	127
15	H	1	125	127	133.76	5635.386	DR	0.0160	6.481	12.106	4.648	1.595	142
15	H	1	135	137	133.86	5639.346	DR	0.0179	6.115	13.030	5.676	1.229	119
15	H	1	145	147	133.98	5643.307	DR	0.0173	6.022	12.753	5.375	1.293	127
15	H	2	15	17	134.16	5651.227	DR	0.0195	7.022	10.915	4.190	1.894	151
15	H	2	25	27	134.26	5655.188	DR	0.0187	7.178	11.824	4.995	1.529	137
15	H	2	35	37	134.36	5659.148	DR	0.0165	6.013	12.990	5.787	1.727	136
15	H	2	45	47	134.46	5663.108	DR	0.0122	6.245	13.785	5.710	2.690	165
15	H	2	55	57	134.56	5667.068	DR	0.0150	5.857	13.503	4.961	1.489	133
15	H	2	65	67	134.66	5671.029	DR	0.0165	5.764	12.841	5.396	1.435	147
15	H	2	75	77	134.76	5674.989	DB	0.0155	5.708	13.156	5.797	1.522	141
15	H	2	85	87	134.88	5678.949	DB	0.0121	5.407	14.266	6.134	1.266	131
15	H	2	105	107	135.06	5686.870	DB	0.0097	5.756	16.081	6.429	1.164	117
15	H	2	115	117	135.16	5690.830	DB	0.0090	4.480	15.556	8.140	1.248	128
15	H	2	123	125	135.24	5693.999	DB	0.0075	5.343	13.747	8.704	0.896	116
15	H	2	135	137	135.36	5698.751	M	0.0073	5.437	14.623	6.834	1.139	114
15	H	2	145	147	135.46	5702.711	M	0.0093	4.594	16.541	6.750	1.009	124
15	H	3	5	7	135.56	5706.671	M	0.0066	4.158	16.518	6.903	1.391	126
15	H	3	15	17	135.68	5710.632	M	0.0063	3.927	15.778	6.233	0.958	139
15	H	3	25	27	135.76	5714.592	M	0.0080	4.362	15.141	7.618	1.022	139
15	H	3	35	37	135.86	5718.552	M	0.0089	5.232	14.366	5.202	1.589	143
15	H	3	45	47	135.96	5722.513	M	0.0073	4.411	14.829	7.435	1.005	130
15	H	3	55	57	136.06	5726.473	M	0.0098	5.766	13.112	5.940	1.507	128
15	H	3	65	67	136.16	5730.433	M	0.0078	5.694	14.707	6.526	1.180	126
15	H	3	75	77	136.26	5734.394	M	0.0099	5.460	14.493	5.504	1.614	133
15	H	3	85	87	136.36	5738.354	M	0.0104	7.939	15.183	6.475	2.070	160
15	H	3	95	97	136.46	5742.314	M	0.0100	5.108	14.397	6.779	1.159	126

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
15	H	3	105	107	136.58	5746.274	M	0.0108	5.174	14.493	5.983	1.322	133
15	H	3	115	117	136.66	5750.235	M	0.0097	5.471	14.756	5.917	1.260	124
15	H	3	125	127	136.76	5754.195	M	0.0100	5.417	14.786	6.963	1.489	127
15	H	3	137	139	136.88	5758.947	M	0.0082	6.396	14.354	6.226	1.434	111
15	H	3	145	147	136.96	5762.116	M	0.0098	5.564	14.340	6.842	1.183	121
15	H	4	5	7	137.06	5766.076	M	0.0087	5.182	14.736	5.500	1.231	130
15	H	4	15	17	137.16	5770.036	M	0.0070	4.231	18.251	6.739	1.481	119
15	H	4	25	27	137.26	5773.997	M	0.0078	5.616	14.492	6.770	1.275	115
15	H	4	35	37	137.38	5777.957	M	0.0098	5.967	14.590	6.094	1.258	112
15	H	4	45	47	137.46	5781.917	M	0.0122	5.353	13.394	4.842	1.815	146
15	H	4	55	57	137.56	5785.877	M	0.0205	7.986	8.968	3.635	2.358	168
15	H	4	65	67	137.66	5789.838	DR	0.0251	8.950	9.447	3.994	2.483	148
15	H	4	85	87	137.86	5797.758	DR	0.0229	7.150	9.893	3.718	1.838	164
15	H	4	95	97	137.96	5801.719	DB	0.0188	6.821	11.730	5.014	1.592	128
15	H	4	105	107	138.06	5805.679	DB	0.0136	5.830	12.445	5.819	1.324	140
15	H	4	115	117	138.16	5809.639	DB	0.0137	5.507	13.521	6.409	1.308	128
15	H	4	125	127	138.26	5813.600	DB	0.0109	7.167	14.348	5.931	1.416	128
15	H	4	135	137	138.36	5817.560	M	0.0083	7.174	14.878	6.858	1.302	104
15	H	4	145	147	138.48	5821.520	M	0.0133	6.299	13.175	5.519	1.466	133
15	H	5	5	7	138.56	5825.480	M	0.0090	6.433	14.530	6.965	1.306	114
15	H	5	15	17	138.66	5829.441	M	0.0091	6.484	14.843	6.933	1.263	105
15	H	5	25	27	138.76	5833.401	DB	0.0129	5.480	13.617	4.825	1.962	179
15	H	5	35	37	138.86	5837.361	DB	0.0154	5.762	13.389	5.748	2.002	161
15	H	5	45	47	138.96	5841.322	DB	0.0179	5.851	11.915	5.194	1.619	150
15	H	5	55	57	139.06	5845.282	DR	0.0189	5.967	10.607	4.021	1.792	155
15	H	5	65	67	139.16	5849.242	DR	0.0195	6.239	10.813	4.542	1.870	165
15	H	5	75	77	139.26	5853.202	DR	0.0193	6.220	11.496	4.860	1.619	130
15	H	5	85	87	139.38	5857.163	DR	0.0209	6.383	10.677	4.602	1.429	154
15	H	5	95	97	139.46	5861.123	DR	0.0214	6.501	9.948	3.728	1.837	164
15	H	5	105	107	139.56	5865.083	DR	0.0176	5.839	11.885	5.100	1.830	146
15	H	5	115	117	139.66	5869.044	DB	0.0127	4.663	14.849	7.510	1.095	126
15	H	5	125	127	139.76	5873.004	DB	0.0098	4.855	15.553	7.187	1.060	133
15	H	5	135	137	139.86	5876.964	M	0.0089	5.034	15.938	6.447	0.938	113
15	H	5	145	147	139.96	5880.925	M	0.0088	4.342	16.735	9.034	0.936	122
15	H	6	15	17	140.16	5888.845	M	0.0092	4.582	16.968	9.022	0.811	123
15	H	6	25	27	140.26	5892.805	M	0.0074	5.557	14.946	6.160	0.987	121
15	H	6	35	37	140.38	5896.766	M	0.0099	4.702	15.396	7.176	0.927	130
15	H	6	45	47	140.46	5900.726	M	0.0086	4.640	17.128	7.979	0.957	120
15	H	6	55	57	140.56	5904.686	M	0.0080	3.849	17.328	9.379	0.854	113
15	H	6	65	67	140.66	5908.647	M	0.0078	3.828	17.439	6.892	0.870	123
15	H	6	75	77	140.76	5912.607	M	0.0072	3.767	16.907	9.125	0.800	117
15	H	6	85	87	140.86	5916.567	M	0.0072	3.743	17.589	9.658	0.771	116
15	H	6	95	97	140.96	5920.528	M	0.0075	4.225	16.288	8.469	0.908	129
15	H	6	105	107	141.06	5924.488	M	0.0081	7.186	15.108	6.011	1.297	106
15	H	6	115	117	141.16	5928.448	M	0.0079	6.560	15.131	7.068	1.198	112
15	H	6	125	127	141.26	5932.408	M	0.0091	6.547	14.777	7.794	1.003	108
15	H	7	5	7	141.36	5936.369	DR	0.0081	8.154	14.747	6.866	1.153	86
15	H	7	15	17	141.46	5940.329	DR	0.0076	7.605	15.963	6.794	0.959	95

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
15	H	7	25	27	141.56	5944.289	DR	0.0076	4.985	15.128	6.615	1.008	131
15	H	7	35	37	141.66	5948.250	DR	0.0074	5.797	15.590	7.236	1.128	110
16	H	1	5	7	142.06	5964.091	DR	0.0143	6.178	12.231	5.458	1.287	145
16	H	1	15	17	142.16	5968.051	DR	0.0134	6.575	12.485	5.686	1.553	144
16	H	1	25	27	142.26	5972.011	DR	0.0136	8.297	9.941	4.581	4.922	164
16	H	1	35	37	142.36	5975.972	DR	0.0235	8.436	9.476	3.985	1.942	155
16	H	1	45	47	142.46	5979.932	DR	0.0238	8.801	8.798	3.660	1.912	162
16	H	1	65	67	142.66	5987.853	DR	0.0115	6.070	12.503	5.319	1.590	136
16	H	1	75	77	142.78	5991.813	DR	0.0231	10.280	7.743	3.063	2.080	169
16	H	1	85	87	142.86	5995.773	DR	0.0256	11.378	7.691	3.239	2.160	159
16	H	1	95	97	142.96	5999.733	DR	0.0262	10.730	8.305	3.491	2.192	152
16	H	1	105	107	143.06	6003.694	DR	0.0225	9.415	9.239	3.923	2.231	158
16	H	1	115	117	143.16	6007.654	DR	0.0173	9.540	8.744	5.664	2.358	145
16	H	1	125	127	143.26	6011.614	DR	0.0175	6.973	11.477	5.103	1.427	141
16	H	1	145	147	143.46	6019.535	DR	0.0175	6.776	12.095	5.617	1.147	126
16	H	2	5	7	143.58	6023.495	DR	0.0238	10.331	9.304	4.068	1.736	148
16	H	2	15	17	143.66	6027.456	DR	0.0240	11.922	7.527	3.359	2.368	172
16	H	2	25	27	143.76	6031.416	DR	0.0206	9.860	8.723	3.383	1.957	162
16	H	2	35	37	143.86	6034.314	DR	0.0161	7.176	11.099	4.940	1.526	153
16	H	2	45	47	143.96	6036.504	DR	0.0141	6.134	12.650	5.884	1.356	141
16	H	2	55	57	144.06	6038.694	DR	0.0186	7.279	10.652	4.710	1.540	161
16	H	2	65	67	144.16	6040.884	DR	0.0215	7.449	9.864	4.481	1.695	145
16	H	2	75	77	144.26	6043.074	DR	0.0201	7.968	10.329	3.580	1.788	149
16	H	2	85	87	144.36	6045.264	DR	0.0123	2.390	19.620	3.263	2.106	157
16	H	2	95	97	144.46	6047.454	DR	0.0245	10.790	7.212	3.006	2.298	190
16	H	2	105	107	144.56	6049.644	DR	0.0245	11.666	8.538	3.569	2.005	149
16	H	2	115	117	144.68	6051.834	DR	0.0221	10.467	8.818	3.772	2.272	153
16	H	2	125	127	144.76	6054.024	DR	0.0216	9.079	8.758	3.952	2.031	171
16	H	2	135	137	144.86	6056.214	DR	0.0175	6.989	11.245	5.024	1.333	135
16	H	2	145	147	144.96	6058.404	DR	0.0169	7.424	11.932	5.462	1.478	117
16	H	3	5	7	145.06	6060.594	DR	0.0202	9.524	9.418	4.160	1.963	136
16	H	3	15	17	145.16	6062.784	DR	0.0133	8.663	11.559	4.938	2.608	138
16	H	3	25	27	145.26	6064.974	M	0.0087	5.479	14.318	7.500	1.310	129
16	H	3	35	37	145.38	6067.164	M	0.0091	5.976	15.030	6.740	1.182	130
16	H	3	45	47	145.46	6069.354	M	0.0097	5.013	16.132	8.996	0.952	136
16	H	3	55	57	145.58	6071.544	M	0.0080	5.647	15.437	7.264	0.972	123
16	H	3	65	67	145.66	6073.734	M	0.0093	6.294	15.001	8.430	1.172	109
16	H	3	85	87	145.88	6078.114	M	0.0080	5.120	15.640	7.411	0.853	121
16	H	3	95	97	145.96	6080.304	M	0.0090	5.328	15.441	6.976	1.172	123
16	H	3	105	107	146.06	6082.494	M	0.0122	8.738	10.846	4.374	9.309	193
16	H	3	115	117	146.16	6084.684	M	0.0086	5.039	14.981	7.390	1.219	135
16	H	3	125	127	146.26	6086.874	M	0.0079	5.703	15.456	7.380	1.060	104
16	H	3	135	137	146.38	6089.064	M	0.0078	3.730	16.308	7.299	0.852	126
16	H	3	145	147	146.46	6091.254	M	0.0085	4.693	15.896	7.960	1.053	134
16	H	4	5	7	146.58	6093.444	M	0.0081	5.383	16.904	7.375	1.187	103
16	H	4	15	17	146.66	6095.634	M	0.0101	5.300	14.111	5.964	1.325	135
16	H	4	25	27	146.76	6097.824	M	0.0108	7.310	14.316	6.435	1.937	154
16	H	4	35	37	146.86	6100.014	M	0.0109	5.972	13.951	6.304	2.329	164

Core, Section, Interval (cm)					Depth (mbsf)	Age (ka)	Facies Code	Ba/Al	Si/Al	Al/Ti	K/Ti	Zr/Rb	Sr
16	H	4	45	47	146.96	6102.204	DR	0.0246	9.596	9.049	3.810	2.157	166
16	H	4	55	57	147.06	6104.394	DR	0.0007	4.531	21.842	7.125	1.646	140
16	H	4	65	67	147.16	6106.584	DR	0.0128	6.291	13.531	6.124	1.595	138
16	H	4	75	77	147.26	6108.774	DR	0.0120	6.730	12.401	5.794	2.036	145
16	H	4	85	87	147.36	6110.964	DR	0.0135	7.100	12.824	5.517	1.896	139
16	H	4	95	97	147.46	6113.154	DR	0.0186	8.448	9.416	3.646	1.883	172
16	H	4	105	107	147.58	6115.344	DR	0.0129	6.754	13.948	6.018	1.535	141
16	H	4	115	117	147.66	6117.534	DR	0.0110	7.464	14.157	6.822	1.911	146
16	H	4	125	127	147.76	6119.724	DR	0.0109	6.002	14.283	6.660	2.125	155
16	H	4	135	137	147.86	6121.914	DR	0.0117	7.428	16.085	7.011	2.231	166
16	H	4	145	147	147.96	6124.104	DR	0.0217	9.006	8.574	3.656	2.372	175
16	H	5	5	7	148.06	6126.294	DR	0.0130	6.465	12.515	5.724	1.386	135
16	H	5	15	17	148.16	6128.484	DR	0.0099	7.146	13.384	6.020	1.366	111
16	H	5	25	27	148.26	6130.674	DR	0.0199	8.406	9.336	4.068	1.590	142
16	H	5	35	37	148.38	6132.864	DR	0.0219	8.893	8.252	3.310	1.860	170
16	H	5	45	47	148.46	6135.054	DR	0.0236	10.122	8.012	3.172	2.131	157
16	H	5	55	57	148.56	6137.244	DR	0.0219	8.394	10.756	4.340	1.879	134
16	H	5	65	67	148.66	6139.434	DR	0.0030	2.851	22.664	3.937	3.017	138
16	H	5	75	77	148.76	6141.624	DR	0.0228	9.284	8.928	3.600	2.598	163
16	H	5	85	87	148.86	6143.814	DR	0.0160	8.522	9.324	3.860	4.770	169
16	H	5	95	97	148.96	6146.004	DR	0.0224	7.779	9.606	4.098	2.304	161
16	H	5	105	107	149.06	6148.194	DR	0.0211	8.183	10.057	4.443	1.670	150
16	H	5	115	117	149.16	6150.384	DR	0.0202	8.598	9.594	3.933	2.520	157
16	H	5	125	127	149.26	6152.574	DR	0.0162	8.030	9.892	4.163	2.758	171
16	H	5	135	137	149.38	6154.764	DR	0.0214	8.004	9.855	3.870	2.309	161
16	H	5	145	147	149.46	6156.954	DR	0.0219	8.515	9.720	3.894	2.160	163
16	H	6	5	7	149.56	6159.144	DR	0.0131	5.002	15.626	4.879	1.254	131
16	H	6	15	17	149.66	6161.334	DR	0.0212	8.064	8.655	3.533	2.515	181
16	H	6	25	27	149.76	6163.524	DR	0.0227	8.455	9.565	3.920	2.511	158
16	H	6	55	57	150.06	6170.094	DR	0.0205	7.810	10.945	5.173	1.406	129
16	H	6	65	67	150.16	6172.284	DR	0.0111	6.567	13.995	7.923	1.704	169
16	H	6	75	77	150.26	6174.474	DR	0.0240	8.926	9.420	4.182	1.908	157
16	H	6	85	87	150.38	6176.664	DR	0.0208	8.407	10.171	4.345	2.054	145
16	H	6	95	97	150.46	6178.854	DB	0.0190	8.256	9.910	4.603	1.900	149
16	H	6	106	108	150.57	6181.263	DB	0.0129	5.789	13.784	7.003	1.455	131
16	H	7	5	7	150.68	6183.672	DB	0.0088	5.244	14.695	7.956	1.281	132
16	H	7	15	17	150.78	6185.862	DB	0.0090	6.001	14.984	8.120	1.049	123
16	H	7	25	27	150.88	6188.052	M	0.0088	4.713	17.154	9.509	0.896	122
16	H	7	35	37	150.98	6190.242	M	0.0086	5.181	16.468	7.435	1.053	110
16	H	7	45	47	151.08	6192.432	M	0.0128	9.097	10.271	4.210	5.674	170
16	H	7	55	57	151.18	6194.622	M	0.0120	6.693	14.552	6.474	0.927	115
16	H	7	65	67	151.28	6196.812	M	0.0095	5.541	13.061	6.548	1.400	133
16	H	7	75	77	151.38	6199.002	M	0.0094	5.173	15.529	6.870	1.088	140
16	H	7	84	86	151.47	6200.973	M	0.0076	5.993	15.030	6.929	1.156	106
16	H	c	5	7	151.54	6202.506	M	0.0082	5.311	15.245	6.595	1.135	114
16	H	c	15	17	151.64	6204.696	M	0.0086	4.557	16.102	8.827	0.864	125