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12 **Millennial-scale climate change and intermediate water circulation in the**
13 **Bering Sea from 90 ka: A high-resolution record from IODP Site U1340**
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Abstract

Millennial-scale climate events in the North Pacific are thought to be related to changes in the circulation of North Pacific Intermediate Water, which may have formed in the Bering Sea in the past. To better constrain past intermediate water circulation and the mechanisms by which millennial-scale events are transmitted, Bering Sea sediment cores from IODP site U1340 were used to construct high-resolution, multi-proxy climate records of the last 90,000 years. Sediment density records show millennial-scale events resembling Dansgaard-Oeschger events, several of which are laminated. Interstadials at U1340 were characterized by 3-5°C warming, increased productivity driven by upwelling, and a reduction in benthic oxygenation. Bering Sea intermediate water also changed over glacial-interglacial timescales; our records show a young, low-salinity intermediate water mass with higher-than-modern oxygen content beginning around 60 ka and persisting until the beginning of the deglaciation. During the Bølling-Allerød, U1340 was characterized by high productivity and laminated sediments, as well as a strong denitrification signature. Paired benthic-planktonic radiocarbon measurements indicate better intermediate water ventilation, implying a decoupling of ventilation and oxygenation during the BA. Our data supports the idea that productivity-derived changes in oxygenation at intermediate water source regions may have contributed to the observed oxygen depletion across the North Pacific during the Bølling-Allerød.

1. Introduction

Global climate over the last 90 kyrs has been characterized by abrupt changes and millennial-scale climate oscillations, including Dansgaard Oeschger (DO) events during the glacial, and the Bølling-Allerød (BA) and Younger Dryas (YD) during the deglaciation. These events, first observed in Greenland ice cores [Dansgaard *et al.*, 1993], are thought to result from complex interactions between ice sheets, sea ice, ocean circulation, and atmospheric greenhouse gases [for review see Clement and Peterson, 2008]. The global nature of these events became clear when sediment cores from the California margin provided evidence that North Pacific benthic oxygenation was synchronous with DO events in Greenland records [Behl and Kennett, 1996; Cannariato and Kennett, 1998]. The causes of these fluctuations in oxygenation have been much investigated in an attempt to identify the processes connecting Atlantic and Pacific millennial-scale climate events.

Hendy and Pederson [2005] found that both export production and intermediate water ventilation influenced benthic oxygenation. Intermediate water at the California margin is affected by the distal extent of two water masses [Hendy and Kennett, 2000]: oxygen-rich North Pacific Intermediate Water (NPIW), which forms in the western North Pacific and circulates at 300-700 m [Talley, 1993], and the low-oxygen California Undercurrent (CU), which is advected from the south at 100-500 m [Hendy *et al.*, 2004; Gay and Chereskin, 2009]. However, Hendy and Pederson (2006), found no evidence that oxygen variations in the Eastern Tropical North Pacific (ETNP) corresponded to those at the California margin, suggesting that California margin oxygen signals were influenced by intermediate water from the *north* and implicating changes in NPIW.

NPIW forms today in the Oyashio and Kuroshio region east of Japan [Talley *et al.*, 1995], but formation and circulation of this water mass in the past is poorly understood. High $\delta^{13}\text{C}$ at intermediate depths during the last glacial suggests that the Bering Sea may have been an area of more active intermediate water formation in the past [Gorbarenko, 1996; Keigwin, 1998; Matsumoto *et al.*, 2002], potentially helping to link high latitude climate to the millennial-scale oceanographic changes observed throughout the North Pacific. To investigate the role of NPIW circulation in millennial-scale climate change in the North Pacific, we generated multi-proxy, high-resolution records of past conditions at Integrated Ocean Drilling Program (IODP) Site U1340 in the Bering Sea.

2. Study area

The Bering Sea is a high productivity marginal sea with high iron, nitrate-limited conditions at the continental shelf, extremely high productivity related to enhanced vertical mixing at the shelf break, and high nitrate, iron-limited conditions in the open sea [Walsh *et al.*, 1989; Banse and English, 1999; Aguilar-Islas *et al.*, 2007], where Site U1340 is located.

The primary source of surface water entering the Bering Sea is the relatively warm Alaskan Stream, which enters the basin through straits between the Aleutian Islands and contributes to the cyclonic gyre circulation in the open basin (Figure 1). Deep and intermediate water transport in/out of the basin occurs through several of the deeper straits. Deep Pacific Water primarily enters the Bering Sea through the Kamchatka Strait at depths below 2500 m (Figure 1) and return flow at ~2500 m is inferred from the distribution of silicic acid [Stabeno *et al.*, 1999].

Temperature, salinity and oxygen profiles from the Bering Sea show that NPIW lacks a strong expression in the basin today, and the depth of the 26.8 σ_θ surface (at which NPIW is

typically centered) is quite shallow (~300 m depth) [Roden, 1995]. However, anthropogenic chlorofluorocarbons in bottom waters indicate that a small amount of ventilation has occurred in the recent past [Warner and Roden, 1995]. Additionally, there is paleoceanographic evidence that Bering Sea intermediate water distribution may have been different in the past; for example, changes in radiolarian assemblages indicate the presence of cold, well-oxygenated intermediate water at deeper water depths during the last glacial period, with some changes over millennial time scales [Wang and Chen, 2005; Tanaka and Takahashi, 2005; Itaki et al., 2009].

3. Materials

3.1. Sediment Cores

This study uses sediment cores collected during IODP Expedition 323 [Takahashi et al., 2011]. Site U1340 is located on Bowers Ridge, an extinct arc system extending 300 km north from the Aleutian Island arc (Figure 1). The water depth at U1340 is 1324 m, in the present day oxygen minimum zone (OMZ) (Figure 2). Samples were taken from the top 24 m of U1340A and U1340D. The splice used for this study differs from that published in the Exp. 323 proceedings [Takahashi et al., 2011] due to better expression of DO-type laminated intervals in U1340A (figure 3), and was constructed based on the correlation of shipboard magnetic susceptibility (alternate affine data: Table 1, alternate splice data: Table 2)

3.2. Lithology

U1340 sediment cores contain alternating layers of diatomaceous ooze and pelagic mud, and include eight distinct laminated intervals (Table 3). Diatomaceous ooze is comprised of well-preserved diatom valves with grain sizes $>15\ \mu\text{m}$, while pelagic mud is finer grained ($<15\ \mu\text{m}$) and mainly composed of less well preserved diatom valves with some fine siliciclastics [Takahashi et al., 2011; Aiello and Ravelo, in review]. Laminated intervals contain $<1\ \text{mm}$ to 6

mm thick lamina of alternating dark green and olive sediments (Figure 3). The lower boundaries between these laminated intervals and the underlying massive sediment, is typically sharp, whereas the upper boundary is more gradual with bioturbation increasing upwards. These laminated layers are high in biogenic opal, leading to low sediment density, and show excellent preservation of whole diatom frustules [Aiello and Ravelo, in review]. While there are only eight laminated intervals, more than 16 low-density, opal-rich intervals are present, and show characteristics similar to the laminated sediments (Figure 4).

4. Methods

4.1. Stable Isotope Analysis

Samples were taken every 5 cm downcore, with additional samples taken every cm in selected laminated intervals. Sediments were freeze-dried and washed through a 63 μm sieve using deionized water. Diatoms were separated from the heavier foraminifera and lithic material through decantation. Benthic foraminifera *Uvigerina peregrina* was picked from the >250 μm fraction, and planktonic foraminifera *Neogloboquadrina pachyderma* (left-coiling) was picked from the >150 μm fraction.

Foraminifera were sonicated in deionized water for 5 seconds (planktonics) or 15 seconds (benthics) to remove particulates. Two to four *U. peregrina* shells, and 10-20 *N. pachyderma* shells were analyzed for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ isotopes using an automated common acid bath carbonate device interfaced to a Fisons Prism III dual-inlet isotope ratio mass spectrometer (IRMS) at UCSC. External precision based on replicates of Carrera Marble is $\pm 0.08\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 0.05\text{‰}$ for $\delta^{13}\text{C}$. Data is expressed relative to the Vienna Pee Dee Belemnite (VPDB).

Bulk $\delta^{15}\text{N}$ was measured on samples taken every ~25 cm downcore and every 1-5 cm through the deglaciation, early Holocene and select laminated intervals. Freeze-dried and

crushed 40 mg samples were analyzed using a Carlo Erba 1108 elemental analyzer interfaced to a Thermo Finningan Delta Plus XP IRMS at UCSC. External precision based on replicate analyses is $\pm 0.02\text{‰}$, and data is expressed relative to atmospheric N_2 .

4.2. Alkenone Analysis

SSTs were estimated for select laminated intervals and the deglaciation and early Holocene using alkenone paleothermometry. Lipids were extracted from 2-3 g of crushed, freeze-dried sediment with dichloromethane using a Dionex ASE 200. Total lipid extracts were dried with a nitrogen stream, then redissolved in 150 μl of toluene spiked with hexatriacontane and heptatriacontane internal standards. Separation of organic compounds was done on a Hewlett-Packard 6890 GC-FID at UCSC following the protocol used by *Dekens et al.* [2007]. Reproducibility of liquid standard replicates included in each run was $\pm 0.007 \text{ U}^{\text{k}'}_{37}$ units. Reproducibility of a Bering Sea sediment standard from IODP site U1342 was $\pm 0.028 \text{ U}^{\text{k}'}_{37}$ units. SSTs were calculated from $\text{U}^{\text{k}'}_{37}$ using the Müller et al. (1998) calibration.

5. Age Model

5.1. Radiocarbon Analysis

For radiocarbon analyses, benthic foraminifera *U. peregrina* and mixed planktonic foraminifera (primarily *N. pachyderma* and *Globigerina bulloides*) were picked from the $>250 \mu\text{m}$ and $>150 \mu\text{m}$ fractions, respectively. In some cases there were not enough *U. peregrina*, and other species were picked and used (see Table 4 for details). Initially, samples were sonicated to remove fine particulates, but this practice caused significant destruction of planktonic tests and was discontinued. Radiocarbon dating was performed at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory. Ages are expressed in conventional

¹⁴C yrs BP using the Libby half-life of 5568, and results include a background subtraction based on ¹⁴C-free calcite and measured $\delta^{13}\text{C}$ (cf. Stuiver and Polach, 1977).

Our age model is based on eight planktonic radiocarbon measurements (Table 4). Radiocarbon ages were converted into calendar years before present with Calib 6.0.1 [Stuiver *et al.*, 2005], using the Marine09 calibration dataset [Reimer *et al.*, 2009] and incorporating a global ocean reservoir correction (R) of 400 yrs. A constant ΔR of 375 yrs [Robinson and Thompson, 1981; Dumond and Griffin, 2002; Kuzmin *et al.*, 2001; Kovanen and Easterbrook, 2002; Cook *et al.*, 2005; Kuzmin *et al.*, 2007; Caissie *et al.*, 2010] was applied prior to calibration. We incorporate a large uncertainty in ΔR (± 300 yrs) because reservoir age in this region is poorly constrained and likely fluctuated temporally during periods like the deglaciation, when deep and intermediate water circulation underwent significant changes [Okazaki *et al.*, 2010; Lund *et al.*, 2011; Thornalley *et al.*, 2011]. Dates between ¹⁴C ages were linearly interpolated. Additionally, we incorporate one tie point to the Lisiecki and Raymo [2005] benthic $\delta^{18}\text{O}$ stack at the beginning of MIS 3 (Figure 5).

5.2. Sedimentation rates

Sedimentation at U1340A averages ~ 30 cm/ kyr (Figure 6). Sedimentation was low during MIS 3 (22 cm/kyr), then high during the LGM (51 cm/kyr) and deglaciation (60 cm/kyr). Sedimentation rates dropped to 32 cm/kyr during the BA, fell further to 12 cm/kyr during the brief YD period, and then increased to 25 cm/kyr during a laminated Pre-Boreal period. Holocene sedimentation rates averaged 15 cm/kyr.

6. Results

6.1. Ventilation Age

Apparent ventilation (Δ_{b-p}) was reconstructed by measuring the radiocarbon age difference (in ^{14}C years) between co-occurring benthic and planktonic samples (Table 4, Figure 7). Modern Δ_{b-p} in the eastern Bering Sea was estimated to be 1200 yrs (using a $\Delta^{14}\text{C}$ profile [Key *et al.*, 2004] to calculate ^{14}C at depth and $R + \Delta R$ to estimate surface ^{14}C). Δ_{b-p} was high during the early deglaciation (~ 9100 yrs), but had fallen to ~ 1000 yrs just prior to the BA. Two measurements during the late BA and the YD show an even further reduction to ~ 350 years, but by the end of the Pre-Boreal laminated interval, Δ_{b-p} had risen to ~ 2400 years.

6.2. Oxygen isotopes

6.2.1. U1340 benthic $\delta^{18}\text{O}$

The benthic $\delta^{18}\text{O}$ record contains two primary features (Figure 8). The first is a pronounced drop of $\sim 0.75\text{‰}$ at the beginning of MIS 3 that lasted until about 35 ka before returning the pre-Holocene average of $\sim 4.5\text{‰}$. The second is a rapid increase to $\sim 5\text{‰}$ at ~ 16 ka, with high values persisting until the onset of BA lamination, when $\delta^{18}\text{O}$ fell sharply. A final drop to the Holocene average of $\sim 3.75\text{‰}$ occurred at the beginning of the Pre-Boreal lamination.

6.2.2. U1340 planktonic $\delta^{18}\text{O}$

The reduction in the benthic $\delta^{18}\text{O}$ record at MIS 3 is also obvious in the planktonic record, but is preceded by an abrupt $\sim 1\text{‰}$ increase at ~ 66 ka that is not present in the benthic record (Figure 8). Low values persisted until the deglaciation, when $\delta^{18}\text{O}$ rose sharply by $\sim 1.1\text{‰}$ to $\sim 4\text{‰}$, and then fell abruptly to 3.1‰ and continued to decline more gradually through the BA and Pre-Boreal laminated periods. Millennial-scale light $\delta^{18}\text{O}$ peaks occurred during MIS 3-4, mostly corresponding to laminated opal-rich intervals. The lack of light $\delta^{18}\text{O}$ peaks during non-laminated opal-rich intervals may be due to lower sampling resolution and bioturbation.

6.3. Carbon isotopes

6.3.1. U1340 benthic $\delta^{13}\text{C}$

Benthic $\delta^{13}\text{C}$ values were lowest (-1.5‰) in the oldest portion of the record. At the beginning of MIS 3, $\delta^{13}\text{C}$ increased to higher values coeval with the reduction in $\delta^{18}\text{O}$ (Figure 8). Values averaged -1.1‰ , until the deglaciation when $\delta^{13}\text{C}$ dropped back to -1.5‰ . During the BA, $\delta^{13}\text{C}$ values increased slightly, peaking at -0.75‰ , then dropped to -1.2‰ during the YD and Pre-Boreal. Holocene values averaged -0.75‰ at ~ 9.5 ka.

6.3.2. U1340 planktonic $\delta^{13}\text{C}$

Planktonic $\delta^{13}\text{C}$ trends largely replicate benthic $\delta^{13}\text{C}$ (Figure 8). Lowest values (-0.2‰) occurred prior to MIS 3, when $\delta^{13}\text{C}$ began rising gradually. A maximum of $\sim 0.2\text{‰}$ occurred at 51 ka, and $\delta^{13}\text{C}$ values remained elevated until the deglaciation. A brief negative excursion to -0.2‰ occurred at ~ 16 ka and persisted for 800 years, after which $\delta^{13}\text{C}$ rose again to $\sim 0.3\text{‰}$. During the BA and YD, $\delta^{13}\text{C}$ rose slowly to the Holocene average of $\sim 0.5\text{‰}$. Millennial-scale low $\delta^{13}\text{C}$ spikes of relatively small amplitude (~ 0.2 - 0.5‰) occurred during MIS 3-4 in most of the laminated, opal-rich intervals. The lack of light $\delta^{18}\text{C}$ values during non-laminated opal-rich intervals may be due to lower sampling resolution and bioturbation.

6.4. U1340 bulk $\delta^{15}\text{N}$

Bulk $\delta^{15}\text{N}$ was relatively high in the oldest part of the record, dropped at the beginning of MIS 3, and then began rising again at ~ 27 ka (Figure 8). LGM and deglaciation values averaged $\sim 5.5\text{‰}$. During the BA, $\delta^{15}\text{N}$ rose rapidly, peaking at 8.24‰ , then decreased to values of $\sim 6.7\text{‰}$ and remained relatively constant through the late BA and the YD. Midway through the Pre-Boreal lamination, $\delta^{15}\text{N}$ began to fall again, stabilizing at the Holocene average of $\sim 4.4\text{‰}$. Sharp $\delta^{15}\text{N}$ peaks occurred during three laminated intervals, with magnitudes of 1.5 - 2‰ and maximum

values of 6-7‰ (Figure 8). The lack of dramatic $\delta^{15}\text{N}$ peaks during non-laminated low-density intervals may be due to lower sampling resolution and bioturbation.

6.5. U1340 alkenone temperatures

Although it was possible to detect alkenones in some samples, the concentrations were generally low, possibly due to the fact that *Coccolithus pelagicus* (non-alkenone producing) is more abundant in the open Bering Sea than *Emiliania huxleyi* (alkenone producing) [Takahashi *et al.*, 2000]. Nevertheless, alkenone studies in the Bering Sea have found good agreement of $\text{U}^{k'}_{37}$ -based temperatures with observed spring/fall surface and mixed layer temperatures [Shin *et al.*, 2002; Harada *et al.*, 2003], and we were able to constrain sea surface temperature (SST) in some intervals.

SST during laminated intervals at U1340 are characterized by sharp, 5-6°C increases, from ~4°C during stadials (non-laminated intervals) to peak interstadial temperatures of ~9°C. LGM temperatures averaged ~6°C, and deglacial warming began around 19 ka (Figure 9). Temperatures fluctuated considerably through the deglaciation, averaging ~8.5°C. BA warming began with the onset of laminated sediments and peaked at 11.2°C. Temperatures fell briefly to ~8°C during the YD, then rose to 10.7°C during the Pre-Boreal lamination. Holocene temperatures averaged ~7.7°C.

7. Discussion

7.1. Bering Sea Paleoenvironment since the Last Glacial Maximum

7.1.1. Chronological issues

Deglacial sediments at U1340 feature two low-density laminated sections separated by a brief bioturbated interval (Figure 7). This sedimentary sequence is also seen in many well-dated Bering Sea and North Pacific cores, with laminated intervals dating to the BA and Pre-Boreal

warm periods on either side of bioturbated sediment of YD age [Behl and Kennett, 1996, Zheng et al., 2000; Cook et al., 2005; Ikehara et al., 2006; Brunelle et al., 2007; Ishizaki et al., 2009; Kim et al., 2011]. Our reconstruction of environmental conditions at U1340 during the BA laminated interval is consistent with Pacific-wide climate conditions during this period, providing convincing evidence that we have correctly identified the BA (see section 7.1.3.). However, our ^{14}C age model (which assumes a constant ΔR) indicates that the onset of lamination occurred at ~ 15.3 ka, 600 years prior to the start of the BA in well-dated Greenland ice cores [Alley et al., 1993] and California margin sediments [Hendy et al., 2002].

Evidence for similar apparent early warming has been observed in other records [Seki et al., 2002; Kiefer and Keinast, 2005; Sarnthein et al., 2006] where calibrated radiocarbon ages (assuming no changes in ΔR) are used and the early warming is assumed to be real. We suggest that a more reasonable explanation for the apparent lead in these and our records is local changes in ΔR . Several northeastern Pacific studies have calculated unusually old reservoir ages (1100-1800 yrs) during the deglaciation, BA, and YD [Kovanen and Easterbrook, 2002; Cosma and Hendy, 2008], in contrast to modern reservoir ages of ~ 800 yrs. By assuming that the laminated BA interval at U1340 occurred concurrently with the BA in Greenland ice cores, the local reservoir age during this period can be estimated to be 1300-1400 yrs (within the range estimated by Kovanen and Easterbrook [2002] and Cosma and Hendy [2008]). The apparent early onset of the BA at U1340 and other subarctic Pacific sites may thus simply be the result of uncertainties and underestimation of reservoir age.

7.1.2. The LGM and the Deglaciation

The deglaciation at U1340 is characterized by a warm-cold pattern very similar to that observed in other subarctic Pacific records (Figure 9). Rapid warming occurred at 19.3 ka,

followed by cooling from 16-15.3 ka, just prior to the BA. This pattern was noted by *Kiefer and Kienast* [2005] in other subarctic Pacific records of the deglaciation, and small differences in timing between records can be explained by uncertainties in the radiocarbon age models and reservoir age assumptions for each of these studies.

$\delta^{18}\text{O}$ values were extremely high during the deglaciation, with planktonic $\delta^{18}\text{O} > 1.5\text{‰}$ higher than Holocene values (Figure 9). Although some of the $\delta^{18}\text{O}$ shift can be attributed to eustatic ice volume changes [eg. *Fairbanks*, 1989], the marked high $\delta^{18}\text{O}$ values from 18-16 ka indicate relatively high local $\delta^{18}\text{O}$ (and salinity) and/or cool temperatures consistent with a southward expansion of sea ice in the LGM [*Tanaka and Takahashi*, 2005] and lasting until 16 ka. Benthic $\delta^{18}\text{O}$ trends after 16 ka are more akin to a (mostly) eustatic component and track well with the *Lisiecki and Raymo* [2005] benthic stack (Figure 5). Prior to 18 ka, U1340 benthic $\delta^{18}\text{O}$ diverged significantly from the global trend; these differences are discussed in section 7.2.).

7.1.3. The Bølling-Allerød, Younger Dryas and early Holocene

Throughout the North Pacific, the BA warm period (14.7-12.8 ka) was characterized by reduced subsurface oxygenation and the preservation of laminations [*Kennett and Ingram*, 1995; *Behl and Kennett*, 1996; *Cannariato and Kennett*, 1999; *Zheng et al.*, 2000; *Cook et al.*, 2005; *Ikehara et al.*, 2006; *Brunelle et al.*, 2007; *Ishizaki et al.*, 2009, *Kim et al.*, 2011]. Changes in oxygenation arose from changes in primary productivity [*Mix et al.*, 1999; *Stott et al.*, 2000; *Crusius et al.*, 2004; *Dean*, 2007], NPIW ventilation [*Duplessy et al.*, 1988; *Kennett and Ingram*, 1995; *Zheng et al.*, 2000; *Ahagon et al.*, 2003; *Sagawa and Ikehara*, 2008] or some combination of the two [*Hendy and Pedersen*, 2005; *Ishizaki et al.*, 2009, *Kim et al.*, 2011].

At U1340, relatively high productivity during the BA is inferred from low sediment density (Figure 9) indicative of higher biogenic opal content [*Takahashi et al.*, 2011] and from

the presence of ubiquitous well-preserved whole diatom frustules compared to the highly fragmented diatom values that dominate the massive sediment intervals before and after the BA [Aiello and Ravelo, in review]. Additionally, the occurrence of *Chaetoceros* resting and vegetative spores, indicative of high primary productivity, increases sharply with the onset of lamination (Table 5). Kim *et al.* [2011] provide a mechanism for this change, suggesting that a stronger Bering Slope Current and increased glacial meltwater led to a greater supply of nutrients and thus high productivity during the BA. This idea is supported at U1340 by an sharp increase in the abundance of *Neodenticula seminae*, a species indicative of Alaskan Stream waters, at the onset of lamination (Table 5). This was likely spurred by rising sea levels and enhanced flow into the Bering Sea, which could have contributed to the stronger Bering Slope Current proposed by Kim *et al.* [2011]. In support of Crusius *et al.* [2004], ours results indicate that high productivity in the western subarctic Pacific during the BA reduced intermediate water oxygenation, providing an explanation for the observed North Pacific-wide subsurface oxygen minima without invoking reduced ventilation.

The onset of the BA was also characterized by a rapid 5°C warming that peaked at ~11°C (Figure 9) and a sharp reduction in the occurrence of the resting spores of sea-related diatom species *Thalssiosira Antarctica*, implying reduced spring sea ice cover (Table 5). Planktonic $\delta^{18}\text{O}$ decreased abruptly ~130 yrs prior to the rise in temperature, most likely indicating freshening along with the effect of reduced global ice volume.

Local/regional denitrification appears to have intensified in the Bering Sea during the BA. Bulk $\delta^{15}\text{N}$ at U1340 rose sharply to 8.24‰ during the early BA (Figure 9), and although productivity likewise increased, higher nitrate utilization cannot fully explain this change. When surface nitrate is completely utilized, the accumulating particulate organic matter bears the

isotopic signature of the source nitrate [Sigman *et al.*, 2009] which is ~5-6‰ in the Bering Sea today [Lehman *et al.*, 2005]. It is thus unlikely that enhanced nitrate utilization produced the 7-8‰ values of the BA without a concurrent change in the $\delta^{15}\text{N}$ of the nitrate source. In addition, changes in utilization typically affect both $\delta^{15}\text{N}$ and $\delta^{13}\text{C}$ of DIC, and U1340 values of $\delta^{13}\text{C}$ of planktonic foraminifera (which reflect the $\delta^{13}\text{C}$ of DIC) were constant through the BA (Figure 8). This suggests that the U1340 $\delta^{15}\text{N}$ signature was caused by water column denitrification, which significantly raises $\delta^{15}\text{N}$ [Sigman *et al.*, 2009]. Our data is consistent with another Bering Sea study that measured 8-9‰ peaks in bulk $\delta^{15}\text{N}$ during the BA [Brunelle *et al.*, 2007] and with research suggesting a widespread rise in denitrification during the BA [Emmer and Thunell, 2000; Kienast *et al.*, 2002; Kao *et al.*, 2008; Brunelle *et al.*, 2010]. The denitrification signature at U1340 may have been transmitted to the Bering Sea via intermediate water originating elsewhere in the region, but the laminations at U1340 suggest that oxygenation was low enough for water column denitrification to occur locally.

The coherent light-density BA interval is comprised of two distinct laminated sections separated by 23 cm (~60 yrs) of organic-rich bioturbated sediments (Figure 9). This non-laminated layer may represent the brief Older Dryas cool period, which occurs in some records between the Bølling and Allerød oscillations, or the Inter-Allerød Cold Period, an even shorter cool interval during the Allerød [Benson *et al.*, 1997]. This lapse in lamination may be indicative of a slight reduction in productivity, such that oxygen was not completely depleted at depth.

The YD is not a prominent feature of the U1340 records, and is identified as the period immediately following BA lamination and preceding a second, Pre-Boreal lamination (Figure 7). Sedimentation dropped sharply to ~10 cm/kyr during this 11 cm interval (Figure 6), thus the YD is represented by only a few data points. Bioturbation during this period likely destroyed the top

few cm of underlying BA laminated sediments, introducing BA-aged organic material into the YD sediments and skewing the data towards characteristics more typical of the BA. Relatively high planktonic $\delta^{18}\text{O}$ values and one relatively cool alkenone SST data point suggest a cooling of 1-2°C during the YD (Figure 9).

7.1.4. Intermediate water ventilation during the deglaciation and early Holocene

Changes in ocean circulation and heat transport are thought to be important in millennial-scale climate change during the deglaciation. In the North Atlantic, an inferred reduction or shutdown in NADW formation occurred during cold periods like the YD [Keigwin and Lehman, 1994; McManus *et al.*, 2004] while ventilation increased during the BA and the Holocene [Robinson *et al.*, 2005; Thornalley *et al.*, 2011]. In contrast, North Pacific data suggests that intermediate depth ventilation was stronger during cold periods [Duplessy *et al.*, 1989; Ahagon *et al.*, 2003; Sagawa and Ikehara, 2008], consistent with a modeling study by Mikolajewicz *et al.*, [1997]. Okazaki *et al* [2010] reproduce this anti-phase pattern of overturning circulation in a second model, and suggest that it could have been important in maintaining poleward heat transport when NADW formation was reduced.

Our Δ_{b-p} apparent ventilation reconstruction differs significantly from the anti-phase model proposed by Okazaki *et al.* [2010]. U1340 apparent ventilation age was lowest (~350 yrs) in two samples from the YD (Figure 7), which likely incorporate, through bioturbation, a significant amount of BA-age foraminifera (see section 7.1.3.). These ventilation values are typical of regions with active overturning and well-ventilated intermediate/deep waters [Matsumoto, 2007]. Apparent ventilation age at U1320 was also young (~900 yrs) in the older BA, when North Atlantic overturning was strong and North Pacific records indicate old ventilation ages. Furthermore, the highest ventilation ages at U1340 occurred during the

Holocene (~2500 yrs) and early deglaciation (9000 yrs), in contrast with the low ventilation ages observed elsewhere in the North Pacific during these periods [for review see *Okazaki et al.*, 2010]. This disparity between U1340 and other North Pacific sites suggests that U1340 data reflects local ventilation changes. Further ventilation age reconstructions at intermediate-depth sites elsewhere in the western North Pacific would aid in tracing the path of the well-ventilated water mass we identify at U1340.

The U1340 data strongly supports the idea that high productivity in NPIW source regions could decouple NPIW oxygenation and ventilation during the BA [*Crusius et al.*, 2004]. During the BA, there was high primary productivity and suboxic intermediate water at U1340, but very young ventilation age. A radiocarbon dataset from the Okhotsk Sea and Emperor Seamounts also shows young ventilation ages at 1-2km depths during the BA [*Cook and Keigwin*, in prep]. Laminations during this period thus appear to have been caused primarily by increased productivity, rather than poor intermediate water ventilation.

The $\Delta_{b,p}$ value of 9000 years at 17.8 ka (Figure 7) is much older than typically observed, and reminiscent of the anomalously old intermediate and abyssal ventilation data from elsewhere in the Pacific [*Sikes et al.*, 2000; *Marchitto et al.*, 2007; *Stott et al.*, 2009]. These irregularities may have resulted from the release of an isolated, ^{14}C -deplete carbon reservoir from the deep ocean during the “Mystery Interval” [17.5-14.5 ka, *Denton et al.*, 2006], when atmospheric $\Delta^{14}\text{C}$ dropped by ~190‰ [*Beck et al.*, 2001; *Hughen et al.*, 2004; *Fairbanks et al.*, 2005; *Broecker and Barker*, 2007]. However, there is no evidence for an anomalously old water mass in the equatorial or South Pacific during this period [*Broecker et al.*, 2004; *Broecker et al.*, 2008; *De Pol-Holz et al.*, 2010]. Model simulations suggest that the existence of an abyssal, radiocarbon-deplete reservoir is unlikely and insufficient to explain the observed intermediate depth

radiocarbon anomalies [Hain *et al.*, 2011]. Although the existence of this old carbon reservoir thus remains an open question, U1340 data may suggest an Arctic or northern source for any such low- ^{14}C water mass.

7.2. Intermediate water formation during MIS 3

At ~60ka, there are pronounced shifts in the U1340 stable isotope records toward lower planktonic and benthic $\delta^{18}\text{O}$ and bulk $\delta^{15}\text{N}$ values, and higher planktonic and benthic $\delta^{13}\text{C}$ values (Figure 10). These trends are absent from California margin and North Atlantic $\delta^{18}\text{O}$ records (Figure 11) [McManus *et al.*, 1999; Hendy *et al.*, 2002; Hendy and Kennett, 2003; Lisiecki and Raymo, 2005], and from North Pacific $\delta^{15}\text{N}$ records (Figure 12) [Emmer and Thunell, 2000; Kienast *et al.*, 2002; Hendy *et al.*, 2004; Brunelle *et al.*, 2010], implying that U1340 records reflect regional or local processes rather than global trends. The isotopic shifts of at 60ka could be explained by a change in either biogeochemical cycling (e.g., productivity, nutrient/particulate organic matter fluxes, etc) or in preformed water mass characteristics throughout the water column at the site. The first possibility is unlikely because sediment lithology does not change, as indicated by the sediment density record (Figure 8) and a nearby biogenic opal accumulation record [Brunelle *et al.*, 2007]. Also, the benthic and planktonic $\delta^{13}\text{C}$ trends are in the same direction and of the same large magnitude (0.6-0.75‰, Figure 10) whereas a change in carbon flux should produce opposing trends. Finally, it is unclear how a biogeochemical process could have affected the planktonic and benthic $\delta^{18}\text{O}$ records to such a degree. Rather, the shifts in isotope records at ~60ka likely reflect changes in the preformed water mass characteristics affecting site U1340.

Higher $\delta^{13}\text{C}$ values after 60 ka indicate a greater contribution of ‘young’ water, which typically has high oxygen concentrations. The decrease in $\delta^{18}\text{O}$ indicates that this water mass was

either warmer by 3-4°C or had lower $\delta^{18}\text{O}_{\text{water}}$ which would translate to lower salinity. Limited U1340 alkenone data shows no significant change in baseline temperatures, and Bering Sea radiolarian assemblages (discussed below) indicate the presence of cold intermediate water during the last glacial. Thus, the shift toward lower planktonic and benthic $\delta^{18}\text{O}$ values probably reflects a decrease in salinity in the surface and at depth. Together, U1340 records indicate the presence of a cold, low-salinity, well-oxygenated intermediate water mass, implying that NPIW could have been more dominant and extensive from 60ka until the deglaciation (MIS3-2) compared to the modern ocean.

The presence of newly formed intermediate water in the Bering Sea during MIS 3-2 is supported by the abundances of radiolarian *Cycladophora davisiana* in Bering Sea sediments. Today, *C. davisiana* is abundant (>20%) only in the Okhotsk Sea, particularly in cold, low salinity intermediate water at depths of 300-1000 m where it is sustained by high microbial biomass at these depths [Nimmergut and Abelman, 2002; Okazaki et al., 2003]. This biomass is supported by a high flux of particulate organic carbon related to the formation of seasonal sea ice and brine [Okazaki et al., 2003]. *C. davisiana* is used in paleoclimate to identify cold, fresh intermediate water. High relative abundances in the Bering Sea during MIS 2-3 support the presence of NPIW-like water [Tanaka and Takahashi, 2005], and provide a mechanism (increased seasonal sea ice and brine formation) for local intermediate water formation.

U1340 $\delta^{15}\text{N}$ is also consistent with local intermediate water formation. The rapid, 2‰ drop in bulk $\delta^{15}\text{N}$ (Figure 10) is most likely related to reduced nutrient utilization [Altabet, and Francois, 1994], caused by a drop in productivity or an increase in nitrate availability via vertical mixing. The lack of a clear change in biogenic opal productivity at another Bowers Ridge site [Brunelle et al., 2007] suggests that the U1340 $\delta^{15}\text{N}$ signal reflects higher nutrients, most likely

caused by a reduction in Bering Sea stratification. Reduced stratification would provide an environment ideal for intermediate water formation, in contrast to the open northwest Pacific, where conditions were more stratified and nutrient utilization was higher [Brunelle *et al.*, 2010; Jaccard *et al.*, 2005]. The lack of a low nitrate utilization signal outside of the Bering Sea is compelling evidence for local, rather than subarctic Pacific-wide, changes in stratification and intermediate water formation.

Bulk sediment $\delta^{15}\text{N}$ and diatom-bound $\delta^{15}\text{N}$ ($\delta^{15}\text{N}_{\text{db}}$) records from a different Bowers Ridge site, JPC17 (2209 m), lacks the pronounced drop observed at U1340 (Figure 13) [Brunelle *et al.*, 2007]. However, diagenetic effects [Sigman *et al.*, 2009] could have had a greater impact on the $\delta^{15}\text{N}$ of bulk sediment record at JPC17, which is deeper and more oxygenated than U1340. Furthermore, although $\delta^{15}\text{N}_{\text{db}}$ is thought to be impervious to diagenetic alteration [Sigman *et al.*, 1999], species-specific vital effects during biomineralization are poorly quantified, and changes in $\delta^{15}\text{N}_{\text{db}}$ may thus incorporate the effects of shifting diatom assemblages. Lower sedimentation (~12 cm/kyr at JPC17 vs. ~30 cm/kyr at U1340) may also obscure sharp changes in JPC17 records.

The influence of well-ventilated intermediate water at U1340 appears to have ended during the deglaciation, between 20-17 ka. Benthic $\delta^{18}\text{O}$ began to rise at 20 ka, while planktonic $\delta^{18}\text{O}$ rose sharply at 18 ka (Figure 10). U1340 alkenones showed no clear temperature trends during this interval (Figure 9), suggesting that the $\delta^{18}\text{O}$ signal was caused by rising salinity. Benthic $\delta^{13}\text{C}$ dropped from 18-16.5 ka, while planktonic $\delta^{13}\text{C}$ fell abruptly at 18 ka (Figure 10). Changes in all of these proxies towards the typical values observed prior to MIS 3 suggest the end of active intermediate water formation in the Bering Sea. This change may have been

influenced by increasing stratification due to warming surface temperatures and the input of glacial meltwater.

7.3. Characteristics of millennial-scale events

Approximately 15 millennial-scale low-density events prior to 20ka were identified at U1340 (Table 3, Figure 4). These appear to be high productivity events based on their high biogenic opal content and enhanced preservation of diatom valves [*Takahashi et al.*, 2011; *Aiello and Ravelo*, in review]. Five of these events were accompanied by laminations similar to those occurring elsewhere in the North Pacific during DO events [e.g., *Behl and Kennett*, 1996], providing striking evidence for significant variations in bottom water oxygenation at U1340. Correlation of each millennial-scale event at U1340 with each specific DO interstadial in the Greenland ice core record was not possible, likely due to uncertainty in our age model beyond the oldest radiocarbon date (~29 ka). With average sedimentation rates of 22cm/kyr, it is also possible that bioturbation obscured some events when anoxia did not develop. Nonetheless, because the low-density peaks at U1340 occur over millennial time-scales and number roughly the same as DO events, we refer to these intervals as “DO-type” events.

DO-type events at U1340 were characterized by warmer, fresher surface waters. The abrupt transition to laminated sediments was accompanied by a sharp, 0.7-0.9‰ drop in planktonic $\delta^{18}\text{O}$ (Figure 14), representing 3-4°C warming if attributed solely to temperature fluctuations. This is roughly consistent with the alkenone-derived temperature record, which indicates ~5°C warming during laminated intervals (Figure 14). Negative $\delta^{18}\text{O}$ peaks were probably also influenced by a freshening of surface waters due to sea ice melting and/or increased runoff into the Bering Sea. Benthic $\delta^{18}\text{O}$ remained constant, implying that temperature/salinity changes were limited to surface waters. Our results are in contrast to a

Emperor Seamounts study by *Kiefer et al.* [2001] suggesting an anti-phase pattern for DO events in the North Atlantic and the North Pacific

Sediment lithology and $\delta^{13}\text{C}$ records suggest increased productivity during DO-type events, with several laminations containing negative planktonic $\delta^{13}\text{C}$ peaks (Figure 14). During one laminated interval, the percentage of fragmented diatom valves dropped from ~75% to ~3%, indicating much higher preservation and suggesting increased opal flux, consistent with a rise in productivity. Given this productivity signal, we attribute negative $\delta^{13}\text{C}$ peaks to amplified upwelling. Higher export production is also consistent with the reduction in benthic oxygen that is evident during DO-type events.

The upwelling signature (and implied nutrient-rich environment) at U1340 during DO-type events suggests that co-occurring high $\delta^{15}\text{N}$ peaks were caused by a change in the preformed isotopic signature of subsurface nitrate rather than increased nutrient utilization. Peak values of 5.6-6.9‰ (Figure 14) cannot be accounted for by complete nutrient utilization if subsurface nitrate retained its modern signature (5-6‰, [*Lehman et al.*, 2005]). Further, it seems doubtful that nitrate would be completely utilized, given that the Bering Sea today is a HNLC, iron-limited environment with only 50% nitrate utilization [*Brunelle et al.*, 2007]. The $\delta^{15}\text{N}$ peaks at U1340 were therefore most likely due to a denitrification signature in subsurface nitrate during DO-type events. This denitrification signal may have originated locally, given the suboxic benthic conditions, or elsewhere in the North Pacific, suggesting possible changes in subsurface circulation during DO-type events. Further high-resolution studies in the subarctic Pacific are necessary to reconstruct any millennial-scale changes in regional denitrification.

8. Summary

1. The Bølling-Allerød at U1340 was characterized by rapid warming, high primary productivity, suboxic benthic conditions, and local or regional denitrification. Intermediate waters were well-ventilated, indicating the decoupling of ventilation age and oxygenation during this period. Our radiocarbon age model places the onset of the BA at U1340 several hundred years earlier than the North Atlantic, most likely a result of large fluctuations in reservoir age in the North Pacific during the deglaciation.

2. From 60 ka through the deglaciation there was a strong presence of cold, low-salinity, well-oxygenated intermediate water at U1340. The Bering Sea also appears to have been less well stratified than elsewhere in the subarctic Pacific during this period, suggesting that conditions were ideal for this intermediate water mass to have formed locally.

3. Dansgaard-Oeschger interstadials at U1340 were characterized by rapid warming and freshening, along with high upwelling-driven productivity that contributed to benthic suboxia/anoxia and led to enhanced water column denitrification (either locally or regionally).

4. Our reconstruction of Bering Sea conditions during the last 85 kyrs indicates that changes in intermediate water formation and ventilation have occurred over both glacial-interglacial and millennial timescales, implying the subarctic northwest Pacific has the potential to play an active role in global climate change.

5. The anoxia and denitrification signals present at U1340 during the BA and DO events show that denitrification occurred more broadly in the past over varied timescales, and is potentially an important dynamic feedback for global climate. During the timescale of our study, this feedback appears to be positive, with denitrification occurring during warm periods when a reduction in bioavailable nitrogen could slow the carbon pump and result in higher atmospheric greenhouse gases (N₂O and CO₂).

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References

- Aguilar-Islas A. M., M. P. Hurst, K. N. Buck, B. Sohst, G. J. Smith, M. C. Lohan, and K. W. Bruland (2007), Micro- and macronutrients in the southeastern Bering Sea: Insight into iron-replete and iron-depleted regimes, *Prog. Oceanogr.*, 73, 99-126, doi:10.1016/j.pocean.2006.12.002.
- Ahagon, N., K. Ohkushi, M. Uchida, and T. Mishima (2003), Mid-depth circulation in the northwest Pacific during the last deglaciation: Evidence from foraminiferal radiocarbon ages, *Geophys. Res. Lett.*, 30, 2097, doi:10.1029/2003GL018287.
- Alley, R. B., D. A. Meese, C. A. Shuman, A. J. Gow, K. C. Taylor, P. M. Grootes, J. W. C. White, M. Ram, E. D. Waddington, P. A. Mayewski and G. A. Zielinski (1993), Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event, *Nature*, 362, 527-529, doi:10.1038/362527a0.
- Altabet, M. A., and R. Francois (1994), Sedimentary nitrogen isotopic ratio as a recorder for surface ocean nitrate utilization, *Global Biogeochem. Cycles*, 8, 103-116, doi:10.1029/93GB03396.
- Banase, K. and D. C. English (1999), Comparing phytoplankton seasonality in the eastern and western subarctic Pacific and the western Bering Sea, *Prog. Oceanogr.*, 43, 235-288, doi:10.1016/S0079-6611(99)00010-5
- Beck, J. W., D. A. Richards, R. L. Edwards, B. W. Silverman, P. L. Smart, D. J. Donahue, S. Hererra-Osterheld, G. S. Burr, L. Calsoyas, A. J. T. Jull, and D. Biddulph (2001), Extremely large variations of atmospheric ^{14}C concentration during the last glacial period, *Science*, 292(5526), 2453-2458, doi:10.1126/science.1056649.
- Behl, R. J., and J. P. Kennett (1996), Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr, *Nature*, 379, 243-246, doi:10.1038/379243a0.
- Benson, L., J. Burdett, S. Lund, M. Kashgarian and S. Mensing (1997), Nearly synchronous climate change in the Northern Hemisphere during the last glacial termination, *Nature*, 388, 263-265,
- Broecker, W., [S. Barker](#), [E. Clark](#), [I. Hajdas](#), [G. Bonani](#) and [L. Stott](#) (2004), Ventilation of the glacial deep Pacific Ocean, *Science*, 306(5699), 1169-1172, doi:10.1126/science.1102293.
- Broecker, W., and S. Barker (2007), A 190‰ drop in atmosphere's ^{14}C during the “mystery interval” (17.5 to 14.5 kyr), *Earth Planet. Sci. Lett.*, 256(1-2), 90-99, doi:10.1016/j.epsl.2007.01.015.

- Broecker, W., E. Clark and S. Barker (2008) Near constancy of the Pacific Ocean surface to mid-depth radiocarbon-age difference over the last 20 kyr, *Earth Planet. Sci. Lett.*, 274(3-4), 322-326, [doi:10.1016/j.epsl.2008.07.035](https://doi.org/10.1016/j.epsl.2008.07.035).
- Brunelle, B. G., D. M. Sigman, M. S. Cook, L. D. Keigwin, G. H. Haug, B. Plessen, G. Schettler, and S. L. Jaccard (2007), Evidence from diatom-bound nitrogen isotopes for subarctic Pacific stratification during the last ice age and a link to North Pacific denitrification changes, *Paleoceanography*, 22, PA1215, doi:10.1029/2005PA001205.
- Brunelle, B. G., D. M. Sigman, S. L. Jaccard, L. D. Keigwin, B. Plessen, G. Schettler, M. S. Cook, and G. H. Haug (2010), Glacial/interglacial changes in nutrient supply and stratification in the western subarctic North Pacific since the penultimate glacial maximum, *Quat. Sci. Rev.*, 29, 2579-2590, doi:10.1016/j.quascirev.2010.03.010.
- Cannariato, K.G., and J.P. Kennett (1999), Climatically related millennial-scale fluctuations in strength of California margin oxygen-minimum zone during the past 60 K.y., *Geology*, 27, 975-978, doi: 10.1130/0091-7613(1999)027<0975:CRMSFI>2.3.CO;2 .
- Caissie, B. E., J. Brigham-Grette, K. T. Lawrence, T. D. Herbert, and M. S. Cook (2010), Last Glacial Maximum to Holocene sea surface conditions at Umnak Plateau, Bering Sea, as inferred from diatom, alkenone, and stable isotope records, *Paleoceanography*, 25, PA1206, doi:10.1029/2008PA001671.
- Clement, A. C., and L. C. Peterson (2008), Mechanisms of abrupt climate change of the last glacial period, *Rev. Geophys.*, 46, RG4002, doi:10.1029/2006RG000204.
- Cook, M.S., L.D. Keigwin, and C.A. Sancetta (2005), The deglacial history of surface and intermediate water of the Bering Sea, *Deep-Sea Res. II*, 52, 2163-2173, doi:10.1016/j.dsr2.2005.07.004.
- Cosma, T.N., I.L. Hendy, and A.S. Chang (2008), Chronological constraints on Cordilleran Ice Sheet glaciomarine sedimentation from core MD02-2496 off Vancouver Island (western Canada), *Quat. Sci. Rev.*, 27, 941-955, doi:10.1016/j.quascirev.2008.01.013.
- Crusius, J., T.F. Pedersen, S. Kienast, L. Keigwin, and L. Labeyrie (2004), Influence of northwest Pacific productivity on North Pacific Intermediate Water oxygen concentrations during the Bolling-Allerod interval (14.7-12.9 ka), *Geology*, 32, 633-636, doi:10.1130/G20508.1.
- De Pol-Holz, R., L. Keigwin, J. Southon, D. Hebbeln and M. Mohtadi (2010), No signature of abyssal carbon in intermediate waters off Chile during deglaciation, *Nature Geoscience*, 3, 192-195, doi:10.1038/ngeo745.
- Dansgaard, W., S. J. Johnsen, H. B. Clausen, D. Dahl-Jensen, N. S. Gundestrup, C. U. Hammer, C. S. Hvidberg, J. P. Steffensen, A. E. Sveinbjörnsdottir, J. Jouzel and G. Bond (1993),

- Evidence for general instability of past climate from a 250-kyr ice-core record, *Nature*, 364, 218-220, doi:10.1038/364218a0.
- Dean, W. E. (2007), Sediment geochemical records of productivity and oxygen depletion along the margin of western North America during the past 60,000 years: teleconnections with Greenland Ice and the Cariaco Basin, *Quat. Sci. Rev.*, 26, 98-114, doi:10.1016/j.quascirev.2006.08.006.
- Dekens, P. S., A. C. Ravelo, and M. D. McCarthy (2007), Warm upwelling regions in the Pliocene warm period, *Paleoceanography*, 22, PA3211, doi:10.1029/2006PA001394.
- Denton, G., W. Broecker, and R. Alley (2006), The mystery interval 17.5 to 14.5 kyrs, *PAGES News*, 13(2), 14–16.
- Dumond, D. E. and D. G. Griffin (2002), Measurements of the marine reservoir effect on radiocarbon ages in the Eastern Bering Sea, *Arctic*, 55, 77-86.
- Duplessy, J.C., N.J. Shackleton, R.G. Fairbanks, L. Labeyrie, and D. Oppo (1988), Deep water source variations during the last climatic cycle and their impact on the global deep water circulation, *Paleoceanography*, 3, 343-360.
- Emmer, E., and R. C. Thunell (2000), Nitrogen isotope variations in Santa Barbara Basin sediments: Implications for denitrification in the eastern tropical North Pacific during the last 50,000 years, *Paleoceanography*, 15, 377-387, doi:10.1029/1999PA000417.
- Expedition 323 Scientists (2010), Bering Sea paleoceanography: Pliocene–Pleistocene paleoceanography and climate history of the Bering Sea, *IODP Prel. Rept.*, 323. doi:10.2204/iodp.pr.323.2010.
- Fairbanks, R.G. (1989), A 17,000 year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep ocean circulation, *Nature*, 342, 637-642.
- Fairbanks, R. G., R. A. Mortlock, T.-C. Chiu, L. Cao, A. Kaplan, T. P. Guilderson, T. W. Fairbanks, A. L. Bloom, P. M. Grootes, and M.-J. Nadeau (2005), Radiocarbon calibration curve spanning 0 to 50,000 years BP based on paired $^{230}\text{Th}/^{234}\text{U}/^{238}\text{U}$ and ^{14}C dates on pristine corals, *Quat. Sci. Rev.*, 24(16–17), 1781-1796, doi:10.1016/j.quascirev.2005.04.007.
- Gay, P. S., and T. K. Chereskin (2009), Mean structure and seasonal variability of the poleward undercurrent off southern California, *J. Geophys. Res.*, 114, C02007, doi:10.1029/2008JC004886.
- Gorbarenko, S. A. (1996), Stable isotope and lithologic evidence of late-glacial and Holocene oceanography of the northwestern Pacific and its marginal seas, *Quat. Res.*, 46, 230-250, doi:10.1006/qres.1996.0063.

- 682
683 Hain, M. P., D. M. Sigman, and G. H. Haug (2011), Shortcomings of the isolated abyssal
684 reservoir model for deglacial radiocarbon changes in the mid-depth Indo-Pacific Ocean,
685 *Geophys. Res. Lett.*, 38, L04604, doi:10.1029/2010GL046158.
686
- 687 Harada, N., K. H. Shin, A. Murata, M. Uchida, and T. Nakatani (2003), Characteristics of
688 alkenones synthesized by a bloom of *emiliana huxleyi* in the Bering Sea, *Geochim.*
689 *Cosmochim. Acta*, 67, 1507-1519, doi:10.1016/S0016-7037(02)01318-2.
690
- 691 Hendy, I. L., and J. P. Kennett (2000), Dansgaard-Oeschger Cycles and the California Current
692 System: Planktonic foraminiferal response to rapid climate change in Santa Barbara
693 Basin, Ocean Drilling Program Hole 893A, *Paleoceanography*, 15, 30-42,
694 doi:10.1029/1999PA000413.
695
- 696 Hendy, I.L., and J.P. Kennett (2003), Tropical forcing of North Pacific intermediate water
697 distribution during Late Quaternary rapid climate change?, *Quat. Sci. Rev.*, 22, 673-689,
698 doi:10.1016/S0277-3791(02)00186-5.
699
- 700 Hendy, I. L., and T. F. Pedersen (2005), Is pore water oxygen content decoupled from
701 productivity on the California Margin? Trace element results from Ocean Drilling
702 Program Hole 1017E, San Lucia slope, California, *Paleoceanography*, 20, PA4026,
703 doi:10.1029/2004PA001123.
704
- 705 Hendy, I. L., and T. F. Pedersen (2006), Oxygen minimum zone expansion in the eastern tropical
706 North Pacific during deglaciation, *Geophys. Res. Lett.*, 33, L20602,
707 doi:10.1029/2006GL025975.
708
- 709 Hendy, I. L., J. P. Kennett, E. B. Roark, and B. L. Ingram (2002), Apparent synchronicity of
710 submillennial scale climate events between Greenland and Santa Barbara Basin,
711 California from 30–10 ka, *Quat.Sci. Rev.*, 21, 1167-1184, doi:10.1016/S0277-
712 3791(01)00138-X.
713
- 714 Hendy, I. L., T. F. Pedersen, J. P. Kennett, and R. Tada (2004), Intermittent existence of a
715 southern Californian upwelling cell during submillennial climate change of the last 60
716 kyr, *Paleoceanography*, 19, PA3007, doi:10.1029/2003PA000965.
717
- 718 Hughen, K., S. Lehman, J. Southon, J. Overpeck, O. Marchal, C. Herring and J. Turnbull (2004),
719 ^{14}C activity and global carbon cycle changes over the past 50,000 years, *Science*,
720 303(5655), 202-207, doi:10.1126/science.1090300.
721
- 722 Ikehara, K., K. Ohkushi, A. Shibahara, and M. Hoshiba (2006), Change of bottom water
723 conditions at intermediate depths of the Oyashio region, NW Pacific over the past
724 20,000 yrs, *Global Planet. Change*, 53, 78-91, doi:10.1016/j.gloplacha.2006.01.011.
725

- Ishizaki, Y., K. Ohkushi, T. Ito and H. Kawahata (2009), Abrupt changes of intermediate-water oxygen in the northwestern Pacific during the last 27 kyr, *Geo-Marine Letters*, 29, 125-131, doi:10.1007/s00367-008-0128-0.
- Itaki, T., M. Uchida, S. Kim, H.-S. Shin, R. Tada, and B.-K. Khim (2009), Late Pleistocene stratigraphy and palaeoceanographic implications in northern Bering Sea slope sediments: evidence from the radiolarian species *Cycladophora davisiana*, *J. Quat. Sci.*, 24, 856-865, doi:10.1002/jqs.1356.
- Jaccard, S. L., G. H. Haug, D. M. Sigman, T. F. Pedersen, H. R. Thierstein and U. Rohl (2005), Glacial/interglacial changes in subarctic North Pacific stratification, *Science*, 308, 1003-1006, doi:10.1126/science.1108696.
- Kao, S. J., K. K. Liu, S. C. Hsu, Y. P. Chang, and M. H. Dai (2008), North Pacific-wide spreading of isotopically heavy nitrogen during the last deglaciation: Evidence from the western Pacific, *Biogeosciences*, 5, 1641-1650, doi:10.5194/bg-5-1641-2008.
- Keigwin, L. D. (1998), Glacial-age hydrography of the far northwest Pacific Ocean, *Paleoceanography*, 13, 323-339, doi:10.1029/98PA00874.
- Keigwin, L. D., and S. J. Lehman (1994), Deep circulation change linked to HEINRICH Event 1 and Younger Dryas in a middepth North Atlantic Core, *Paleoceanography*, 9, 185-194, doi:10.1029/94PA00032.
- Kennett, J.P., and B.L. Ingram (1995), A 20,000-year record of ocean circulation and climate change from the Santa Barbara Basin, *Nature*, 377, 510-514, doi:10.1038/377510a0.
- Key, R. M., A. Kozyr, C. L. Sabine, K. Lee, R. Wanninkhof, J. Bullister, R. A. Feely, F. Millero, C. Mordy and T.-H. Peng (2004), A global ocean carbon climatology: Results from GLODAP. *Global Biogeochem. Cycles*, 18, GB4031.
- Kiefer, T., M. Sarnthein, H. Erlenkeuser, P. M. Grootes, and A. P. Roberts (2001), North Pacific response to millennial-scale changes in ocean circulation over the last 60 kyr, *Paleoceanography*, 16(2), 179-189, doi:10.1029/2000PA000545.
- Kiefer, T. and M. Kienast (2005), Patterns of deglacial warming in the Pacific Ocean: A review with emphasis on the time interval of Heinrich event 1, *Quat. Sci. Rev.*, 24, 1063-1081, doi:10.1016/j.quascirev.2004.02.021.
- Kienast, S. S., S. E. Calvert, and T. F. Pedersen (2002), Nitrogen isotope and productivity variations along the northeast Pacific margin over the last 120 kyr: Surface and subsurface paleoceanography, *Paleoceanography*, 17, 1055, doi:10.1029/2001PA000650.
- Kim, S., B. K. Khim, M. Uchida, T. Itaki and R. Tada (2011), Millennial-scale paleoceanographic events and implication for the intermediate-water ventilation in the

771 northern slope area of the Bering Sea during the last 71 kyrs, *Global Planet. Change*, 79,
772 89-98, doi:10.1016/j.gloplacha.2011.08.004.
773

774 Kovanen, D. J., and D.J. Easterbrook (2002), Paleodeviations of radiocarbon marine reservoir
775 values for the northeast Pacific, *Geology*, 30, 243-246,
776 doi:10.1130/0091-7613(2002)030<0243:PORMRV>2.0.CO;2

777 Kuzmin, Y. V., G. S. Burr, and A. J. T. Jull (2001), Radiocarbon reservoir correction ages in the
778 Peter the Great Gulf, Sea of Japan, and eastern coast of the Kunashir, Southern Kuriles
779 (Northwestern Pacific), *Radiocarbon*, 43, 477-481.

780 Kuzmin, Y. V., L. A. Nevesskaya, S. K. Krivonogov and G. S. Burr (2007), Apparent ^{14}C ages
781 of the 'pre-bomb' shells and correction values (R, DR) for Caspian and Aral Seas (Central
782 Asia), *Nuclear Instruments and Methods in Physics Research B*, 259, 463-466,
783 doi:10.1016/j.nimb.2007.01.308

784 Lehman, M. F., D. M. Sigman, D. C. McCorkle, B. G. Brunelle, S. Hoffmann, M. Kienast, G.
785 Cane, and J. Clement (2005), Origin of the deep Bering Sea nitrate deficit: Constraints
786 from the nitrogen and oxygen isotopic composition of water column nitrate and benthic
787 nitrate fluxes, *Global Biogeochem. Cycles*, 19, GB4005, doi:10.1029/2005GB002508.
788

789 Lisiecki, L. E., and M. E. Raymo (2005), A Pliocene-Pleistocene stack of 57 globally distributed
790 benthic $\delta^{18}\text{O}$ records, *Paleoceanography*, 20, PA1003, doi:10.1029/2004PA001071.
791

792 Lund, D. C., Mix, A. C., and J. Southon (2011), Increased ventilation age of the deep northeast
793 Pacific Ocean during the last deglaciation, *Nature Geoscience*, 4, 771-774,
794 doi:10.1038/ngeo1272.
795

796 Marchitto, T. M., S. J. Lehman, J. D. Ortiz, J. Flückiger, and A. van Geen (2007), Marine
797 radiocarbon evidence for the mechanism of deglacial atmospheric CO_2 rise, *Science*,
798 316(5830), 1456-1459, doi:10.1126/science.1138679.
799

800 Matsumoto, K. (2007), Radiocarbon-based circulation age of the world oceans, *J. Geophys. Res.*,
801 112, C09004, doi:10.1029/2007JC004095.
802

803 Matsumoto, K., T. Oba, J. Lynch-Stieglitz, and H. Yamamoto (2002), Interior hydrography and
804 circulation of the glacial Pacific Ocean, *Quaternary Sci. Rev.*, 21, 1693-1704,
805 doi:10.1016/S0277-3791(01)00142-1.
806

807 McManus1, J. F., D. W. Oppo1 and J. L. Cullen (1999), A 0.5-million-year record of millennial-
808 scale climate variability in the North Atlantic, *Science*, 283, 971-975,
809 doi:10.1126/science.283.5404.971.
810

811 Mcmanus, J. F., R. Francois, J. M. Gherardi, L. D. Keigwin, and S. Brown-Leger (2004),
812 Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial
813 climate changes, *Nature*, 428, 834-837.

- Mikolajewicz, U., T. J. Crowley, A. Schiller, and R. Voss (1997), Modelling teleconnections between the North Atlantic and North Pacific during the Younger Dryas, *Nature*, 387, 384-387, doi:10.1038/387384a0.
- Mix, A. C., D. C. Lund, N. G. Pisias, P. Bodén, L. Bornmalm, M. Lyle, and J. Pike (1999), Rapid climate oscillations in the Northeast Pacific during the last deglaciation reflect Northern and Southern Hemisphere sources, in *Mechanisms of Global Climate Change at Millennial Time Scales*, *Geophys. Monogr. Ser.*, vol. 112, edited by U. Clark, S. Webb, and D. Keigwin, pp. 127–148, doi:10.1029/GM112p0127, AGU, Washington, D. C.
- Müller, P. J., G. Kirst, G. Ruhland, I. von Storch, and A. Rosell-Melé (1998), Calibration of the alkenone paleotemperature index U_{37}^{K-} based on core-tops from the eastern South Atlantic and the global ocean (60°N-60°S), *Geochimica et Cosmochimica Acta*, 62, 1757-1772, doi:10.1016/S0016-7037(98)00097-0.
- Nimmergut, A. and A. Abelmann (2002), Spatial and seasonal changes of radiolarian standing stocks in the Sea of Okhotsk, *Deep-Sea Res. I*, 49, 463-4493, doi:10.1016/S0967-0637(01)00074-7.
- North Greenland Ice Core Project members (2004), North Greenland Ice Core Project Oxygen Isotope Data, IGBP PAGES/World Data Center for Paleoclimatology, Data Contribution Series # 2004-059, NOAA/NGDC Paleoclimatology Program, Boulder CO, USA.
- Okazaki, Y., K. Takahashi, T. Nakatsuka, and M. C. Honda (2003), The production scheme of *Cycladophora davisiana* (Radiolaria) in the Okhotsk Sea and the northwestern North Pacific: implication for the paleoceanographic conditions during the glacials in the high latitude oceans, *Geophys. Res. Lett.*, 30(18), 1939, doi:10.1029/2003GL018070.
- Okazaki, Y., A. Timmermann, L. Menviel, N. Harada, A. Abe-Ouchi, M. O. Chikamoto, A. Mouchet and H. Asahi (2010), Deepwater formation in the North Pacific during the last glacial termination, *Science*, 329, 200-204, doi:10.1126/science.1190612.
- Reimer, P. J., M. G. L. Baillie, E. Bard, A. Bayliss, J. W. Beck, P. G. Blackwell, C. Bronk Ramsey, C. E. Buck, G. S. Burr, R. L. Edwards, M. Friedrich, P. M. Grootes, T. P. Guilderson, I. Hajdas, T. J. Heaton, A. G. Hogg, K. A. Hughen, K. F. Kaiser, B. Kromer, F. G. McCormac, S. W. Manning, R. W. Reimer, D. A. Richards, J. R. Southon, S. Talamo, C. S. M. Turney, J. van der Plicht, C. E. Weyhenmeyer (2009), IntCal09 and Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP, *Radiocarbon*, 51, 1111-50.
- Robinson, L. F., J. F. Adkins, L. D. Keigwin, J. Southon, D. P. Fernandez, S-L Wang, and D. S. Scheirer (2005) Radiocarbon Variability in the Western North Atlantic During the Last Deglaciation, *Science*, 310, 1469-1473, doi:10.1126/science.1114832.

- Robinson, S.W., and G. Thompson (1981), Radiocarbon corrections for marine shell dates with application to southern Pacific Northwest Coast prehistory, *Syesis*, 14, 45-57.
- Roden, G. (1995) Aleutian Basin of the Bering Sea: Thermohaline, oxygen, nutrient, and current structure in July 1993, *J. Geophys. Res.*, 100(C7), 13539–13554, doi:10.1029/95JC01291.
- Sagawa, T., and K. Ikehara (2008), Intermediate water ventilation change in the subarctic northwest Pacific during the last deglaciation, *Geophys. Res. Lett.*, 35, L24702, doi:10.1029/2008GL035133.
- Sarnthein M., T. Kiefer, P. M. Grootes, H. Elderfield, and H. Erlenkeuser (2006), Warmings in the far northwestern Pacific promoted pre-Clovis immigration to America during Heinrich event 1, *Geology*, 34, 141–144, doi: 10.1130/G22200.1.
- Seki, O., R. Ishiwatari, and K. Matsumoto (2002), Millennial climate oscillations in NE Pacific surface waters over the last 82 kyr: New evidence from alkenones, *Geophys. Res. Lett.*, 29, 2144, doi:10.1029/2002GL015200.
- Shin, K.-H., N. Tanaka, N. Harada, and J.-C. Marty (2002), Production and turnover rates of C37 alkenones in the eastern Bering Sea: implication for the mechanism of a long duration of *Emiliania huxleyi* bloom, *Prog. Oceanogr.*, 55, 113-129, doi:10.1016/S0079-6611(02)00073-3.
- Sigman, D. M., M. A. Altabet, R. Francois, D. C. McCorkle, and J.-F. Gaillard (1999), The isotopic composition of diatom-bound nitrogen in Southern Ocean sediments, *Paleoceanography*, 14(2), 118–134, doi:10.1029/1998PA900018.
- Sigman D. M., K. L. Karsh, and K. L. Casciotti (2009), Nitrogen isotopes in the ocean, in *Encyclopedia of Ocean Sciences (Second Edition)*, edited by J.H. Steele, K. K. Turekian, and S. A. Thorpe, pp. 40-54, Academic Press, Oxford, doi:10.1016/B978-012374473-9.00632-9.
- Sikes, E. L., C. R. Samson, T. P. Guilderson, and W. R. Howard (2000), Old radiocarbon ages in the southwest Pacific Ocean during the last glacial period and deglaciation, *Nature*, 405, 555-559, doi:10.1038/35014581.
- Stabeno, P. J., J. D. Schumacher, and K. Ohtani (1999), The physical oceanography of the Bering Sea, in *Dynamics of the Bering Sea*, edited by T. R. Loughlin and K. Ohtani, pp. 1-28, University of Alaska Sea Grant, Fairbanks, AK.
- Stott, L. D., W. Berelson, R. Douglas and D. Gorsline (2000), Increased dissolved oxygen in Pacific intermediate waters due to lower rates of carbon oxidation in sediments, *Nature*, 407, 367-370, doi:10.1038/35030084.

- Stott, L., J. Southon, A. Timmermann, and A. Koutavas (2009), Radiocarbon age anomaly at intermediate water depth in the Pacific Ocean during the last deglaciation, *Paleoceanography*, 24, PA2223, doi:10.1029/2008PA001690.
- Stuiver, M. and H. A. Polach (1977), Discussion: Reporting of ^{14}C data, *Radiocarbon*, 19, 355-363.
- Stuiver, M., P. J. Reimer, and R. W. Reimer (2005) CALIB 6.0.1
- Talley, L. D. (1993), Distribution and formation of North Pacific Intermediate Water, *J. Phys. Oceanogr.*, 23, 517-538, doi:10.1175/1520-0485(1993)023<0517:DAFONP>2.0.CO;2.
- Talley, L. D., Y. Nagata, M. Fujimura, T. Kono, D. Inagake, M. Hirai, and K. Okuda (1995) North Pacific intermediate water in the Kuroshio/Oyashio mixed water region, *J. Phys. Oceanogr.*, 25, 475-501, doi:10.1175/1520-0485(1995)025<0475:NPIWIT>2.0.CO;2.
- Takahashi, K. N. Fujitani, M. Yanada, and Y. Maita (2000), Long-term biogenic particle fluxes in the Bering Sea and the central subarctic Pacific Ocean, 1990-1995, *Deep-Sea Res. I*, 47, 1723-1759, doi:10.1016/S0967-0637(00)00002-9.
- Takahashi, K., A.C. Ravelo, C. Alvarez Zarikian, and the Expedition 323 Scientists (2011), *Proc. IODP*, 323, Tokyo (Integrated Ocean Drilling Program Management International, Inc.), doi:10.2204/iodp.proc.323.2011.
- Tanaka, S., and K. Takahashi (2005), Late Quaternary paleoceanographic changes in the Bering Sea and the western subarctic Pacific based on radiolarian assemblage, *Deep-Sea Res. II*, 52, 2131-2149, doi:10.1016/j.dsr2.2005.07.002.
- Thornalley, D. J. R., S. Barker, W. S. Broecker, H. Elderfield and I. N. McCave (2011) The deglacial evolution of North Atlantic deep convection, *Science*, 331, 202-205, doi:10.1126/science.1196812.
- Walsh, J. J., C. P. McRoy, L. K. Coachman, J. J. Goering, J. J. Nihoul, T. E. Whitledge, T. H. Blackburn, P. L. Parker, C. D. Wirick, P. G. Shuert, J. M. Grebmeier, A. M. Springer, R. D. Tripp, D. Hansell, S. Djenidi, E. Deleersnijder, K. Henriksen, B. A. Lund, P. Andersen, F. E. Müller-Karger and K. Dean (1989), Carbon and nitrogen cycling with the Bering/Chukchi Seas: Sources regions for organic matter affecting AOU demands of the Arctic Ocean, *Prog. Oceanogr.*, 22, 277-359, doi:10.1016/0079-6611(89)90006-2.
- Wang, R. J., and R.H. Chen (2005), *Cycladophora davisiana* (radiolarian) in the Bering Sea during the late Quaternary: a stratigraphic tool and proxy of the glacial Subarctic Pacific Intermediate Water, *Science in China Series D-Earth Sciences*, 48, 1698-1707, doi:10.1360/02yd0009.

949 Warner, M. J., and G. I. Roden (1995), Chlorofluorocarbon evidence for recent ventilation of the
950 deep Bering Sea, *Nature*, 373, 409–412, doi:10.1038/373409a0..
951
952 Zheng, Y., A. van Geen, R. F. Anderson, J. V. Gardner, and W. E. Dean (2000), Intensification
953 of the Northeast Pacific oxygen minimum zone during the Bölling-Alleröd Warm Period,
954 *Paleoceanography*, 15, 528-536, doi:10.1029
955

Figure captions

Figure 1. Bering Sea bathymetry, surface circulation (red arrows), subsurface circulation (yellow arrows), and the location of Site U1340 [*Expedition 323 Scientists*, 2011].

Figure 2. Vertical oxygen profile for the Bering Sea [*Expedition 323 Scientists*, 2011], with U1340 located in the present day OMZ.

Figure 3. Image of U1340A cores from 0-3.5 m [*Takahashi et al.*, 2011], with increased contrast. Reddish brown sediments are diatom-rich intervals, many of which are laminated.

Figure 4. U1340 sediment density record shows millennial-scale events similar to the Dansgaard–Oeschger events recorded in Greenland ice cores. Yellow bars mark laminated sediments.

Figure 5. Comparison of U1340 benthic $\delta^{18}\text{O}$ to the LR04 [*Lisiecki and Raymo*, 2005] global benthic stack. In addition to the radiocarbon dates in the upper section of core, the U1340 age model incorporates a tie-point to the LR04 stack at 62 ka.

Figure 6. U1340 age with depth (diamonds), and sedimentation rates (squares). Yellow bars mark laminated sediments.

Figure 7. U1340 ventilation age compared with core photos, planktonic $\delta^{18}\text{O}$, bulk $\delta^{15}\text{N}$, and sediment density. Modern day ventilation age indicated by a yellow diamond. Yellow bars mark laminated intervals.

Figure 8. All U1340 isotope records generated for this study, with GRA bulk density [*Expedition 323 Scientists*, 2011]. Yellow bars mark laminated sediments.

Figure 9. U1340 $\delta^{18}\text{O}$, GRA density, bulk $\delta^{15}\text{N}$, and alkenone records from the LGM to the early Holocene. Our radiocarbon age model erroneously places the BA several hundred years earlier than in the N. Atlantic, which is attributed to large fluctuations in reservoir age during the deglaciation. Yellow bars mark laminated sediments.

Figure 10. U1340 records with arrows marking the large isotopic shifts at the beginning of MIS 3 and during the deglaciation.

Figure 11. U1340 $\delta^{18}\text{O}$ compared to records from the N. Pacific and N. Atlantic, with arrows marking the large $\delta^{18}\text{O}$ changes at U1340.

Figure 12. U1340 $\delta^{15}\text{N}$ compared to other N. Pacific records, with an arrow marking the large $\delta^{15}\text{N}$ change at U1340.

Figure 13. U1340 $\delta^{15}\text{N}$ compared to $\delta^{15}\text{N}$ at JPC17, a deeper Bowers Ridge site.

1001 **Figure 14.** Changes in temperature, planktonic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, bulk $\delta^{15}\text{N}$, and GRA density at
1002 U1340 over two separate DO-type intervals. Yellow bars mark laminated sediments.
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Tables

Table 1. U1340 alternate affine table

Site	Hole	Core	Depth offset (m)
U1340	A	1	-0.0605
U1340	A	2	1.1318
U1340	A	3	1.6634
U1340	A	4	1.9757
U1340	D	1	0.2622
U1340	D	2	2.7392

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Table 2. U1340 alternate splice table

Sample ID	Interval depth (cm)	Mbsf	Mcd	TIE	Sample ID	Interval depth (cm)	Mbsf	Mcd
U1340A-1H-03	38.0	3.380	3.320	–	U1340D-1H-03	5.7	3.057	3.320
U1340D-1H-05	84.0	6.840	7.102	–	U1340A-2H-02	57.0	5.970	7.102
U1340A-2H-07	36.4	13.264	14.395	–	U1340D-2H-03	135.6	11.656	14.395
U1340D-2H-04	71.3	12.513	15.253	–	U1340A-3H-01	18.9	13.589	15.253

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Table 3. U1340 low density intervals numbered by ascending age, with any associated laminations (shaded rows)

#	Top depth (mcd)	Bottom depth (mcd)	Thickness (cm)	Top age (cal yrs BP)	Bottom age (cal yrs BP)	Duration (yrs)
1	0.8665	1.0845*	22	11935	12864*	929
1a	0.8354	1.0845	25	11852	12865	1014
2	1.2665	1.8505	58	13785	15355	1570
2a	1.2145	1.3745	16	13629	14113	484
2b	1.4845	1.8345	35	14446	15327	881
3	6.5768	6.7768	20	24028	24417	390
4	7.8038	7.9038	10	26418	26613	195
4a	7.7968	7.8968	10	26402	26597	195
5	9.0458	9.3208	28	28837	29756	919
6	9.7378	9.8628	13	31641	32206	565
6a	9.7568	9.8368	8	31724	32086	362
7	10.8398	10.9648	13	36623	37188	565
7a	10.9018	10.9318	3	36901	37036	136
8	11.3708	11.4458	7	39023	39362	339
9	11.9468	12.0718	13	41627	42192	565
10	12.5538	12.7798	23	44371	45393	1022
11	13.3298	13.5558	23	47879	48901	1022
12	14.3458	14.4208	7	52472	52811	339
13	16.0654	16.2404	18	60246	61037	791
13a	16.1334	16.2134	8	60553	60915	362
14	17.6494	17.7244	7	67466	67466	0
15	18.5424	18.8684	18	71503	72976	814
15a	18.6434	18.8234	18	71900	72714	814
16	19.2184	19.4444	18	74559	75580	814
17	19.5774	19.7774	18	76181	77086	814

*gap in density data, bottom depth assumed to be equivalent to the bottom depth of the associated lamination

Table 4. Age model with radiocarbon dates, sedimentation rates, and ventilation; calculated using $\Delta R = 375 \pm 300$

Foraminifera species used	Depth (mcd)	^{14}C yrs BP $\pm \sigma$	Calendar yrs BP	1 σ range	Ventilation age (Δ_{b-p})	Sed. rate (cm/kyr)
MPL ¹	0.000	5700 \pm 35	5800	5490 6160		
MPL <i>Brizalina aff alata</i>	0.807	10890 ² \pm 57 13310 ² \pm 108	11740 14730	11230 12180 14050 15170	2420 \pm 441	14
MPL <i>Uvigerina peregrina</i>	1.135	11980 \pm 35 12360 \pm 35	13070 13450	12750 13330 13160 13760	370 \pm 427	25
MPL <i>Uvigerina peregrina</i>	1.185	12440 \pm 35 12790 \pm 35	13540 13960	13230 13830 13430 14460	350 \pm 427	11
MPL <i>Brizalina aff alata</i> , <i>Bolivina earlandi</i> , <i>Bulimina exilis</i>	3.385	13440 \pm 40 14320 \pm 60	14960 16470	14180 15470 15960 16970	890 \pm 431	33
MPL <i>Uvigerina peregrina</i>	1.885	13640 \pm 40 14710 \pm 40	15430 17060	14880 16290 16760 17460	1070 \pm 429	49
MPL <i>Uvigerina peregrina</i>	3.385	15420 ² \pm 64 24610 ² \pm 163	17810 28670	17410 18470 28280 29450	9190 \pm 459	63
MPL	9.172	25090 \pm 120	29080	28790 29280		51
N/A (LR04 tie-point)	16.453	N/A	62000	N/A		22

¹mixed planktonic species

²average of two measurements

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Table 5. Key diatom species percentages through the transition from bioturbated deglaciation sediments to laminated Bølling-Allerød sediments.

Sediment type	Depth (mcd)	<i>Chaetoceros</i> sp. (high productivity indicator)	<i>Actinocyclus curvatulus</i> (associated with post-bloom conditions)	<i>Neodenticula seminae</i> (Alaskan Stream indicator)	<i>Thalassiosira antarctica RS</i> (sea ice-related species)	<i>Thalassiosira trifulta</i> group	Fragmented valves % of total	total no. of valves counted
Laminated	1.5445	17.41	0	55.72	0	16.42	16.92	201
	1.6045	36.22	1.02	25.51	2.55	24.49	28.57	196
	1.6545	3.49	1.31	77.73	0	12.66	13.97	229
	1.7045	56.28	0	21.11	0.5	4.52	6.03	199
	1.8045	25.37	1.49	38.81	0	5.47	12.44	201
Transitional	1.8445	58.21	1	9.95	0.5	11.44	12.94	201
Bioturbated	1.9495	8.94	11.38	9.76	17.89	26.02	59.35	123
	2.1495	1.45	30.43	11.59	27.54	20.29	81.16	69

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