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A Sedimentological Analysis of the Siliciclastic Fraction in Pliocene Core Sediments from Bowers Ridge, Bering Sea (IODP Exp 323)

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A SEDIMENTOLOGICAL ANALYSIS OF THE SILICICLASTIC FRACTION IN
PLIOCENE CORE SEDIMENTS FROM BOWERS RIDGE, BERING SEA
(IODP EXP 323)

A Thesis

Presented to

The Faculty of the Department of Marine Science

San José State University

In Partial Fulfillment

of the Requirements for the Degree

Master of Science

by

Melinda J. Tanner

May 2016

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The Designated Thesis Committee Approves the Thesis Titled
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ABSTRACT

A SEDIMENTOLOGICAL ANALYSIS OF THE SILICICLASTIC FRACTION IN PLIOCENE CORE SEDIMENTS FROM BOWERS RIDGE, BERING SEA (IODP EXP 323)

by Melinda J. Tanner

The analysis and interpretation of sediment cores constitutes an important data record that can be used to evaluate paleoclimatic conditions and changes over time. The causes for the cooling trend that began in the Eocene, and its intensification with the onset of North Hemisphere Glaciation (NHG), remain under debate. Using sediment core samples from Site U1341 in the Bering Sea, the principal goal of this thesis was to examine the record at 1.9-4.2 Ma, using sedimentologic tools to interpret the siliciclastic fraction and ascertain whether and which portion is due to ice-related input. Previous research estimates the onset to begin at 2.75 Ma by using IRD occurrence and a sharp decline in opal accumulation (Haug et al., 2005). A reevaluation of this critical period will be important for gaining a better understanding of the onset of NHG. In this study, over 200 samples from 1.9-4.2 Ma were used to study the onset of NHG. The $> 250 \mu\text{m}$ siliciclastic mass accumulation rates (MAR) for the samples indicate an initial increase in abundance between 3.3 and 3.6 Ma before decreasing between 2.7 and 3.2 Ma. Another increase in abundance was observed between 2.7 Ma and the end of the record at 1.9 Ma. The provenance, composition, and surface texture features of quartz were also analyzed. It is uncertain whether the first occurrence of $> 250 \mu\text{m}$ siliciclastics was the result of ice-related processes or from other mechanisms like local re-sedimentation from Bowers Ridge. The implications of both possibilities were evaluated.

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The Bering Sea

An important opportunity to study the paleoclimate and evolution of the cryosphere in the Northern Hemisphere came in 2009 with International Ocean Discovery Program (IODP)'s Expedition 323 to the Bering Sea. The long intact sediment cores that were recovered from different parts of this marginal basin contain a 5 million year (my) record of the oceanographic and climatic conditions of the region, including such major events as onset of North Hemisphere Glaciation (NHG), the mid-Pleistocene transition (MPT), and glacial-interglacial cycles. According to this record, the Northern Hemisphere experienced a significant change in climate at the end of the Pliocene with the onset of NHG. Ice sheets began to form, expand, and persist, sea surface temperatures decreased, the thermohaline circulation weakened, the concentration of atmospheric carbon dioxide decreased, sea level lowered, and the ocean became more stratified (Ravelo et al., 2004). The cause for these changes remains under debate. The shift in global climate variability that occurred during MPT, 1.2 to 0.7 million years from the 40 thousand year (ky) cycles that dominated the Pliocene and early Pleistocene to a 100 ky cycle, can still not be fully explained (Hönisch et al., 2009) and alternative hypotheses concerning the causes of this shift have been formulated (Clark et al., 2006; Raymo and Huybers, 2008). During this time, the Northern Hemisphere experienced major climatic events in the last 5 my. Evidence of those events is preserved in the sediment record in the marine cores recovered by IODP Exp 323. Understanding the mechanisms and the dynamics responsible for significant variations in climate, both regionally and globally, is

becoming increasingly relevant in light of current projections for present and future climate change. Research into past changes in climate and ocean conditions is critical in further elucidating that system.

The geography of the Bering Sea and the surrounding land mass play an important role in both atmospheric circulation, biological productivity, and ocean circulation. The capacity of marginal seas in sub-polar regions, like the Bering Sea, to absorb atmospheric carbon dioxide allows these areas to have a substantial impact on the global carbon cycle, and therefore, on changes in global climate as well (Takahashi, 1998). The Bering Sea also influences regional climate through sea ice, which affects climate through albedo, stratification, and evaporation, and is a critical control in biological productivity and ice-related blooms (Smith et al., 2003; Katsuki and Takahashi, 2005; Caissie et al., 2010). A shallow (0-200 m) neritic area takes up about one half of the Bering Sea and today is seasonally covered by sea ice. In the deeper Aleutian Basin, however, sea ice is absent (Hood, 1983; Niebauer et al., 1999; Takahashi, 2005). Along the continental shelf is a highly productive region, in which Springer et al. (1996) estimate primary production to be 175 to 275 g C/m² yr or 260% higher than the other areas in the ocean. The Bering Sea is connected to the North Pacific by the Kamchatka Strait and to the Arctic Sea through the Bering Strait, and subsequently to the North Atlantic and is critical in ocean circulation (Takahashi, 1998; Takahashi, 2005). The exchange of surface water between the Bering Sea and the North Pacific through the Aleutian Islands is considerable and allows the Bering Sea to play a role in the conditions of the North Pacific (Aiello and Ravelo, 2012). The Bering Sea also affects the strength

of North Pacific Intermediate Water (NPIW), which forms in the Okhotsk Sea, by varying intermediate water ventilation that occurs during glacials and interglacials (Kim et al., 2011). Thus, the Bering Sea, and the sediments that it bears, is an important region for studying major changes in climate and oceanographic conditions.

1.1 Glacials vs Interglacials

There are significant differences in surface water circulation, sea level, extent of sea ice, and sedimentation of the Bering Sea between glacial periods and interglacials. During glacials, sea ice formation is more extensive and nearly perennial, which results in a drop in sea level and the closure of Unimak Pass and the Bering Strait (Katsuki and Takahashi, 2005). This diminishes both the inflow of North Pacific water into the Bering Sea and Bering Sea water into the Arctic. Biogenic sedimentation is also lower during glacials than interglacials, and the source of terrigenous sediment changes. During glacials, older rocks ranging from 40-80 Ma (based on ^{40}Ar - ^{39}Ar data) from the Yukon River basin and northeast Russia are the dominant source of sediment (VanLaningham et al., 2009). The Bering Sea is a significant source of sediment to the Meiji Drift during glacials, whereas during interglacials, the younger (2-15 Ma based on Nd isotopic data from VanLaningham et al., 2009) Kamchatkan and Aleutian arcs are the source of sediment to the Meiji Drift just south of the Bering Sea (VanLaningham et al., 2009).

1.2 Ice-rafted Debris

Paleoenvironmental proxies are important tools used in reconstructing Earth's climate history. Deep-sea sediment cores, which provide the longest and most continuous

record of paleoclimatic conditions and changes available on Earth, can go back as far as 150-170 My (Boyle, 2008). Ice-rafted debris (IRD) has widely been used as a direct indicator of glaciation. IRD are the detrital materials in marine sediments from ocean basins. Among these are the siliciclastics greater than 250 μm (von Huene et al., 1973; von Huene et al., 1976; Krissek, 1995; St. John and Krissek, 1999). Icebergs and sea ice are considered the likely transport mechanisms of sediment this size since these larger particles settle out relatively rapidly during fluid flow (Tripathi et al., 2008; Ruddiman, 1977). Deposited far from a continental shelf, these larger, coarser terrigenous grains are an anomaly among the finer grained pelagic or hemipelagic sediment matrix (Krissek, 1995). While pyroclastics and authigenics can be ice-rafted, there are other more likely transportation mechanisms and no clear way of identifying which portion if any has been ice-rafted (Figure 1). Therefore, to be conservative, only the siliciclastic component is considered IRD.

Figure 1. Transportation Mechanisms. This figure shows the different transport mechanisms of sediment to the deep sea (Modified from Lisitsyn, 2002).

The placement of IRD gives two important pieces of information. First, the composition and geographic locations of IRD can show the glaciated source areas (see Table 8). Second, analysis of the stratigraphic distribution of IRD can reveal when continental glaciers reached sea level (Krissek, 1995). Since other possible transportation mechanisms could have been responsible for grain deposition to the deep sea, such as nepheloid layers, turbidites, and local re-sedimentation, other considerations are needed to determine if the sediment has been ice-rafted. These considerations include surface microtextures of quartz grains, provenance, and degree of sorting. To better understand the temporal and spatial onset of the NHG, IRD was used in this study as a proxy for past continental glaciation.

1.3 Sea Ice vs Glacial Ice

Although ice sheets are the largest and most influential factor of the Earth's climate systems (Boulton, 2006), glaciers, ice caps, and ice shelves are forceful parts of the climate as well. Ice caps are smaller than ice sheets and tend to be dome-shaped ice and snow that completely covers a mountainous area, preventing peaks or Nunataks from being visible (Molina, 2004). In contrast glaciers are classified by their size, location, and thermal regime. As glaciers and ice sheets shear the rock bed below, sediment becomes incorporated into the basal ice through abrasion, plucking, and erosion by subglacial meltwater (Glazovskii, 1989; Boulton, 2006). This sediment is delivered to the seafloor by icebergs and can be a significant input of sediment. A range of fine (silts and clays) and coarse (boulders, pebbles) granulometric fractions distinguishes deposits from iceberg rafted sediment. Therefore the sediment is less sorted and tends to be more

rounded (Lisitzin, 2002). The sediment in sea ice comes from suspension freezing of particles in the water column or on the sediment bed, which is known as sediment entrainment. Sediment entrainment is caused by subfreezing temperatures, strong winds, and powerful turbulence in shallow, open water (Reimnitz et al., 1998). Anchor ice forms by freezing on the seafloor, capturing sediment, which later floats to the surface. The entrained sediment needs to be coarse enough to resist the drag and buoyancy of ice buildup (Arden and Wigle, 1972; Reimnitz et al., 1987). Consequently, the sediment entrained in anchor ice is typically sand and pebbles, and is less sorted than other types of sea ice, like frazil ice. When water crystallizes with suspended sediment at its center, frazil ice is formed (Lisitzin, 2002). Frazil ice commonly forms on the surface of the water where temperatures are coldest. This is commonly an unconsolidated mix of ice crystals and supercooled water. It commonly precedes the formation of pancake ice, which rafts together to form sea ice. Frazil ice preferentially entrains finer and better-sorted sediment (Reimnitz et al., 1987, Aiello and Ravelo, 2012). Melting also causes sorting, as the sediment dispersed throughout the ice is concentrated at the surface of the sea ice. As the ice begins to melt, fine sediment is carried into the sea by the meltwater, while the coarse grains remain (Reimnitz et al., 1998; Aiello and Ravelo, 2012). Overall, sediment entrained in sea ice through suspension freezing generally tends to be finer than iceberg-rafted sediment (Nürnberg et al., 1994). The sediment size of the siliciclastic fraction provides clues to determining the transport mechanism for sea ice (Table 1).

Table 1

Transportation by Ice

Type	Mode	Sediment Description	Source	Texture	References
Sea Ice	Frazil	Predominantly clay and silt, sorted	Water Column	Rounded, sub- rounded	Reimnitz et al., (1987); Lisitzin, 2002
	Anchor	Coarse, less well sorted	Seafloor	Rounded, sub- rounded	Reimnitz et al., (1987); Lisitzin, 2002
Glaciers	Fracturing, Abrasion, Erosion	Large range of size fractions, less well sorted	One or few point sources	Angular, rough, uneven with scores, grooves and striations	Lisitzin, 2002
Ice Sheets	Fracturing, Abrasion, Erosion	Large range of size fractions, less well sorted	Multiple point sources	Angular, rough, uneven with scores, grooves and striations	Lisitzin, 2002

1.4 Previous IRD Studies

The Atlantic has been an area of interest for paleoclimate research. The Spectral Mapping (SPECMAP) project in the 1980s created a timescale going back to the early Pleistocene based on oxygen isotopes from planktonic foraminifera (Imbrie et al., 1984; Gornitz, 2009). Other important studies in the North Atlantic and Greenland Sea were done on Ocean Drilling Program (ODP) Legs 108, 151, and 172. While there have been many studies looking at IRD in the Atlantic (Shackleton et al., 1984; Andrews, 1999; Caissie et al., 2010), there has been relatively little research in the North Pacific and even less in the Bering Sea. Among the first to obtain sediment cores from the Bering Sea were Creager and Scholl (1973) on Deep Sea Drilling Project (DSDP) Leg 19. Leg 19 recovered cores from the North Pacific near the Aleutian Islands in 1992 (Rea et al., 1993; März et al., 2013). Previous IRD studies in the North Pacific have found limited IRD prior to 2.6 Ma, but IRD significantly increasing at around 2.6 Ma and 1.0 Ma (Krissek, 1995, Table 2). This finding suggests that the onset of NHG (occurring on a continent scale) at 2.6 Ma, and the intensification of NHG climate fluctuations at around 1.0 Ma (Stewart et al., 1979; Fullam et al., 1973; Conolly and Ewing, 1970; Kent et al., 1971; Rea and Schrader, 1985; Krissek et al., 1989; Krissek, 1995). Due to outdated drilling technology and poor recovery, however, it was not until IODP Expedition 323 in 2009 that long intact sediment cores from the Bering Sea were recovered and with it a 5 my climatic and oceanographic record of the North Hemisphere (Cook et al., 2005; Okazaki et al., 2005; Takahashi et al., 2005; IODP Exp 323 Scientists, 2010).

Table 2

Past Research

Previous Work	Location	Key Findings
Kent et al., 1971	North Pacific	Found IRD at 2.6 Ma with increase of ice-rafting beginning at ~1.2 Ma
von Huene et al., 1973	Gulf of Alaska, DSDP Leg 18	Demonstrated that the abundance of the coarse-sand fraction (250 μm to 2 mm) is a valid indicator of IRD abundances
Rea and Schrader, 1985	North Pacific, Bering Sea, DSDP Legs 18, 19	Onset of IRD at 2.6 Ma
Krissek et al., 1985	NW Pacific, ODP Leg 145	Found IRD at 2.6 Ma with an increase beginning at 1.0 Ma
Krissek, 1996	NW & NE Pacific, ODP Leg 145	IRD occurrence consistent with Kent et al., 1971
McKelvey et al., 1995	NW & NE Pacific, ODP Leg 145	NW Pacific IRD derived from the western Bering Sea and the Kamchatka Peninsula, NE Pacific IRD derived from SE Alaska
Prueher and Rea, 1998	NW & NE Pacific, ODP Leg 145	Sudden and drastic input of IRD a 2.67 Ma in the North Pacific. Propose intensification could be linked to increased volcanism
St. John and Krissek, 1999	North Pacific, DSDP 580	Found regional spatial and temporal differences in IRD in the North Pacific
März et al., 2013	Bering Sea, IODP Exp 323	$\text{SiO}_{2\text{XS}}$ accumulation rates suggest NHG onset at 2.6 Ma

1.5 Evolution of the Cryosphere

The sedimentary record has revealed that Earth's climate over the last 65 million years has ranged from warm periods with ice-free poles to cold periods with expansive continental ice sheets (Zachos et al., 2001). These changes in climate are largely reflections in the changes of Earth's orbital parameters: eccentricity, obliquity, and precession, and plate tectonics (Zachos et al., 2001). Using delta oxygen-18 isotopes ($\delta^{18}\text{O}$) from benthic foraminifera collected from marine sediment cores globally distributed, several researchers have been able to establish a record of paleoclimate and polar ice volume, as can be seen in Figure 2 (Zachos et al., 2001; Lisiecki and Raymo, 2005).

Figure 2. Evolution of the Cryosphere. Ice sheets began forming in Antarctica in the late Eocene whereas glaciation began in the northern hemisphere around the Pliocene (Adapted from Zachos et al., 2001 and Lisiecki and Raymo, 2005).

These records show that ice sheets began to develop in Antarctica nearly 40 million years ago and near the end of the Pliocene in the North Hemisphere. The onset and development of Antarctic glaciation during the early Eocene is widely agreed upon (Zachos et al., 2001; DeConto and Pollard, 2003; Coxall et al., 2005); however, the beginning of glaciation in the North Hemisphere remains under debate (Tripathi et al., 2005). Eldrett et al (2007) point to problems in preservation of the sedimentary record through the early Cenozoic in cores from the North Hemisphere that could have been caused by stronger ocean circulation and/or a drop in sea level due to Antarctic ice sheets (Eldrett et al., 2007). Due to the poor records that have been recovered so far, they argue that the lack of ice-rafted debris (IRD) found in sediments that date back to before the mid-Miocene cannot rule out an earlier onset of glaciation in the North Hemisphere.

1.6 North Hemisphere Glaciation

The North Hemisphere experienced a significant change in climate at the end of the Pliocene (Figure 3). Ice sheets began to form, expand, and persist. Sea surface temperatures cooled, the thermohaline circulation weakened, the concentration of atmospheric carbon dioxide decreased, sea level lowered, and the ocean became more stratified (Ravelo et al., 2004).

Figure 3. Onset of NHG (LR04). The gradual cooling that began at the beginning of the Pliocene intensified with onset of North Hemisphere Glaciation. The late Pliocene was characterized by warmer SSTs (red) than the Pleistocene (blue). The area in outlined in the black box is the time period covered by this study (Adapted from Lisiecki and Raymo 2005).

The cause for these changes is still under debate. According to the Milankovitch theory, glaciation needs cool summers in the northern latitudes to prevent melting and ablation (Maslin et al, 1998). This, along with enough moisture, allows the snowpack to persist and expand which increases surface albedo and acts as a positive feedback (Driscoll and Haug, 1998). These favorable conditions for glaciation were present at the end of the Pliocene through lower summer insolation and possibly, an enhanced thermohaline circulation that would have transported more moisture northward (Raymo et al., 1992; Driscoll and Haug, 1998). These factors, in combination with certain tectonic events, could have led to the development of North Hemisphere Glaciation (NHG) (Maslin et al., 1998). Some of the theories include an increase in volcanism (Prueher and

Rea, 2001), the closure of the Central American seaway (Driscoll and Haug, 1998), the uplift of the Himalayas (Raymo et al, 1988), and the ending of a permanent El Nino (Molnar and Cane, 2002; Philander and Fedorov, 2003; Ravelo et al., 2004; Wara et al., 2005; Lunt et al., 2008).

1.7 Geology of Alaska

In determining the spatial scale of glaciation in the North Hemisphere, understanding the geology of Alaska and Kamchatka is critical. Alaska is divided into two key tectonic and geological regions: the Arctic coastal plains and the Cordillera. Within the Cordillera, there are three main subdivisions. These include a northern mountainous region, a central intermontane area, and a southern mountainous region (Plafker and Berg, 1994). The Cordillera includes the Yukon River Basin and the Anadyr River Basin. Both are dominated by sedimentary and metasedimentary geologies with some volcanic, plutonic, and metamorphic regions (Figure 5; VanLaningham et al, 2009). The geology of Kamchatka is similar, but with a few more regions dominated by volcanics than Alaska (VanLaningham et al, 2009).

Figure 4. Geology of Alaska. Alaska is divided into two key tectonic and geological regions: the Arctic coastal plains and the Cordillera. Paleozoic and Mesozoic sedimentary rocks are the most widespread (Adapted from U.S. Geological Survey digital data).

St. John and Krissek (1999) found that the Northeast Pacific supplied more IRD than the Northwest Pacific. They also were able to determine that the presence of chert and rhodochrosite indicated the source to be Kamchatka Peninsula. Their conclusions pinpoint Southeast Alaska as the greatest source of IRD in the North Pacific possibly due to atmospheric circulation and geography that promoted more extensive glaciers.

1.8 Principal Site: U1341

There are two major topographic features in the Bering Sea: Shirshov Ridge, which separates the Aleutian Basin into an eastern and western part, and Bowers Ridge, which extends north from the Aleutian Island Arc (Creager and Scholl, 1973; Takahashi, 2005). Other important features include Bering Slope and Umnak Plateau. Researchers

on the Integrated Oceanic Drilling Program (IODP) Expedition 323 were able to collect marine sediment cores from the Bering Sea at seven sites: three along the Bering Sea shelf (U1343, U1344, U1345), three along Bowers Ridge (U1340, U1341, U1342), and one on the Umnak Plateau (U1339). Previous work on past DSDP and Ocean Drilling Program (ODP) sites (Leg 145: Sites 881, 882, 883, 884) near the Bering Sea found a sharp increase in IRD at 2.6-2.75 Ma (Krissek, 1995; Maslin et al., 1996; Figure 5).

Site U1341 is located on the western side of Bowers Ridge, which is a submerged volcanic arc formed during the Tertiary age (Cooper et al, 1987; Märtz et al., 2013). It provides a 4.3 million year record of the climatic and oceanographic conditions of the Bering Sea. For this reason, and because of its distance from the shelf, it can be used to study IRD.

Figure 5. IODP Expedition 323. Drilling Sites on Bowers Ridge in the Bering Sea from IODP Expedition 323 in 2009 (Adapted from IODP Expedition 323 Scientists, 2010).

Three holes were drilled at this site, with Hole U1341B being the deepest at 604.5 meters below seafloor (mbsf) (Exp 323 Scientists, 2010). The sedimentation rates at this site are approximately linear at 12 cm/ky, which is relatively high compared to other Bering Sea sites (IODP Exp 323 Scientists). The environmental conditions during sediment deposition are reflected in the distribution of the sedimentary components. The variations in lithology indicate that large-scale changes or events have occurred in the Bering Sea and are recorded in the sediment cores (Exp 323 Scientists, 2010). These events include onset of NHG. The lithology of the sediments are predominantly biogenic and siliciclastic with minor amounts of volcanic material (Figure 6).

Figure 6. Stratigraphic Column. The sediment at U1341B is dominated by biogenics especially diatoms (Adapted from Aiello and Ravelo, 2012).

The biogenics are mainly composed of diatoms with some radiolarians, foraminifera, and nannofossils (Exp 323 Scientists, 2010). U1341B is divided into two lithologic units. The younger unit (Unit I) is composed of alternating subunits of diatom silt, diatom clay, and diatom ooze. At 1.6 Ma, this changes to Unit II, which is characterized by its predominance of diatom ooze (Exp 323 Scientists, 2010). This change in lithology coincides with changes in the intensity of the magnetic susceptibility record and the abundance of calcareous tests (Exp 323 Scientists, 2010). Laminated intervals occur throughout U1341B, which indicate dysoxia. Dropstones begin occurring at U1341B at roughly 2.6 Ma and increase in frequency at 1.8 Ma (Exp 323 Scientists, 2010). Isolated dropstones were found at U1340 as early as 2.8 Ma (Exp 323 Scientists, 2010). There is also a volcaniclastic layer several meters thick that occurs around 2.7 Ma in U1340A right above a gap in the sediment record.

1.9 Sediment Transport Mechanisms

Due to the open ocean setting, there are potentially fewer sediment sources capable of transferring sediment 250 μm to Bowers Ridge (Figure 1). These include nepheloid layers, turbidity currents, local re-sedimentation from Bowers Ridge, sea ice, and icebergs. Bottom nepheloid layers are areas of increased turbidity as a result of thermohaline currents and internal waves (McCave, 2013). They typically consist of particles less than 2 μm and occur near continental shelves (McCave 1986; McCave 2013). While they are an effective means of transporting finer grained sediment and larger aggregated particles, they lack the necessary energy to transport significant amount

of coarse particles over long distances. Turbidity currents are capable of moving a wide range of size particles down slopes. Site U1341 is on the slope of Bowers Ridge and is in an area that could possibly be affected by turbidity currents. The $> 250 \mu\text{m}$ siliciclastics found at U1341 could be the result of local re-sedimentation from Bowers Ridge. If this were the case, the lithology of the $> 250 \mu\text{m}$ siliciclastics recovered would be similar to Bowers Ridge, which is submerged volcanic arc. According to Cooper et al (1970) and Wanke et al (2012) Bowers Ridge had subsided below sea level by the middle Miocene. Since Bowers Ridge was possibly no longer subaerial, local re-sedimentation of sediment after the middle Miocene is not likely or of minor significance. Sea ice and icebergs are effective methods of transporting sediment in a range of sizes over great distances. Sea ice would have extended over the continental shelf and possibly parts of the deeper Aleutian basin. As glaciers and ice sheets shear the rock bed below, sediment becomes incorporated into the basal ice through abrasion, plucking, and erosion by subglacial meltwater (Glazovskii, 1989; Boulton, 2006). This sediment is delivered to the seafloor by icebergs and can be a significant input of sediment. The sediment transported via ice related processes would be predominantly terrigenous, and the quartz grains would display microtextural features indicative of ice transport.

1.10 Thesis Rationale and Key Questions

Sediment cores are an important source of paleoclimatic data. The sediment cores act as recording devices, providing insight into the occurrence of geologic and climatic events. The causes for the observed cooling trend that began in the Eocene, the timing,

the differences in evolution of the cryosphere between hemispheres and between oceans and whether the intensification of the NHG occurred at ~2.7 Ma, remain under debate.

Using sediment core samples from Site U1341 in the Bering Sea, the chief aim of this thesis is to examine the record between approximately 1.9 and 4.2 Ma with a sample every 5ky to every 10ky, using sedimentologic tools to quantify temporal variabilities at millennial scales the coarse ($> 250 \mu\text{m}$) fraction of the siliciclastic component. Using sedimentologic criteria based on our modern understanding of the mechanism that entrains and delivers IRD, the goal is also to ascertain whether and which portion of the siliciclastic input is due to ice-related processes. Since a number of previous research (Kent et al., 1971; Krissek, 1995; Haug et al., 2005) suggest that the onset of the NHG began ~2.7 Ma this research focused on this critical period. Over 200 samples from the time period of interest are used to study the onset of NHG. With this data, it will be possible to determine if the NHG occurred gradually or abruptly, and to develop an improved record of how the modern cryosphere evolved.

The key questions that are addressed are:

1. What is the overall trend of $> 250 \mu\text{m}$ siliciclastics through time for the Being Sea and how does that compare to other IRD records from the North Pacific? Is there a clear change in flux of IRD at ~2.7 Ma?
2. What is the mechanism of delivery responsible for the deposition of $> 250 \mu\text{m}$ siliciclastics to Bowers Ridge?
3. Do changes in the source of IRD occur during the study period? Is there a difference in IRD concentration and composition before and after 2.7 Ma?

Materials and Methods

A > 250 μm siliciclastics record from Bowers Ridge was determined using 250 samples with a temporal resolution that ranges from one sample every 7 ky to 28 ky (Table 3, see age model in Table 4, Figure 7).

Figure 7. Methodology Flowchart. This figure summarizes the key elements in the methodology used in this project.

Table 3

Age Between Samples

	Range (Age Ma)	Average (ky)
All	1.95-4.27	10.37
Interval A	1.95-2.49	28.26
Interval B	2.49-2.97	9.02
Interval C	2.97-3.19	9.80
Interval D	3.19-3.53	7.52
Interval E	3.53-4.27	8.68

2.1 Core Collection/Age Model

The marine sediment cores used in this study were obtained from the United States Implementation Organization drill ship, JOIDES Resolution. The coring systems used on board were the Advanced Piston Coring, which can drill to a depth of 438 m below seafloor (DSF) and the Extended Coring Barrel that can drill to 600 m DSF (März et al., 2013; Exp 323 Scientists, 2010). A composite depth scale was made from a splice of Holes 1341A, B, and C according to their stratigraphic position (März et al., 2013; Table 4). A preliminary age model was developed shipboard and updated onshore using diatom biostratigraphy, magnetic properties, and assuming linear sedimentation rates (Expedition 323 Scientists, 2010; Takahashi et al., 2011; Teraishi et al., 2012; März et al., 2013).

Table 4

Age Model

Depth CSF-A (m)	Depth CCSF-A (m)	Control point	Sedimentation rate (m/myrs)	Age (Ma)
0	0		113.962	0
32.9	34.2	Depth of LO <i>Proboscia curvirostris</i>	111.384	0.3
82.5	87.8	Brunehes bottom	202.085	0.781
137.8	146.6	Jaramillo bottom	162.015	1.072
245.5	261.0	Depth of RI <i>N. seminae</i>	157.898	1.778
367.8	387.7	Gauss top	137.900	2.581
535.9	555.8	Depth of FO <i>N. koizumii</i>	137.900	3.8

Note. Updated from Expedition 323 Scientists.

2.2 Sieving and Weighing

The first step in developing a mass accumulation rate is isolating the $> 250 \mu\text{m}$ siliciclastics and calculating its weight relative to the beginning sample weight. Roughly

10 g of each sample was measured out and placed in the oven for approximately 20 minutes or until dry. The samples were then reweighed to determine the dry weight. Next, the samples were submerged in 5 mL of 10% hydrogen peroxide and warm deionized water (40°C). The samples then sat for 5 minutes. The hydrogen peroxide and water were carefully removed using a 0.10 mL pipette. Fresh deionized water and a few drops of sodium hexametaphosphate were added, and the samples were disaggregated ultrasonically for 5 minutes. Then the samples were wet sieved at 63 μm and 250 μm , dried using heat lamps, and weighed again after cooling to room temperature. For a handful of samples, disaggregation was not completely effective and ultrasonication had to be repeated for an additional 5 minutes.

Since only the siliciclastic component of the $> 250 \mu\text{m}$ fraction is considered IRD, it is important to remove all other components. So, once the $> 250 \mu\text{m}$ size fraction was isolated, the biogenics, authigenics, and pyroclastics were removed under a dissecting microscope using 400x magnification and the samples were weighed again (Figures 8 and 9).

The weight was then normalized to the starting dry weight and is a measure of medium and coarse sand abundance. The resulting ratio of the dry weight of the medium to coarse siliciclastic fraction to the dry sample weight is the $> 250 \mu\text{m}$ siliciclastic weight percent (referred to as IRD Wt% in the below equation).

Figure 8. Authigenic Sediment. Example of authigenic sediment found in several U1341B samples (Photograph by author).

Figure 9. Common Biogenics. Forams (left), sponge spicules and radiolarians (right) were among the most common types of biogenics found in U1341B samples (Photograph by author).

The 63-250 μm fraction was recovered, dried, and weighed. However, due to the relatively high abundance of grains in this fraction, the siliciclastic portion was not isolated or quantified beyond a weight percent.

2.3 Mass Accumulation Rates (MAR)

Mass accumulation rates (MAR) are used to quantify the flux of a sedimentary component taking in consideration both the sedimentation rate and bulk density of the sediment. Both weight percent and MARs are calculated for each sample.

For the $> 250 \mu\text{m}$ siliciclastics the MAR is calculated it using the equation below:

$$\text{IRD MAR} = \text{IRD Wt\%} \times \text{DBD} \times \text{LSR}$$

Where IRD wt% is the normalized weight percent of the $> 250 \mu\text{m}$ siliciclastic size fraction, DBD is the dry bulk density (g/cm^3), and LSR is the linear sedimentation rate (cm/ky) (Krissek, 1995). The dry bulk density data and linear sedimentation rate were obtained from Exp 323 Shipboard data. DBD ranged from 1-2 g/cm^3 and LSR ranged from 13.79-15.23 cm/ky .

2.4 Provenance

The purpose of a provenance analysis is to determine the source rock of a siliciclastic sediment and based on a knowledge of the geology of the potential source areas, to reconstruct the location of the rock from which the sediment originated from. This investigation involves examining the composition and texture of the sediment (Pettijohn et al., 1987; Weltje and Eynatten, 2004). Dickinson and Suczek (1973) established a relationship between sandstone composition and plate tectonics. The three general groups that Dickinson and Suczek (1973) use in their classification are: continental block, magmatic arc, and recycled orogeny (Figure 10).

Figure 10. Ternary Diagram. The ternary diagram relates sediment composition to provenance type (Adapted from Dickinson and Suczek, 1973).

By using ternary diagrams showing the relative framework ratios of monocrystalline quartz, feldspars (potassium feldspar as well as plagioclase), and lithics it is possible to discern the primary provenance type based on Dickinson's work with sandstones. Only framework grains (i.e. > 5% of the sediment) are considered in the ratio since the interstitial cement or matrix are products of diagenesis (Dickinson and Suczek 1973; Dickinson et al., 1983)

The composition of the > 250 µm siliciclastics was determined by identifying each grain and the interpretation was done using the three components used by Dickinson's provenance study: quartz, feldspar (K-feldspar and plagioclase), and lithics. This was done visually using a binocular dissecting microscope. A sample was placed into a plastic tray lined with graph paper to make counts easier. Lithics included all minerals other than quartz and feldspar, volcanics, rock fragments, metamorphics, and

polycrystalline quartz. Quartz was identified by its lack of cleavage and its transparency. The criteria for feldspar included opaqueness and a 90° cleavage plane. To aid in the identification Simon and Schuster's Guide to Rocks and Minerals was referenced as well Ward's Geology Collection. Some commonly occurring grains were examined with SEM and EDX to more confidently identify them.

2.5 Quartz Surface Texture Analysis

By examining surface microtextures of quartz grains using a scanning electron microscope (SEM) it is possible to differentiate between various transportation mechanisms (Krinsley and Donahue, 1968; Ehrlich and Weinberg, 1970; Culver et al., 1983; Helland and Holmes, 1997; Dunhill, 1998; Bull and Morgan, 2006; Molén, 2014). Of interest to this study are those transportation mechanisms related to ice. These include sea ice rafted debris (SIRD) and glacial ice rafted debris (IRD).

To analyze the microtexture features on the surface of quartz grains, 30 transparent quartz grains distributed throughout the research time period of 1.9-4.2 Ma were chosen for analysis with the scanning electron microscope (SEM) and energy dispersive x-ray spectroscopy (EDX). Stubs were prepared with the selected quartz grains. The stubs were then coated with gold to a thickness of 9 nm. All the grains were analyzed with EDX to verify that they were quartz (SiO₂). They were then examined with the SEM. The analysis of quartz surface microtextures is based largely on Molén's (2014) method. This involves identifying the major microtextures and determining the relative history of each. The diagnostic microtextures indicating a glacial environment include mechanical (abrasions, fractures), chemical (weathering, solution, precipitation), and

crystallization (nodes, embayments, intergrowths, overgrowths, new growth) (Molén, 2014).

Three geological histories are considered: History-0 (fresh), History-1 (recent), and History-2 (old). The abundance of each feature along with its relative age is determined visually.

Results

A total of 226 samples covering the time period between 1.9 and 4.27 Ma from U1341B were processed to establish a $> 250 \mu\text{m}$ MAR record and to create a ternary plot to determine provenance. Surface texture analysis was performed on 30 quartz grains. The results are presented below.

3.1 Mass Accumulation Rate

The mass accumulation rates plotted as a function of time show distinct intervals. Each interval was established by considering average MAR (Figure 11, Table 5) and composition. Some intervals are distinguishable by their low average MAR and weight percent, while others have numerous peaks. There is another clear division around 2.75 Ma (Interval B). The abundance and frequency of peaks substantially increases beginning at 2.7 Ma. The abundance and frequency of peaks substantially increases beginning at 2.7 Ma.

Figure 11. 250 μm siliciclastic MAR. This plot shows the 5 intervals based on average MAR values and composition: Interval A (red), Interval B (blue), Interval C (orange), Interval D (yellow), and Interval E (green).

The MAR for the medium to coarse siliciclastic fraction range from 0 to 40.9 $\text{g}/\text{cm}^2/\text{ky}$ (0 - 0.0116 wt%). The greatest MAR value occurs at 2.66 Ma (~ 378.5 mbsf). The following descriptions are from the oldest to the youngest part of the record (Table 5).

Table 5

MAR Summary (g/cm²/ky)

	Interval A	Interval B	Interval C	Interval D	Interval E
Ages (Ma)	1.95-2.49	2.49-2.97	2.97-3.19	3.19-3.53	3.53-4.27
Avg > 250 μm MAR	0.011	0.013	0.003	0.008	0.001
Std Dev	0.014	0.019	0.003	0.015	0.003
Max > 250 μm MAR	0.051	0.409	0.011	0.085	0.017
Min > 250 μm MAR	0.001	0.000	0.000	0.000	0.000
# of Samples	20	102	27	45	79

Interval E: There is essentially no or very low occurrence of > 250 μ m siliciclastics; the oldest part of the record, from 3.6 to 4.27 Ma, and the ~3.22 to 2.67 Ma interval similarly have very little to none. The maximum MAR value for this interval is 0.085 g/cm²/ky.

Interval D: Following this period of low MAR values, there is a sudden increase at 2.66 Ma that continues until the end of the record at 1.94 Ma

Interval C: This interval is characterized by a low average MAR of 0.003 g/cm²/ky. The maximum MAR value for this interval is only 0.011 g/cm²/ky.

Interval B: 2.75 Ma falls within this interval. The abrupt increase in MAR occurs at ~2.66 Ma where the maximum MAR value of 0.409 g/cm²/ky occurs.

Interval A: Has a moderate average MAR of 0.011 g/cm²/ky.

Interval A: 1.95 to 2.49 Ma

Interval A is composed of samples ranging in age from 1.95 to 2.49 Ma. This interval contains the youngest samples.

Figure 12. Interval A MAR. This figure shows Interval A, which ranges in age from 1.95 to 2.49 Ma. The average 250 μm siliciclastic MAR values are relatively high.

The average weight percent of > 250 μm siliciclastics is 0.054. This is second highest of the intervals. The average MAR value is 0.011 g/cm²/ky. There are three distinct increases in abundance during this time period shown in Figure 10. Each increase in abundance includes a distinct increase in MAR that is 2 to 3 times the average within the interval. The first occurs from 1.97 to 2.14 Ma, the second at 2.14 to 2.29 Ma, and the last at 2.29 to 2.48 Ma. The increase lasts 17.03 ky, 15.56 ky, and 19.13 ky respectively.

Interval B: 2.49 to 2.97 Ma

Interval B is composed of samples ranging in age from 2.49 to 2.97 Ma shown in Figure 13. This interval has one large spike at 2.66 Ma with a MAR at 0.409, which is significantly larger than other values within the interval and in any other interval.

Figure 13. Interval B MAR. Interval B that covers the period from 2.49 to 2.97 Ma. It is characterized by high average 250 μm siliciclastic MAR values.

The average weight percent of $> 250 \mu\text{m}$ siliciclastics is 0.064 and the average MAR value is $0.013 \text{ g/cm}^2/\text{ky}$. This is the largest average weight percent and MAR among the intervals and was the only one to contain a pebble (weight not shown in weight percent and MAR plots) shown in Figure 14. There are eight increases in abundance within this interval.

Figure 14. Pebble from Interval B. This is a picture of the only pebble found in U1341B sediment in the study period (Photograph by author).

Interval C: 2.97 to 3.19 Ma

Interval C is composed of samples ranging in age from 2.97 to 3.19 Ma shown in Figure 15.

Figure 15. Interval C MAR. Interval C, which ranges from 2.97 to 3.19 Ma. This period has low average 250 μm siliciclastic MAR values.

The average weight percent of $> 250 \mu\text{m}$ siliciclastics is 0.013. This is the second lowest average weight percent among the intervals. This interval has a consistently fairly low level of MAR. The average MAR value is $0.003 \text{ g/cm}^2/\text{ky}$. It occurs between two intervals with relatively high average weight percent and MARs.

Interval D: 3.19 to 3.53 Ma

Interval D is composed of samples ranging in age from 3.19 to 3.53 Ma shown in Figure 16.

Figure 16. Interval D MAR. Interval D covers the period from 3.19 to 3.53 Ma. This interval is the first to show a moderate abundance of 250 μm siliciclastic MAR values.

The average weight percent of $> 250 \mu\text{m}$ siliciclastics is 0.045. This is the first interval with a major presence of $> 250 \mu\text{m}$ siliciclastics. The average MAR value is $0.008 \text{ g/cm}^2/\text{ky}$.

Interval E: 3.53 to 4.27 Ma

Interval E is composed of samples ranging in age from 3.53 to 4.27 Ma shown in Figure 17. This interval has the oldest samples. This interval is characterized by having the lowest average weight percent of $> 250 \mu\text{m}$ siliciclastics at 0.011 and the lowest average MAR values at $0.001 \text{ g/cm}^2/\text{ky}$.

Figure 17. Interval E MAR. Interval E ranges from 3.5-4.3 Ma. This interval is characterized by low average 250 μ m siliciclastic MAR values.

3.2 Provenance/QFL Diagram

Following Dickinson and Suczec's (1979) use of a ternary diagram to plot the three dominant compositional groups: monocrystalline quartz, feldspar, and lithics in the samples to determine provenance, a QFL diagram was plotted with the grain counts (Figure 18).

Figure 18. QFL Ternary Diagram. The older samples in Intervals D and E are concentrated near the lithics, whereas the younger Intervals A and B are more centered.

By highlighting each of the intervals within the QFL diagram, a trend from the oldest sediment to the youngest can be detected. The oldest samples (intervals D and E) dominant the lower left of the ternary plot indicating a high percent of lithics and relatively low quartz and feldspar. The younger samples (intervals A and B) still have a high degree of lithics with respect to quartz and lithics, but less than that of the older intervals (D and E). The source of intervals D and E is from an undissected arc, whereas intervals A and B are likely from a transitional arc.

3.3 Quartz Surface Texture Analysis

Surface microtexture analysis was performed on 30 quartz grains from site 1341. Two groups of samples (15 grains in each group) were examined prior to and after 2.75 Ma (Figure 19).

Figure 19. Common Surface Texture Features. Pictured above are some of the most common features of the quartz grains. The samples younger than 2.75 Ma (left) had many fresh fractures, whereas the older samples (right) had a more diverse surface texture (Photographs by author).

Molén's method for defining the surface features of interest was followed. From his simple approach, the quartz grains from each of the two groups were examined for large and small fractures, indications of weathering, crystal growth, abrasions, and embayments/nodes (Table 6).

Table 6

Molén's Method

Symbol	Feature and Extent
F	Large fracture that covers ≥ 20 -25%
f	Small fracture that covers ~ 5 -20%
A	Abrasion that covers ≥ 15 -20%
SP	Weathering that covers ≥ 10 -15%
EN	Embayment/nodes that cover ≥ 10 -15%
C	Crystal growth that covers ≥ 5 -10%

Discussion

One of the most important aspects of this thesis is the use of ice rafted debris as a direct indicator of glaciation. The first step is to determine what is, and what is not, ice rafted debris. Size is one of the most critical factors in this determination since grains much larger than $> 250 \mu\text{m}$ settle out relatively close to the continental shelf and are therefore transported to Bowers Ridge by ice. Size, however, cannot be the only criterion. Composition, provenance, and surface features are additional important criteria to be considered. Previous work done in the North Pacific and Bering Sea has detected a sharp increase in IRD around 2.6-2.75 Ma (Krissek, 1999; März, 2013).

The plot of $> 250 \mu\text{m}$ siliciclastics MAR vs. age reveals the earliest occurrence of significant $> 250 \mu\text{m}$ siliciclastics beginning at 3.6 Ma. Abundance increases until it peaks at 3.3 Ma, and then begins to decline. For 600 ky between 3.2 and 2.6 Ma, the presence of $> 250 \mu\text{m}$ siliciclastics is minimal until it abruptly increases and remains high until the end of the record at 1.9 Ma. There are five intervals based on $> 250 \mu\text{m}$ siliciclastic abundance and age that can be discerned from this plot. Another distinguishable division is at 2.6 Ma. From 1.9 to 2.6 Ma the MAR, of $> 250 \mu\text{m}$ siliciclastics is relatively and consistently high. In contrast, after 2.6 Ma, the $> 250 \mu\text{m}$ siliciclastic abundances vary from low to high. Based on the widespread detection from earlier research in the region that onset of continental glaciation began around 2.75 Ma, it is reasonable to assume that the $> 250 \mu\text{m}$ siliciclastics detected between 1.9 and 2.75 Ma in the U1341 record are indicative of ice-rafting. Therefore the discussion on the record

will be separated for the interval younger than 2.75 Ma and the interval older than 2.75 Ma (Figure 20).

Figure 20. MAR vs Time Plot. This MAR plot is divided into two groups based on age (before and after 2.75 Ma).

4.1 Younger than 2.75 Ma

Previous work from the North Pacific and Bering Sea has detected a sharp increase in IRD around 2.6-2.75 Ma (Haug et al., 1995; Krissek, 1995; März, 2013). It is, therefore, very likely that the $> 250 \mu\text{m}$ siliciclastics younger than 2.75 Ma from 1341B are related to ice processes. Additionally, larger dropstones were recovered from U1341 beginning at approximately 2.8 Ma. This correlation between the appearance of dropstones and the increase in coarse sand provides further support that the sediment younger than 2.75 Ma was ice-rafted.

The average MAR value for the $> 250 \mu\text{m}$ siliciclastic portion younger than 2.75 Ma is $0.015 \text{ g/cm}^2/\text{ky}$. Between 2.66 and 2.75 Ma, the MAR values are moderate with a

few slight increases in abundances. At 2.65 Ma, there is a dramatic sudden increase with a value of $0.409 \text{ g/cm}^2/\text{ky}$. After this sudden increase, there are several more increases in abundances and MAR values remain fairly high until the end of the record at 1.9 Ma. The fluctuation in this part of the record includes intermediate data points that lessen the possibility that these fluctuations are merely reflections of the sampling interval (Krissek, 1995).

The MAR record from U1341 was compared to Krissek's (1995) MAR plot from Sites 881, 883, and 887 on ODP Leg 145 (Figure 21). Overall Krissek's MAR values are higher than those for U1341. This is likely the result of the location of the sites. Sites 881 and 883 are both in the Northwestern Pacific near Kamchatka Peninsula.

Figure 21. Kriisek vs U1341. This figure compares the MAR values for U1341 with Kriisek's (1995) data from sites 881, 883, and 887.

The surface circulation in the Bering Sea is a cyclonic gyre. The Bering Slope Current (BCS), which flows along the continental shelf, forms the eastern boundary of the gyre, and the southward flowing Kamchatka Current (KC) forms the western boundary (Cook et al., 2005). The Alaskan Stream flows through passes in the Aleutian Islands and joins the gyre circulation (Cook et al., 2005). Iceberg drift closely follows circulation patterns (Robe, 2012). Icebergs near the continental shelf would flow northward across the shelf carried by the BCS. As they approach Kamchatka Peninsula, the KC would cause some icebergs to exit the Bering Sea through Kamchatka Strait. Those that were not carried out of the Bering Sea would flow along the Aleutian Islands by the Aleutian North Slope Current (ANC) and past Bowers Ridge (Caissie et al., 2010).

Site 887 is located near the Patton-Murray Seamounts in the Northeastern Pacific.

Icebergs that did not melt after exiting through Kamchakta Strait would travel eastward by the Kuroshio Current and the Sub-Arctic Front before passing over the Patton-Murray Seamounts and moving westward with the Alaskan Stream. Based on these circulation patterns, it seems reasonable that sites 881, 883, and 887 would have more icebergs drift past those locations than U1341. The dominant sources of IRD in the region are the Kurile-Kamchatka margin and the coast along the Gulf of Alaska, so that sites close to those areas would receive the most IRD input compared to more central and southern locations (Krissek, 1995). St. John and Krissek (1999) found that during the Pleistocene the northeast Pacific received a greater supply of IRD than the northwest Pacific. The average MAR values for each site between 1.9 and 2.75 Ma is summarized in Table 7. The U1341 record is similar in many ways to Krissek's.

Table 7

Average MAR Values

Site	U1341	881	883	887
Avg MAR	0.015	0.052	0.001	0.04

Previous work in the North Pacific and Bering Sea has found two primary sources for ice-rafted debris: Kamchatka Peninsula and Alaska. The findings of the major provenance studies are summarized in Table 4. McKelvey et al. (1995) and St. John and Krissek (1999), using data collected from sites 881, 883, and 887 from ODP Leg 145, have determined that IRD from 881 and 883 are most likely from Kamchatka Peninsula, whereas the source of IRD from 887 is probably southeastern Alaska. The St. Elias Mountain's proximity to the coast is a major contributor of sediment to the Gulf of

Alaska (Reece et al., 2011). Following Dickinson and Suczec (1973)'s use of a ternary diagram to plot the ratio of three dominant compositional groups: monocrystalline quartz, feldspar (K-feldspar and plagioclase), and lithics to determine provenance, a clear provenance and age relationship emerges.

The samples younger than 2.75 Ma are primarily from a transitional arc provenance. The lithics in these younger samples are mostly granites with some schists and volcanic rocks. There was possibly chert in these samples, which would indicate that Kamchatka Peninsula is the source. This would fit with the surface circulation patterns and with what St. John and Krissek (1999) found.

4.2 Older than 2.75 Ma

An increase in the abundance of $> 250 \mu\text{m}$ siliciclastics MAR vs age plot begin to occur at 3.6 Ma.

The samples older than 2.75 Ma include part of Interval B, and all of Intervals C, D, and E. Intervals C and E are distinguishable by their lack of $> 250 \mu\text{m}$ siliciclastics. These groups have the lowest average MAR values. Between them is Interval D occurring between 3.27 and 3.55 Ma, which has five increases in abundance in the $> 250 \mu\text{m}$ MAR record and the second overall highest peak.

The most notable difference between the $> 250 \mu\text{m}$ MAR records for U1341 and Krissek's is the earlier presence of several $> 250 \mu\text{m}$ siliciclastics peaks in the U1341 record. Between 3.27 and 3.55 Ma, a series of peaks in the $> 250 \mu\text{m}$ siliciclastics MAR record occur. These are the oldest presence of $> 250 \mu\text{m}$ siliciclastics that appear in the U1341 record. Unlike the younger samples, there are no corresponding peaks between the

two records. This lack of older peaks in Krissek's record and in other records from the North Pacific (Krissek, 1995; Haug et al., 1999), raises the question as whether this coarse siliciclastic fraction is indicative of early ice-rafting activity or is related to mass-gravity sedimentation from relatively local sources. Although the Lisiecki and Raymo (2005) stack shows the first significant increase in $\delta^{18}\text{O}\%$ occurring around 3.3 Ma, Krissek (1995) did find earlier detections of coarse sand in his record. At site 881, it occurred at 6.6, 5.5, 4.8-4.9, 4.5-4.6, 4.2, 3.9, 3.8, and 3.4-3.7 Ma. Krissek (1995) considered this older coarse sand insignificant since each occurrence was based on a single grain. In contrast, the older samples in the U1341 record have an average number of 21 grains per sample. The range is 0 to 488 grains. At site 883, Krissek found a single dropstone, which was described as pumice, and therefore not conclusively IRD. At site 887, coarse sand was found at 5.3-5.4, 5.2, 4.9, 4.3-4.4, 3.7-3.9, 3.5-3.6, 3.3-3.4, and 3.1 Ma. Krissek (1995) attributes the coarse sand older than 4.2 Ma to possible tidewater glaciers as suggested by the coeval siliciclastic deposits found in the Yakataga Formation in southeastern Alaska. Krissek does not consider significant glaciation to begin until 2.6 Ma, which is consistent with other work done in the region. Additional considerations involving the composition, provenance, and surface textures will be important in determining whether or not the sediment has been ice-rafted.

A critical first step in determining the mechanism of transport for the older intervals is pinpointing the source of the $> 250\ \mu\text{m}$ siliciclastics and comparing that to the younger intervals (Figure 22). The older samples (Intervals D and E) dominate the lithics

corner, indicating an undissected arc provenance whereas the younger samples (Intervals A and B) are more centered and are likely from a transitional arc provenance.

Figure 22. QFL Ternary Diagram. This plot shows the relative ratios of quartz, feldspar, and lithics.

The QFL ternary diagram suggests the provenance for Intervals D and E is an undissected arc. Dickinson and Suczek (1979) define this as an active magmatic arc from either an island arc or continental margin. While nearly all the samples in Interval E and most of Interval D fall within the undissected arc provenance, there is considerable overlap between Interval D and the two younger intervals, A and B. All the intervals are near the lithics corner of the ternary plot. To better understand the compositional differences between the younger and older groups, the lithics were subdivided into three groups: granite, metamorphic, and volcanic, recounted, and organized on a ternary plot

(Figure 23). There is a clear shift in lithology from volcanics to granites with time. This reinforces the likelihood that there are at least two separate sources for these samples.

Figure 23. MGV Ternary Diagram. This ternary plot of lithics is divided into Granite, Metamorphic and Volcanic.

Subdividing the lithics highlights the main compositional differences between the younger and older groups. The input of metamorphic clasts seems to remain constant over time. Starting with Interval E, which is clustered in the volcanics corner of the ternary plot, there is a shift over time to the granites corner. This suggests a change in provenance over time. It is clear from the ternary plot that Interval A's > 250 μm siliciclastics has a different source than Interval E.

Volcanics dominate Interval E. Interval D has a more varied composition, but with volcanics making up the majority of grains. There are two potential sources of

volcanics in the region: Bowers Ridge and the Aleutian Island Arc. The nearness of both sources suggests alternate mechanisms unrelated to ice processes could be responsible for the deposition of $> 250 \mu\text{m}$ siliciclastics at U1341. One possible mechanism is mass gravity flows, such as turbidity currents. Site U1340, which is adjacent to U1341, has large gaps and coarse deposits that occur at roughly the same time as the earlier abundance of $> 250 \mu\text{m}$ at U1341. The seismic profile of U1340 also shows that Bowers Ridge was tectonically active at this time, which could cause turbidity currents. For Bowers Ridge to be the source of the earlier abundance of $> 250 \mu\text{m}$ siliciclastics it would need to be subaerially exposed. However, Wanke et al. (2012) found evidence that Bowers Ridge has likely been inactive since the Miocene. Either Bowers Ridge was exposed to subaerial weathering longer than estimated or the $> 250 \mu\text{m}$ siliciclastics possibly came from the weathering of rocks exposed in the Aleutian Island Arc. The occurrence of shallow water sponge spicules (Aiello, personal communication) found in the sediment with the $> 250 \mu\text{m}$ siliciclastics suggests that Bowers Ridge had not been submerged for as long as previously estimated. There are other possible volcanic sources in Alaska so other factors need to be taken into consideration to determine the origin of the earlier abundance of $> 250 \mu\text{m}$ siliciclastics. One consideration is the presence of a terrigenous component in the $> 250 \mu\text{m}$ siliciclastics. VanLaningham et al. (2009) found that during glacial periods, Siberia and the region of Alaska drained by the Yukon River were the dominant sources of sediment to the North Pacific. The geology of the Yukon River basin is varied and that variety would be represented in any ice-rafted debris from that region. Quartz in pelagic sediments from the Pacific are typically derived from

terrigenous or volcanic sources (Rex and Goldberg, 1957). The quartz found among predominately volcanic sediment is usually without inclusions or as quartz phenocrysts, and found in minor amounts (Dickinson and Suzcek, 1979). Looking at just the flux of quartz is a useful way of determining the extent of the contribution of basement uplift sources. Quartz flux was calculated by the following equation:

$$\text{Quartz flux} = \# \text{ quartz/wt} \times \text{DBD} \times \text{LSR}$$

This is similar to the MAR calculation where DBD is dry bulk density and LSR is the linear sedimentation rate (Figure 24).

Figure 24. Quartz Flux. The upper plot shows quartz flux over time. A significant increase in quartz flux occurs right at 2.75 Ma. The lower plot is the MAR from U1341B.

From the plot of quartz flux, there is a considerable amount of quartz in the older samples. The average quartz flux for the samples younger than 2.75 Ma is 31.6 g/cm²/ky while the average value for the period between 3.2-3.55 Ma is 11.3 g/cm²/ky. The first major peak in the quartz record occurs at 2.75 Ma, in agreement with the timing of NHG found in other records (Haug et al., 1995; Krissek, 1995). The amount of quartz in the older samples is not present in comparable amounts to the younger samples, and therefore, it remains unclear how the > 250 µm siliciclastics were transported to Bowers Ridge. One last consideration involves examining the surface features of some of the grains for clues.

Quartz is a stable mineral that resists weathering and chemical alteration while also being abundant in potential sediment (Bull and Morgan, 2006). This makes it ideal for studying its surface for clues to how it was transported (Kransley and Donahue, 1968; Dunhill, 1998; Molén, 2014). Grains transported by glacial ice are distinguishable by surfaces having largely mechanical features such as conchoidal fractures, breakage blocks, and high relief (Dunhill, 1998). Molén (2014) has recently proposed a simpler, and more quantitative, method for evaluating quartz surface microtextures. He reduces the number of features to six and assigns each feature a relative history to indicate how fresh or old the features are compared to the grain's other features. This was done on two groups of 15 quartz grains: younger than 2.75 Ma and older than 2.75 Ma. Both groups of samples displayed the same degree of fracturing (Figure 25). The primary difference between the older and younger samples is the degree of weathering.

Figure 25. Surface Texture Analysis. This plot summarizes the surface features of quartz grains following Molen (2014)'s simple quantitative approach.

They were dominated by large fractures with some small fractures and abrasion.

The primary difference between the two groups was the extent of weathering. The younger samples were fresh whereas the older samples had considerably more solution/precipitation features indicating an older history (Figure 26).

Figure 26. Surface Texture Features Younger than 2.75 Ma. The dominant surface textural features are fresh fractures (Photographs by author).

It is clear that the younger group were transported by ice. It also seems likely that the older group or a portion of them was also transported by ice (Figure 27). There are more solution/precipitation features in these older samples than in the younger quartz grains.

Figure 27. Surface Texture Features Older than 2.75 Ma. The fractures are less fresh than the younger quartz (Photographs by author).

Is it possible the $> 250 \mu\text{m}$ in the samples older than 2.75 Ma are from the Aleutians or the result of local re-sedimentation from Bowers Ridge?

4.3 Hypotheses

There a number of possible explanations to account for the presence of $> 250 \mu\text{m}$ siliciclastics before 2.75 Ma.

1. The petrographic analyses indicate that the oldest samples from Intervals D and E are dominated by volcanic clasts. The QFL ternary plot indicates an undissected arc provenance. There are two potential sources of volcanic rocks to this region of the Bering Sea: the Aleutian Island arc and the Bering Sea basalt province (März et al., 2013). However, the closest potential source is Bowers Ridge is a

submerged magmatic arc that extends north from the Aleutian Island arc. This suggests that the $> 250 \mu\text{m}$ siliciclastics could be the result of local re-sedimentation of Bowers Ridge or the result of local tectonic activity.

A recent paper by Brigham-Grette et al. (2013) supports this hypothesis. The authors present a paleoclimatic record from the Siberian Arctic that challenges the notion that the conditions prior to 2.75 Ma were conducive for glaciation. The evidence that they present suggest that climate from 3.6- 2.2 Ma was as warm as during the Holocene, and therefore, too warm for glaciation to persist.

Horikawa et al. (2015) using neodymium (Nd) and lead (Pb) isotope data from detrital material from Site U1341 found that prior to Marine Isotope Stage (MIS) M2 about 3.32 Ma the terrigenous component of the sediment was likely from local (Aleutian Island) sources. This would support the interpretation that the $> 250 \mu\text{m}$ siliciclastics described in this study is similarly from local volcanic sources. However, Horikawa et al. (2015) did find pulses of terrigenous sediment with Alaskan provenance during this time interval. Two of these pulses occur at 3.319 Ma and 3.55 Ma. It is interesting to observe that the pulses have similar ages as the increase in $> 250 \mu\text{m}$ siliciclastics abundances at 3.33 Ma and 3.55 Ma found in this study (Figure 28). The more negative values of detrital ϵ_{Nd} indicate an Alaskan provenance, whereas the higher values suggest a local Aleutian source. Two of the Alaskan pulses in the detrital ϵ_{Nd} record occurring prior to 2.75 Ma align with two of the increases in $> 250 \mu\text{m}$ MAR abundances (yellow).

Figure 28. Detrital ϵ_{Nd} values. Detrital ϵ_{Nd} values from U1341B (Horikawa et al., 2015). (a) shows Horikawa et al.'s (2015) detrital ϵ_{Nd} values from U1341B (b) compared to the $> 250 \mu\text{m}$ MAR values from the same site.

A potential concern with this hypothesis is that for re-sedimentation to occur, Bowers Ridge would still need to have been sub-aerially exposed. The timing of the $> 250 \mu\text{m}$ siliciclastics found at U1341 occurs after Bowers Ridge was submerged according to Wanke et al. (2012) and Cooper et al. (2005). So if this sediment is from Bowers Ridge, then the previous estimate for its subsidence is off. The authors found that movement of Bowers Ridge had slowed and its volcanic activity had stopped by the end of the Cenozoic. However, the seismic profiles for U1340, a site on Bowers Ridge close to U1341, shows that the ridge was possibly still tectonically active. This could result in slumping and movement of coarse sediments including the $> 250 \mu\text{m}$ fraction. Horikawa et al. (2015) using

neodymium (Nd) and lead (Pb) isotope data from detrital material from U1341 found that prior to Marine Isotope Stage (MIS) M2 about 3.32 Ma the sediment was likely from local Aleutian sources. This would suggest that the $> 250 \mu\text{m}$ siliciclastics used in this study are similarly from local sources. Horikawa et al. (2015) did find pulses of Alaskan sediment during this interval. Two such pulses occurred at 3.319 Ma and 3.55 Ma. This is close to the increase in $> 250 \mu\text{m}$ siliciclastics abundances at 3.33 Ma and 3.55 Ma found in this study.

2. The Pliocene (5.33-2.58 Ma) is an important global climatic time period. The Pliocene was relatively warm and characterized by increased sea surface temperatures (SST) of $3-4^{\circ}\text{C}$, a sea level 10-20 m higher, a 30% higher concentration of atmospheric carbon dioxide, and a stronger thermohaline circulation than today (Ravelo et al., 2004; Brierley et al., 2009). A cooling trend can be observed (Figure 29) in the global benthic oxygen isotope stack (LR04) during the Pliocene. A noticeable feature of the $\delta^{18}\text{O}$ record occurs at Marine Isotope Stage (MIS) M2 at 3.3 Ma. The sharp change in $\delta^{18}\text{O}$ coincides with an increase in $> 250 \mu\text{m}$ siliciclastics abundances at U1341B. This suggests that at least that portion of the $> 250 \mu\text{m}$ siliciclastics record is the result of ice-rafting. The gradual cooling was interrupted for a brief warm period in the mid-Pliocene (3.29-2.97 Ma). This is supported by the $> 250 \mu\text{m}$ siliciclastics record when abundances are very low. Cooling then continued until intensifying at the Pliocene-Pleistocene boundary around 2.75 Ma in the North Hemisphere (De Schepper et al., 2014).

Figure 29. MAR vs Time Plot. Plot from U1341B with the Global Stack of Benthic $\delta^{18}\text{O}$ Record (Lisiecki and Raymo, 2005). (a) shows the $> 250 \mu\text{m}$ MAR values from U1341B, (b) shows the global stack of benthic $\delta^{18}\text{O}$ record from Lisiecki and Raymo (2005).

Krissek (1995) found the early occurrences of IRD at 6.6 and 5.5 Ma, which he attributes to the Yakataga Formation near the Alaskan coast (Lagoe et al., 1993; De Schepper et al., 2014). Following the warm interval during the mid-Pliocene, glaciation intensified and ice sheets became well developed at 2.75 Ma (Krissek, 1995; Reece et al., 2011). It is widely thought that the onset of NHG occurred gradually and in a stepwise manner (Ravelo et al., 2004; Brigham-Grette et al., 2013; De Schepper et al., 2014). Due to the geographic and atmospheric circulation patterns, southwest Alaska experiences conditions favorable to glaciation. For this reason, the IRD prior to 2.75 found in the North Pacific and Gulf of Alaska are thought to be the result of this limited glaciation on coastal

Alaskan mountains and not an indication of widespread continental glaciation (Prueher and Rea, 2001; Reece et al., 2011; De Schepper et al., 2014). The > 250 μm siliciclastics record from U1341B seems to agree with this overall timing.

Arkhipov et al. (1986) found limited glaciers at sea level on Kamchatka Peninsula during the middle Pleistocene. This means that any IRD prior to then was most likely from Alaska. As glaciation became more extensive and moved northward and to Kamchatka Peninsula, it would begin become a source for IRD. This would account for the change in provenance observed at U1341.

3. Another possibility is that the earlier abundance of > 250 μm siliciclastics does not indicate the beginning of significant glaciation. Instead they represent ephemeral glaciation that was restricted to the mountainous southeast Alaska.

Conclusions

1. An important tool in creating a record of ice-rafting for Bowers Ridge was quantifying the amount of > 250 μm siliciclastics from marine core sediment and creating a > 250 μm concentration vs age plot. While it is not clear that all the > 250 μm sediment from the plot was ice-rafted, it elucidated other major processes that contributed to deposition of sediment > 250 μm at Bowers Ridge.
2. Using a QFL ternary diagram developed by Dickinson and Suczek (1979) was critical in detecting a change in provenance with time. The samples older than 2.75 Ma were from an undissected arc, whereas the younger samples were from a transitional arc.

3. The use of microtexture analysis of quartz grains using SEM was useful in detecting diagnostic surface features that indicate ice related transport mechanisms. While there were clear differences in the ages of the samples as indicated by the degree of weathering, both age groups had a preponderance of large and small fractures and abrasions.
4. Whether the coarse siliciclastic component found in U1341 sediments older than 2.75 Ma is the result of ice-related processes or local re-sedimentation from Bowers Ridge is unclear. It is possible that both processes contributed to its deposition at Bowers Ridge.
5. With the exception of Interval D, the samples younger than 2.75 Ma had significantly more coarse siliciclastics than the samples older than 2.75 Ma. There were also important compositional differences between samples older and younger than 2.75 Ma. The younger samples had more quartz and granites. In contrast, the majority of the older samples were composed of volcanic rocks.
6. The U1341 MAR record, when compared to other records from the North Pacific and Bering Sea, was in agreement with the timing of NHG onset. The biggest differences between Kriisek's (1995) record and U1341 were the amount of coarse sediment found and the peaks detected earlier at U1341.

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