1 **Surface nitrate utilization in the Bering Sea since 180 ka BP: Insight**

2 **from sedimentary nitrogen isotopes**

3

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28

29   **Abstract**

30   We present high-resolution records of sedimentary nitrogen ($\delta^{15}N_{bulk}$) and carbon

31   isotope ratios ($\delta^{13}C_{bulk}$) from piston core SO201-2-85KL located in the western Bering

32   Sea. The records reflect changes in surface nitrate utilization and terrestrial organic

33   matter contribution in submillennial resolution that span the last 180 kyr. The $\delta^{15}N_{bulk}$

34   record is characterized by a minimum during the penultimate interglacial indicating

35   low nitrate utilization (~62-80%) despite the relatively high export production inferred

36   from opal concentrations along with a significant reduction in the terrestrial organic

37   matter fraction (m_{terr}). This suggests that the consumption of the nitrate pool at our

38   site was incomplete and even more reduced than today (~84%). $\delta^{15}N_{bulk}$ increases

39   from Marine Isotope Stage (MIS) 5.4 and culminates during the Last Glacial

40   Maximum, which indicates that nitrate utilization in the Bering Sea was raised during

41   cold intervals (MIS 5.4, 5.2, 4) and almost complete during MIS 3 and 2 (~93-100%).

42   This is in agreement with previous hypotheses suggesting that stronger glacial

43   stratification reduced the nutrient supply from the subeuphotic zone, thereby

44   increasing the iron-to-nutrient ratio and therefore the nitrate utilization in the mixed

45   surface layer. Large variations in $\delta^{15}N_{bulk}$ were also recorded from 180 to 130 ka BP

46   (MIS 6) indicating a potential link to insolation and sea-level forcing and its related

47   feedbacks. Millennial-scale oscillations were observed in $\delta^{15}N_{bulk}$ and $\delta^{13}C_{bulk}$ that

48   might be related to Greenland interstadials.

49

50   **1. Introduction**

51   The polar oceans are thought to have been more stratified during past glacial periods

52   and the breakdown of stratification in the Southern Ocean during interglacials has

53   been suggested as a potential control mechanism for the glacial-interglacial cycles in

54   atmospheric carbon dioxide (CO_2) (for a review see Sigman et al., 2010). Recent

55   studies have found supporting evidence that past variations in stratification/ventilation

56   also occurred in the subarctic North Pacific with implications for ocean-atmosphere

57   gas exchange (Jaccard et al., 2005, 2010; Brunelle et al., 2007, 2010; Galbraith et

58   al., 2008a; Okazaki et al., 2010, 2012; Chikamoto et al., 2012; Menviel et al., 2012;

59   Rella et al., 2012; Jaccard and Galbraith, 2013; Max et al., 2014).

60   The modern subarctic North Pacific is characterized by a permanent halocline due to
61   a low-salinity surface layer that limits the exchange of nutrients between the surface
62   and subsurface and prevents the formation of deep water masses (e.g., Warren,
63   1983; Haug et al., 1999; Emile-Geay et al., 2003). At the same time high marine
64   productivity makes this area a net sink for atmospheric CO_2 (Honda et al., 2002;
65   Takahashi et al., 2002b). However, the efficiency of the biological pump in the high-
66   nutrient, low-chlorophyll (HNLC) regions of the subarctic North Pacific is reduced due
67   to iron limitation (e.g., Tsuda et al., 2003), which results in incomplete nitrate
68   utilization.

69   Sedimentary records from the North Pacific and its marginal seas consistently show
70   reduced contents of biogenic opal and barium, $CaCO_3$, and organic carbon during
71   past glacial periods, indicating reduced biological export production (e.g.,
72   Gorbarenko et al., 2002; Narita et al., 2002; Kienast et al., 2004; Nürnberg and
73   Tiedemann, 2004; Jaccard et al., 2005, 2010; Okazaki et al., 2005a; Shigemitsu et
74   al., 2007, 2008; Riethdorf et al., 2013a). Restricted marine productivity both in the
75   North Pacific and in the Antarctic sector of the Southern Ocean is attributed to (i) light
76   limitation due to extensive sea-ice coverage (Elderfield and Rickaby, 2000), or (ii) to
77   enhanced stratification of the upper water column that suppressed nutrient supply to
78   the euphotic zone (Francois et al., 1997).

79   Studies investigating these hypotheses applied stable nitrogen isotope ratios ($\delta^{15}N$)
80   to provide a link to the marine nutrient cycle, but they have been mostly focused on
81   regions not influenced by seasonal sea-ice, which has the potential to modulate
82   biological and terrigenous fluxes. In the NW Pacific the $\delta^{15}N$ signal can be used as a
83   proxy of surface nitrate utilization, whereas in the NE Pacific, it reflects variations in
84   the composition of the subsurface nitrate pool (Brunelle et al., 2007; Galbraith et al.,
85   2008a). The available reconstructions of surface nitrate utilization in the Okhotsk and
86   Bering seas indicate that in both marginal seas enhanced stratification during glacial
87   intervals resulted in a reduced supply of nitrate to the surface and a more complete
88   utilization of surface nitrate because of the continued iron supply from atmospheric
89   deposition (i.e., higher iron-to-nitrate ratio), thereby explaining the low glacial
90   productivity (e.g., Brunelle et al., 2007, 2010; Kim et al., 2011; Khim et al., 2012).

91 For the Bering Sea, recent studies have found indications for millennial-scale
92 oscillations in export production and terrigenous matter supply, which might be
93 connected to changes in stratification and/or sea-ice influence (Gorbarenko et al.,
94 2005, 2010b; Kim et al., 2011; Riethdorf et al., 2013a; Schlung et al., 2013).
95 However, most of the available records of $\delta^{15}N$ are located in the southern Bering
96 Sea and therefore not necessarily influenced by seasonal sea-ice. Moreover, the
97 records are restricted to ~120 ka BP and do not allow for a comparison between the
98 glacial terminations, which due to the large amplitude of climate and environmental
99 changes are considered important periods for the understanding of the carbon cycle
100 (e.g., Yokoyama and Esat, 2011). Especially for Marine Isotope Stage 6 (MIS 6),
101 which in the Okhotsk Sea environment is characterized by rather extreme glacial ice
102 conditions with significantly increased accumulation rates of ice-rafted debris (IRD)
103 (Nürnberg et al., 2011), no records of $\delta^{15}N$ are available.

104 Here, we present isotope geochemical records from a supposedly sea-ice influenced
105 site in the poorly studied western Bering Sea for the last 180 kyr in high-resolution
106 employing sedimentary carbon and nitrogen isotope ratios to reconstruct changes in
107 the contribution of terrestrial organic matter and surface nitrate utilization,
108 respectively. Our results, for the first time, provide information on nitrate utilization in
109 the Bering Sea beyond 120 ka BP and expand the hypothesis of glacial-interglacial
110 stratification changes to cold and warm intervals. Moreover, our record suggests that
111 millennial-scale climate oscillations occurred in the Bering Sea which might be
112 connected to Greenland interstadials.

113

114 **2. Study area**

115 The Bering and Okhotsk seas are marginal seas of the North Pacific, separated from
116 by the Aleutian and Kurile islands, respectively. They are bounded by the coasts of
117 eastern Siberia, the Kamchatkan peninsula and/or western Alaska. Wide and shallow
118 continental shelf areas are found in the northern Okhotsk and in the northern and
119 eastern Bering Sea (Figure 1).

120 With respect to surface circulation, waters from the North Pacific are transported
121 westward along the Aleutian islands by the Alaskan Stream and enter the Bering Sea

122   via the Aleutian passes. There, the Bering Slope Current (BSC) and the East
123   Kamchatka Current (EKC) form boundary currents (Stabeno et al., 1999). Surface
124   outflow is directed into the Arctic through the shallow (~50 m) Bering Strait, whereas
125   surface and deeper waters are transported back into the NW Pacific through the
126   deeper straits, mainly Kamchatka Strait (Figure 1). The EKC and the Oyashio current
127   flow southward and represent western boundary currents of the North Pacific
128   subpolar gyre. The Kurile straits provide entrance and exit pathways to the Okhotsk
129   Sea.

130   Major climatic and oceanographic characteristics of the North Pacific realm are the
131   strong seasonality in sea surface temperatures (SST) and sea-ice formation, the
132   permanent halocline, and a pronounced oxygen minimum zone (OMZ). In the Bering
133   Sea sea-ice is present from September until July reaching its maximum distribution
134   during March/April (Tomczak and Godfrey, 1994; Niebauer et al., 1999). Its formation
135   is related to the interaction of the Siberian High and the Aleutian Low, which results
136   in the advection of cold Arctic air masses, subsequent cooling of the sea surface, and
137   strong winter mixing (Stabeno et al., 1999). Sea-ice is considered as an important
138   transport agent of terrigenous matter in the Okhotsk and Bering seas (Nürnberg and
139   Tiedemann, 2004; Nürnberg et al., 2011; Riethdorf et al., 2013a). Geochemical
140   results indicate that sediments on the eastern Bering Sea shelf and in the Meiji Drift
141   in the NW Pacific are supplied from Yukon–Bering Sea sources (VanLaningham et
142   al., 2009; Asahara et al., 2012; Nagashima et al., 2012). Based on these results
143   Riethdorf et al. (2013a) proposed that terrigenous matter entrained into sea-ice by
144   tidal pumping, suspension freezing, and beach-ice formation, was transported from
145   the eastern Bering Sea shelf to the location studied in this paper, although a
146   contribution by suspension load carried by the BSC could not be excluded.

147   Although sea-ice formation and according brine rejection in the northern Okhotsk Sea
148   drive the modern ventilation of North Pacific Intermediate Water (NPIW) (e.g.,
149   Yasuda, 1997; Yamamoto et al., 2001), the source of NPIW might have shifted to the
150   Bering Sea in the past (Matsumoto et al., 2002; Ohkushi et al., 2003; Tanaka and
151   Takahashi, 2005; Rella et al., 2012), where it nowadays resides at the depth of the
152   26.8 potential density (σ_θ) surface in ~200-400 m (Roden, 1995; Macdonald et al.,
153   2001). The OMZ is found beneath the NPIW with minimum dissolved oxygen

154 concentrations of ~15-20 μmol kg^{-1} at ~900-1100 m (Roden, 2000; Lehmann et al.,
155 2005).

156 In the Bering Sea, high marine productivity is observed, which is mainly associated
157 with shelf areas (e.g., Springer et al., 1996; Stabeno et al., 1999) and dominated by
158 diatoms. Major biological fluxes occur during spring/summer (mainly diatoms) and
159 late summer/early fall (coccolithophores and planktonic foraminifera) (Takahashi et
160 al., 2002a). Nutrients are consumed during the productive seasons and returned from
161 the subsurface by winter mixing. Although winter mixing supplies nutrients from the
162 subsurface into the euphotic zone, near-surface nutrients are not completely
163 consumed by phytoplankton during the productive seasons. Therefore, the western
164 Bering Sea studied here, as well as the central eastern and western parts of the
165 subarctic North Pacific are HNLC regions with perennially high surface nitrate
166 concentrations (e.g., Tyrrell et al., 2005) (Figure 1). As extensively discussed in
167 Brunelle et al. (2007, 2010) changes in the extent of surface nitrate utilization can be
168 reconstructed using records of δ^{15}N, if the underlying assumptions include a constant
169 isotope effect for nitrate assimilation and little or no changes in the δ^{15}N of the source
170 nitrate. There is evidence for the northward propagation of ^{15}N-enriched nitrate from
171 the eastern tropical North Pacific along coastal North America (Liu and Kaplan, 1989;
172 Altabet et al., 1999; Kienast et al., 2002; Sigman et al., 2003). Hence,
173 paleoceanographic interpretations of sedimentary δ^{15}N have to consider changes in
174 the δ^{15}N of the subsurface nitrate pool.

175

176 **3. Material and methods**

177 This study is based on 18.13 m-long piston core SO201-2-85KL (referred to as 85KL
178 hereafter) recovered during R/V Sonne expedition SO201 KALMAR Leg 2 in 2009
179 (Dullo et al., 2009) from Shirshov Ridge, western Bering Sea (57°30.30'N,
180 170°24.77'E, 968 m deep; Figure 1). Sediments from this core mainly consist of
181 terrigenous siliciclastic material bound to the clay- and silt-fractions, but layers of
182 diatomaceous ooze are repeatedly intercalated. Carbonate preservation is poor and
183 at best sporadic, and no sediments younger than 7.5 ka BP were recovered.

184

185   3.1 Bulk sedimentary analyses (TOC, TN, $\delta^{13}C_{bulk}$, $\delta^{15}N_{bulk}$)

186   Total organic carbon (TOC) and total nitrogen (TN) concentrations, as well as

187   sedimentary stable carbon ($\delta^{13}C_{bulk}$) and nitrogen ($\delta^{15}N_{bulk}$) isotope ratios were

188   determined downcore (>9.1 ka BP) every 5 cm from a total of 357 bulk sediment

189   samples. About 25 mg of freeze-dried, hand-ground (agate mortar) sediment was

190   weighed into Ag capsules, acidified with 100 µl hydrochloric acid (3N) to remove

191   inorganic carbon, and dried in a desiccator filled with phosphorus(V)oxide and

192   sodium hydroxide. To ensure complete combustion, the Ag capsules were

193   subsequently wrapped into Sn capsules. TOC, TN, $\delta^{13}C_{bulk}$, and $\delta^{15}N_{bulk}$ were

194   determined at the Center for Advanced Marine Core Research, Kochi University,

195   using a Flash EA 1112 Series elemental analyzer (EA; Thermo Fisher Scientific,

196   USA) coupled with a Delta Plus Advantage isotope ratio mass spectrometer (IRMS;

197   Thermo Fisher Scientific, USA) via a Conflo III interface (He carrier). Stable isotope

198   results are reported in conventional δ-notation and referenced to the Vienna PeeDee

199   Belemnite (VPDB) standard and to atmospheric nitrogen. Molar TN/TOC ratios were

200   corrected for inorganic nitrogen compounds based on a linear regression between

201   TOC and TN following Goñi et al. (1998) (referred to as molar N/C ratios hereafter;

202   Figure 2). Analytical precision (1σ) was determined from two different standards (L-

203   Alanine, n = 58; Sulfanilamide, n = 60) and was <2% RSD (relative standard

204   deviation) for TOC, <5% RSD for TN, ±0.01 mol mol^{-1} for N/C, ±0.14‰ for $\delta^{13}C_{bulk}$,

205   and ±0.29‰ for $\delta^{15}N_{bulk}$. Reproducibility of the samples, determined from replicates

206   (1σ, n = 14), was ±0.01 wt.% for TOC and TN, ±0.01 mol mol^{-1} for N/C, ±0.11‰ for

207    $\delta^{13}C_{bulk}$, and ±0.86‰ for $\delta^{15}N_{bulk}$. This rather high value for $\delta^{15}N_{bulk}$, being significantly

208   higher than instrumental precision, might be related to sample inhomogeneity and

209   potential alteration of the sedimentary organic matter during pre-analysis acid

210   treatment (Brodie et al., 2011a, 2011b).

211

212   3.2 Age model

213   The stratigraphic framework of 85KL is described in detail in Max et al. (2012) and

214   Riethdorf et al. (2013a). Briefly, X-ray fluorescence (XRF) and spectrophotometric

215   (color b*) core logging data were correlated to the $\delta^{18}O$ records of the NGRIP ice

216 core (NGRIP members, 2004; GICC05 timescale, Rasmussen et al., 2006) and the
217 Sanbao stalagmites (Wang et al., 2008). This approach was validated by benthic
218 $\delta^{18}O$ stratigraphy, magnetostratigraphy, AMS ^{14}C dating of planktonic foraminifera,
219 and intercore correlations to neighbouring sediment cores. Linear sedimentation
220 rates vary between 4 and 23 cm kyr^{-1} (average of ~12 cm kyr^{-1}), which translates into
221 a submillennial time-resolution for our reconstructions.

222

223 **3.3 Existing data**

224 For comparison we used logging data and geochemical results reflecting changes in
225 export production and terrigenous matter supply already available for core 85KL.
226 Method details are given elsewhere (Max et al., 2012; Riethdorf et al., 2013a,
227 2013b). In summary, light and color reflectance were measured directly after core
228 recovery every 1 cm using a Minolta CM 508d hand-held spectrophotometer and
229 converted into CIE L*, a*, and b* color space. XRF scanning for elements Al through
230 to Ba was performed at 1 cm sampling resolution using the Avaatech XRF core
231 scanner at Alfred Wegener Institute for Polar and Marine Research, Bremerhaven.
232 Molybdate-blue spectrophotometry was used to determine biogenic opal
233 concentrations (after Müller and Schneider, 1993), and concentrations of $CaCO_3$
234 were calculated from the difference of total carbon (TC) and TOC previously
235 determined using a Carlo Erba CNS analyzer (model NA-1500) at GEOMAR, Kiel.
236 The relative amount of siliciclastics was calculated by subtracting the sum of $CaCO_3$,
237 TOC, and opal concentrations from a total of 100 wt.%. Records of XRF Ca/Ti log-
238 ratios, XRF Br count rates (in counts per second, cps), and color b* correlated with
239 $CaCO_3$ (R^2 = 0.65), TOC (R^2 = 0.64), and opal (R^2 = 0.61), respectively. This finding
240 is in agreement with other studies linking biogenic $CaCO_3$ with normalized XRF
241 records of Ca (Jaccard et al., 2005), TOC with biophilic Br (Ziegler et al., 2008), and
242 opal and organic matter content with color b* (Debret et al., 2006).

243

244 **3.4 Reconstruction of export production**

245   The use of $CaCO_3$, TOC, and opal to reconstruct changes in export production is
246   subject to specific restrictions. $CaCO_3$, especially in the North Pacific, is influenced
247   by carbonate dissolution and might be more indicative of changes in the bottom
248   water calcite saturation state (e.g. Jaccard et al., 2005). With respect to TOC, it is
249   necessary to discriminate between marine and terrestrial carbon sources. The most
250   often used proxy for reconstructions of paleo-export production in the North Pacific
251   realm is biogenic opal (e.g. Kienast et al., 2004).

252   In paleoceanography, fluxes are usually reconstructed using accumulation rates
253   rather than proxy concentrations. However, for the North Pacific and Bering Sea
254   several studies provide evidence for the similar evolution of concentration and
255   accumulation records of biogenic components (e.g. Crusius et al., 2004; Brunelle et
256   al., 2007, 2010). Here, we used dry bulk density measurements (Riethdorf et al.,
257   2013a) to calculate bulk mass accumulation rates (AR Bulk, in g cm^{-2} kyr^{-1}), as well
258   as proxy accumulation rates for $CaCO_3$, TOC, opal, and siliciclastics. The result is
259   shown in Figure 3, clearly demonstrating that sedimentation at Site 85KL is
260   dominated by siliciclastic input. Overall, concentrations of the biogenic components
261   are low, but opal concentrations and opal accumulation rates show a positive linear
262   relationship (R^2 = 0.49). We therefore assume in this paper that at Site 85KL
263   concentrations of opal and color b* logging data are related to export production.

264

265   **4. Results**

266   **4.1 Export production and terrigenous matter supply**

267   In general, concentrations of $CaCO_3$, TOC, and opal, as well as their approximating
268   logging data show increased values during warm intervals (MIS 5.5, 5.3, 5.1, and 1)
269   and Greenland interstadials (GI), but low values during cold intervals (MIS 6, 5.4, 5.2,
270   4, and 2) and Greenland stadials (GS) (Figure 4; Riethdorf et al., 2013a). Because of
271   sedimentary dilution, the proxy records reflecting terrigenous matter supply
272   (%Siliciclastics, XRF data of Al) have an inversed shape with respect to the records
273   reflecting export production. Our EA-IRMS-based TOC results are in excellent
274   agreement with those of Riethdorf et al. (2013a) and in higher temporal resolution
275   extend the respective record by ~30 kyr into MIS 6. The temporal evolution of TOC

276 recorded during MIS 6 strongly corresponds to that observed in color b*, and, to a

277 lesser degree, in XRF Ca/Ti log-ratios and Br count rates (Figure 4). With respect to

278 TOC concentrations, MIS 6 is characterized by a strong variability within the range of

279 ~0.4 to ~1.7 wt.%, showing several short-lived oscillations and highest

280 concentrations during ~156-137 ka BP.

281

282 **4.2 $\delta^{13}C_{bulk}$, $\delta^{15}N_{bulk}$, N/C ratios, and estimation of the terrestrial organic matter**

283 **fraction (m_{terr})**

284 Values for $\delta^{13}C_{bulk}$ (of TOC) and for $\delta^{15}N_{bulk}$ (of TN) ranged from -25.4 to -21.9‰ and

285 from 1.7 to 7.5‰, respectively, and N/C ratios varied between 0.04 and 0.11 mol mol$^-$

286 ^{1} (Table 1; Figures 5 and 6). In general, $\delta^{13}C_{bulk}$ and N/C ratios are more positive

287 during warm intervals (MIS 5.5, 5.3, 5.1, 3, and 1) and show pronounced, but short-

288 lived maxima during GI, which especially for $\delta^{13}C_{bulk}$ exceed analytical precision and

289 reproducibility. The cold intervals (MIS 5.4, 5.2, and 4) and some GS are

290 characterized by decreases in both proxies. A different temporal evolution is

291 recorded for $\delta^{15}N_{bulk}$. The base of core 85KL shows $\delta^{15}N_{bulk}$ values of ~5-6‰, which

292 is followed by a sharp decrease at ~172 ka BP to values of ~2-3‰ (Figure 6).

293 Subsequently, $\delta^{15}N_{bulk}$ again increases until another sudden drop of ~3‰

294 characterizes the transition from MIS 6 to 5.5 (Termination II). During the last

295 interglacial $\delta^{15}N_{bulk}$ remained low at ~3‰ but it decreased even further to minimum

296 values of 1.7‰ at the beginning of cold MIS 5.4. After MIS 5.4, a long-term trend

297 toward higher values that continues into the early Holocene is observed. In between,

298 increasing values were recorded during cold MIS 5.2 and 4, whereas $\delta^{15}N_{bulk}$

299 decreased (MIS 5.1) or remained almost constant (MIS 5.3 and 3) during warm

300 stages. The transition from MIS 2 to 1 (Termination I) is characterized by local

301 minima (~4-5‰) during the cold phases of Heinrich Stadial 1 (HS1; 18.0-14.7 ka BP,

302 Sarnthein et al., 2001) and the Younger Dryas (YD; 12.9-11.7 ka BP, Blockley et al.,

303 2012), and by an intercalated pronounced maximum during the Bølling-Allerød warm

304 phase (B/A; 14.7-12.9 ka BP, Blockley et al., 2012) (up to 6.8‰). Subsequent to the

305 YD, $\delta^{15}N_{bulk}$ continuously increased until highest values of 7.5‰ were recorded

306 during the early Holocene (~10.1 ka BP). Notably, the $\delta^{15}N_{bulk}$ record also features

307 short-lived maxima during GI, which, however, must be considered insignificant with
308 respect to reproducibility.

309 Assuming that the geochemical and isotopic sedimentary composition represents a
310 mixture of marine and terrestrial organic matter we applied a linear mixing model to
311 estimate the fraction of terrestrial-derived organic matter (m_{terr}) using hypothetical
312 endmember compositions. We therefore followed the results of Walinsky et al.
313 (2009). This approach has recently been applied to a sediment core from the
314 northern Gulf of Alaska (Addison et al., 2012). Our Holocene samples lie within the
315 ranges reported by Smith et al. (2002) for surface sediment samples from the
316 southeastern Bering Sea shelf, whereas our glacial samples compare to terrigenous
317 particulate organic matter (POM) from the Yukon River (Guo and Macdonald, 2006)
318 (Figure 5). Accordingly, we assume that the most likely organic matter sources for
319 our site are marine phytoplankton (δ^{13}C: -22 to -20‰; δ^{15}N: >5‰; N/C: 0.10 to 0.15
320 mol mol^{-1}), soil (δ^{13}C: -26.5 to -25.5‰; δ^{15}N: 0 to 1‰; N/C: 0.08 to 0.10 mol mol^{-1}),
321 and vascular plant detritus (VPD; δ^{13}C: -27 to -25‰; δ^{15}N: 0 to 1‰; N/C: 0 to 0.05
322 mol mol^{-1}) (Meyers, 1994; McQuoid et al., 2001; Geider and La Roche, 2002; Smith
323 et al., 2002; Guo et al., 2004; Gaye-Haake et al., 2005; Guo and Macdonald, 2006;
324 Walsh et al., 2008, and references therein) (Figure 5; Table 2). With the influence of
325 soil considered insignificant, m_{terr} was calculated as:

326
$$m_{terr} = (A_{sample} - A_{mar}) / (A_{terr} - A_{mar}) \qquad \textit{(Eq. 1)}$$

327 In Equation 1 A refers to the δ^{13}C$_{bulk}$ or N/C ratio of the sample and the respective
328 average values for the assumed marine and terrestrial endmember composition
329 summarized in Table 2. We preferred using molar N/C over C/N ratios, because
330 mixing lines based on C/N are reported to underestimate the fraction of terrestrially
331 derived organic carbon (Perdue and Koprivnjak, 2007). This approach resulted in
332 similar values but in part different temporal evolutions of the respective m_{terr} records,
333 which is attributed to the low variability in N/C ratios. m_{terr} varied between ~10% and
334 ~90% with average values of ~40-50% (Table 1). It was lowest during warm intervals
335 (MIS 5.5, 5.3, 5.1, 3, and 1) and GI, when marine productivity was high. Reductions
336 of up to 50% occurred during the transitions from cold to warm intervals, whereas

337 pronounced but short-lived increases of ~20% seem to correspond to GS (Figure 6).

338 In the discussion we refer to m_{terr} derived from $\delta^{13}C_{bulk}$.

339

340 **5. Discussion**

341 **5.1 The $\delta^{15}N_{bulk}$ signal and potential alteration**

342 $\delta^{15}N_{bulk}$ values reflect the isotopic signature of the export flux of organic matter plus

343 any secondary alteration of this signal during sinking and burial (e.g., Galbraith et al.,

344 2008b; Robinson et al., 2012, and references therein). Hence, interpretation of

345 variations in the $\delta^{15}N_{bulk}$ record must consider changes in (i) the isotopic composition

346 of the subsurface nitrate pool, which is controlled by nitrogen fixation by diazotrophic

347 bacteria and by denitrification, (ii) the degree of nitrification and surface nitrate

348 utilization, and (iii) secondary alteration.

349 Nitrate is the primary nitrogen source for marine phytoplankton, which preferentially

350 incorporates isotopically light (^{14}N-enriched) nitrate (Pennock et al., 1996; Waser et

351 al., 1998). In the Bering Sea the source nitrate is supplied to the surface from below

352 the euphotic zone with a modern value ($\delta^{15}N_{nitrate}$) of ~5.5‰ (Lehmann et al., 2005),

353 which is slightly higher than the global deep ocean average of ~5‰ because of

354 denitrification in the North Pacific (Sigman et al., 2000, Brunelle et al., 2007).

355 Nitrogen fixation results in $\delta^{15}N_{nitrate}$ values that are isotopically light and close to that

356 of air (0‰; Carpenter et al., 1997), and it is the main reason for the low $\delta^{15}N$ values

357 of nitrate and sinking detritus in the tropical/subtropical ocean basins (e.g., Somes et

358 al., 2010). Water column denitrification occurs under low dissolved oxygen

359 concentrations (<5 µmol l^{-1}; Codispoti et al., 2001) and results in a ^{15}N-enriched

360 nitrate pool (Barford et al., 1999). In this respect, $\delta^{15}N$ might also reflect redox

361 conditions in the past with higher values during bottom water suboxia (Galbraith et

362 al., 2004; Kashiyama et al., 2008; Jaccard and Galbraith, 2012; Robinson et al.,

363 2012). Today, water column denitrification is mainly observed in the Arabian Sea, the

364 eastern tropical North Pacific, and in the eastern tropical South Pacific. Thus, the

365 export of ^{15}N-enriched waters might result in a shift toward higher $\delta^{15}N_{nitrate}$, as

366 observed along coastal North America in the subarctic NE Pacific (Liu and Kaplan,

367   1989; Altabet et al., 1999; Kienast et al., 2002; Sigman et al., 2003). However,

368   modern dissolved oxygen concentrations at Site 85KL lie above the denitrification

369   threshold.

370   Results from benthic foraminiferal assemblages from the same site (Ovsepyan et al.,

371   2013) suggest oxidizing conditions in the surface sediment layer from MIS 3 to the

372   Last Glacial Maximum (LGM), but oxygen-depleted conditions during the mid-B/A

373   and early Holocene. This is in agreement with Kim et al. (2011) who for Site PC23A

374   reported on a dominance of oxic benthic foraminiferal species during MIS 2 and 3,

375   but on dominantly dysoxic species during the B/A, early Holocene, and GI. These

376   observations and our proxy records for export production indicate the presence of

377   mostly oxic bottom waters and strongly reduced export of organic matter during most

378   of the past 180 kyr, arguing against a significant impact of water column

379   denitrification on $\delta^{15}N_{bulk}$. Over the last 180 kyr, one can expect a slightly higher $\delta^{15}N$

380   of nitrate during warm stages because of the greater extent of denitrification zone in

381   the North Pacific. The isotopic impact of such increased denitrification during warm

382   stages is thought to be relatively equal to the one observed today (~0.5‰: Lehmann

383   et al., 2005). This shift toward heavier $\delta^{15}N$ is opposite to the expected isotopic effect

384   of decreased nitrate utilization during interglacial scale. Thus, the relatively small

385   increase in $\delta^{15}N_{bulk}$ during warm stages (~0.5‰) should not mask the larger isotopic

386   variation expected from nitrate utilization change (Brunelle et al., 2007 and reference

387   therein). Thus, we are confident that the observed variations in our $\delta^{15}N_{bulk}$ can be

388   used to assess relative changes in the utilization of nitrate in the Bering Sea surface

389   water except for period where dysoxia was present locally. During phases of local

390   enhanced export production and oxygen-depleted bottom water conditions, as

391   recorded during the B/A and the early Holocene, as well as during GI, denitrification

392   might have resulted in a shift toward heavier $\delta^{15}N_{bulk}$ We will thus need to also

393   consider potential changes in the ^{15}N signature of the source nitrate, therefore our

394   nitrate utilization estimate might not be accurate during those periods and a

395   multiproxy approach will be needed to decipher the exact cause of change in $\delta^{15}N_{bulk}$.

396   Nitrate utilization is incomplete in the modern subarctic North Pacific and in the

397   Bering Sea. Accordingly, the isotopic value of the export flux of organic matter

398   ($\delta^{15}N_{export}$) is lighter than that of $\delta^{15}N_{nitrate}$ (Altabet and Francois, 1994; Sigman et al.,

399   1999; Needoba et al., 2003; Galbraith et al., 2008a). The difference between

400    $\delta^{15}N_{nitrate}$ and $\delta^{15}N_{export}$ is primarily controlled by the nitrate utilization and decreases

401   as it becomes more complete and in most of the global ocean the difference is zero

402   due to almost complete utilization (Altabet et al., 1999; Thunell et al., 2004). We can

403   calculate $\delta^{15}N_{export}$ for the expected integrated organic nitrogen export at Site 85KL

404   assuming Rayleigh fractionation kinetics (Altabet and Francois, 1994; Mariotti et al.,

405   1981) after:

406                  $\delta^{15}N_{export} = \delta^{15}N_{nitrate} + f / (1 - f)\ \varepsilon\ \ln(f)$ *(Eq. 2)*

407   In Equation 2 f is the fraction of unutilized nitrate (i.e., $[NO_3^-]_{summer} / [NO_3^-]_{winter}$) and ε

408   is the isotope effect for nitrate incorporation by phytoplankton, which was assumed to

409   be constant at ~5‰ for simplification (Brunelle et al., 2007, 2010). Using a modern

410   average for f of 0.16 (84% utilization), estimated from WOA 2009 surface nitrate

411   concentrations (Garcia et al., 2010), gives $\delta^{15}N_{export}$ = ~3.8‰ for our site, which is

412   1.7‰ lower than the modern $\delta^{15}N_{nitrate}$. Unfortunately, no coretop $\delta^{15}N_{bulk}$ results are

413   available for verification at our site.

414   Alternatively, secondary alteration (preferential loss of organic nitrogen, leakage of

415   ammonium into pore waters, ammonium absorption into clay minerals,

416   winnowing/size fractionation) might have raised the modern coretop $\delta^{15}N$ value. In

417   general, however, alteration of the $\delta^{15}N_{bulk}$ from that of the sinking flux is not

418   considered to have a considerable influence in organic-rich sediments from high-

419   accumulation regions with low contributions of inorganic nitrogen compounds (for a

420   review see Robinson et al., 2012). Core 85KL has rather high average TOC

421   concentrations (~0.9 wt.%), whereas TIN is low at ~0.017 wt.% (Figure 2a), and the

422   relationship between TN and $\delta^{15}N_{bulk}$ is only weak (R^2 = 0.24; Figure 2b), which

423   argues against alteration. Moreover, in comparison with other Bering Sea records

424   using $\delta^{15}N$ as a proxy for nitrate utilization (Brunelle et al., 2007, 2010; Kim et al.,

425   2011; Schlung et al., 2013) our $\delta^{15}N_{bulk}$ record generally shows a similar evolution

426   (Figure 7). Notably, at our study site $\delta^{15}N_{bulk}$ values >5.5‰ were mainly recorded

427   during Termination I, which, in accordance with the previously mentioned studies is

428   more likely attributed to enhanced deglacial water column denitrification.

429   Finally, there might have been influence from terrestrial nitrogen, since our estimates
430   for m_{terr} suggest significant average contributions (~40-50%) of terrestrial organic
431   matter in Shirshov Ridge sediments. In this respect, variations in $\delta^{15}N_{bulk}$ might reflect
432   changes in the supply of terrestrial nitrogen. However, for the estimation of m_{terr} we
433   assumed that the $\delta^{15}N_{bulk}$ of terrestrial organic matter is lower than that of marine
434   organic matter. If there was considerable influence from terrestrial nitrogen we would
435   expect a strong positive relationship between N/C ratios and $\delta^{15}N_{bulk}$, which is not
436   observed (R^2 = 0.10). Accordingly, although Site 85KL is characterized by overall
437   high m_{terr} values with a strong downcore variability, there seems to have been no
438   significant influence of m_{terr} on the $\delta^{15}N_{bulk}$ signal, which might be explained by an
439   only low fraction of terrestrial nitrogen. We therefore consider the influence of
440   secondary alteration and contamination from terrestrial nitrogen on the $\delta^{15}N_{bulk}$ signal
441   as insignificant and in the following discuss variations in $\delta^{15}N_{bulk}$ by means of
442   changes in surface nitrate utilization.

443

444   **5.2 Cold and warm intervals of the past 180 kyr**

445   At Site 85KL, glacial periods, specifically cold intervals (MIS 6, 5.4, 5.2, and 4 to 2),
446   were characterized by significantly reduced export production and enhanced
447   terrigenous matter supply (Riethdorf et al., 2013a; Figure 4), which is in agreement
448   with other studies from the subarctic North Pacific (e.g., Kienast et al., 2004; Jaccard
449   et al., 2005), the Okhotsk Sea (e.g., Narita et al., 2002; Nürnberg and Tiedemann,
450   2004; Okazaki et al., 2005b; Nürnberg et al., 2011), and the Bering Sea (e.g.,
451   Okazaki et al., 2005a; Brunelle et al., 2007; Kim et al., 2011). This is supported by
452   our reconstruction of m_{terr} indicating an average terrestrial organic matter fraction of
453   ~40-50%, which is significantly reduced only during warm intervals. Today, the
454   organic matter in Bering Sea sediments is dominantly of marine origin (Méheust et
455   al., 2013). The glacial terrigenous matter source of Bering Sea sediments is under
456   debate, but there are indications that they originate from source rocks drained by the
457   Yukon River (VanLaningham et al., 2009), and/or from sea-ice rafting in the NE
458   Bering Sea (Riethdorf et al., 2013a).

459 For the Bering and Okhotsk seas the outlined observations were explained by
460 enhanced sea-ice influence and stronger stratification of the upper water column
461 during cold climate conditions restricting marine productivity (Nürnberg and
462 Tiedemann, 2004; Brunelle et al., 2007, 2010; Kim et al., 2011; Khim et al., 2012;
463 Riethdorf et al., 2013a). This restriction results from the extended sea-ice season and
464 coverage and the subsequent limitation of light availability and vertical mixing
465 (nutrient supply), but temperature limitation is likely to have played an additional role.
466 An extended Bering Sea sea-ice coverage during cold phases is supported by
467 reconstructions from diatom assemblages (Katsuki and Takahashi, 2005) and from
468 the diatom-derived, highly branched isoprenoid sea ice biomarker (IP_{25}) (Max et al.,
469 2012).

470 Records of sedimentary and diatom-bound $\delta^{15}N$ imply enhanced surface nitrate
471 utilization as a result of stronger upper water column stratification in the Bering Sea,
472 especially during MIS 3 and 2 (Brunelle et al., 2007, 2010; Kim et al., 2011). Similar
473 observations are reported for the Okhotsk Sea (Brunelle et al., 2010; Khim et al.,
474 2012) and the subarctic NW Pacific (Galbraith et al., 2008a; Brunelle et al., 2010),
475 indicating that these regions were not always HNLC. This shift towards higher $\delta^{15}N$ in
476 MIS 3 is, however, not observed at IODP Site U1340 at the northeastern flank of
477 Bowers Ridge (Schlung et al., 2013). Instead, a sharp decrease in $\delta^{15}N_{bulk}$ was
478 recorded at this site at ~55 ka BP, which might be related to local stratification
479 changes, or to the influence of turbidites that are reported to compromise records
480 recovered from that area (Nakatsuka et al., 1995). Our $\delta^{15}N_{bulk}$ results are in
481 accordance with the former Bering Sea studies confirming enhanced nitrate
482 utilization during cold intervals, but they seem to reveal a more complex development
483 of stratification and for the first time provide information for MIS 6:

484 At the base of core 85KL (~180-173 ka BP), early MIS 6 is characterized by high
485 $\delta^{15}N_{bulk}$ values indicating almost complete nitrate utilization, when export production
486 was reduced, but maintained. Relatively high $\delta^{15}N_{bulk}$ during MIS 6 were also
487 recorded at Okhotsk Sea sites GGC27 (Brunelle et al., 2010) and GC9A (Khim et al.,
488 2012), as well as at ODP Site 882 and at Site MD01-2416 (Galbraith et al., 2008a).
489 Hence, low insolation and weak seasonal contrasts most probably caused a

490   prolonged sea-ice season, extended sea-ice coverage, and suppressed vertical

491   mixing. At ~172 ka BP, when Northern Hemisphere summer insolation had a local

492   maximum, a sharp decrease in $\delta^{15}N_{bulk}$ implies a sudden drop in nitrate utilization

493   (from ~100% to ~50-70%) in the Bering Sea (Figure 6). At the same time export

494   production was increased, whereas m_{terr} decreased. This might be explained by a

495   shortened sea-ice season, reduced sea-ice coverage, and enhanced winter mixing

496   due to stronger seasonal contrasts, which increased the nutrient supply from the

497   subeuphotic zone.

498   Accordingly, we speculate that changes in nitrate utilization are strongly affected by

499   insolation forcing and a feedback by sea-ice processes that drive the extent of

500   vertical mixing during winter, as well as the input of terrestrial organic matter.

501   Dominant climate control via insolation has already been proposed for the Okhotsk

502   Sea (Gorbarenko et al., 2010a, 2012). The long-term increase in nitrate utilization

503   after ~172 ka BP until Termination II, which is also observed at the Okhotsk Sea and

504   NW Pacific sites (Figure 7), as well as the long-term increase from MIS 5.4 until

505   Termination I might be explained by increasingly fostered stratification (i.e. a

506   reduction in the supply of nutrients into the euphotic zone), which is finally subject to

507   a 'breakdown' during the deglaciations. It is beyond the scope of this paper to

508   decipher the underlying causes of this deglacial breakdown, but increasing insolation

509   resulting in a reduced sea-ice season and strengthened winter mixing, is a likely

510   contributing factor.

511   The observation that the strongest maxima in Northern Hemisphere summer

512   insolation, mainly those of warm intervals (MIS 5.5, 5.3, 5.1, and 1), are reflected by

513   maxima in export production (Figure 4), minima in m_{terr}, and decreasing or constant

514   $\delta^{15}N_{bulk}$ (Figure 6), supports the view that insolation changes affect nutrient-limited

515   marine productivity by a feedback in sea-ice processes and winter mixing. It is also in

516   agreement with previously published concepts proposed to explain glacial-interglacial

517   changes in the Okhotsk (Seki et al., 2004; Okazaki et al., 2005b; Khim et al., 2012)

518   and Bering seas (Nakatsuka et al., 1995; Kim et al., 2011). When applying Equation

519   2, our $\delta^{15}N_{bulk}$ record indicates increasing nitrate utilization during cold MIS 5.4 (from

520   ~50 to ~90%), MIS 5.2 (from ~90 to ~100%), and MIS 4 (from ~80 to ~93%), and

521   almost complete utilization during MIS 3 and 2 (~93-100%). On the other hand,

522   decreasing or constant nitrate utilization was recorded during warm MIS 5.5 (from

523   ~97 to ~62%), MIS 5.3 (~90%), and MIS 5.1 (from ~100 to ~80%). This suggests that

524   stratification was fostered during cold intervals, but weakened during warm intervals

525   due to the processes outlined above.

526   The low glacial $\delta^{15}N_{bulk}$ values of ~2-3‰ at the beginning of MIS 5.4, of ~3.5-4.5‰ at

527   the beginning of MIS 4, and the concurrent increases in m_{terr} (Figure 6) might be

528   explained (i) by a higher contribution of (^{15}N-depleted) terrestrial organic matter, or (ii)

529   by stronger vertical mixing. We disregard the first possibility, because we already

530   discarded the potential effect of terrestrial nitrogen on the $\delta^{15}N_{bulk}$ signal (Section

531   5.1). Stronger vertical mixing in the Bering Sea during MIS 5.4 and 4 might be related

532   to the increased formation and/or ventilation of intermediate waters as inferred from

533   neodymium isotope ratios by Horikawa et al. (2010). The authors suggested that sea-

534   ice formation and according brine rejection led to the subduction of surface waters to

535   intermediate depths. Enhanced formation of sea-ice, acting as the transport agent for

536   terrestrial organic matter would be in accordance with this assumption and explain

537   the higher m_{terr} values. Other studies support the idea of well-ventilated intermediate

538   waters in the Bering Sea and North Pacific during glacial times (Ohkushi et al., 2003;

539   Itaki et al., 2009; Kim et al., 2011) and during severe stadial episodes (Rella et al.,

540   2012). The enhanced formation and/or ventilation of intermediate waters at the end

541   of MIS 6 and during the LGM implied by the record of Horikawa et al. (2010) is not

542   reflected in our $\delta^{15}N_{bulk}$ record, which rather suggests strong stratification during that

543   time. However, these observations are not necessarily contradictory, since

544   intermediate waters could have been formed outside the still-stratified Bering Sea. In

545   fact, recent reconstructions of past ventilation changes in the subarctic North Pacific

546   using radiocarbon-derived ventilation ages in combination with epibenthic stable

547   carbon isotope ratios point to the Okhotsk Sea as the source region of intermediate

548   waters during HS1 and the YD (Max et al., 2014).

549   In addition to insolation forcing, sea-level changes might have influenced the extent

550   of stratification in the Bering Sea. Today, the only shallow (~50 m) Bering Strait

551   allows for oceanic communication between the North Pacific and the N Atlantic.

552   During glacial times the closed Bering Strait prevented the flux of relatively fresh

553   waters into the Atlantic, which is thought to have affected the Atlantic meridional

554   overturning circulation (Hu et al., 2010). Lower glacial sea-level is also likely to have

555   reduced the inflow of Alaskan Stream waters into the Bering Sea (Gorbarenko et al.,

556   2005; Tanaka and Takahashi, 2005). As suggested by relative sea-level

557   reconstructions (e.g., Waelbroeck et al., 2002; Yokoyama and Esat, 2011), the

558   Bering Strait was closed during MIS 6 and in between MIS 4 to 2 until ~12-11 ka BP

559   (Keigwin et al., 2006) (Figure 6). Our $\delta^{15}N_{bulk}$ values indicate almost complete nitrate

560   utilization during late MIS 6 (~150-130 ka BP) and during MIS 3 and 2 as a result of

561   strong stratification. During this time a closed Bering Strait is likely to have fostered

562   stratification due to the pooling of the relatively fresh waters within the Bering Sea,

563   which would have resulted in a strengthened pycnocline. Support for this view and for

564   fresher glacial conditions in the Bering Sea comes from diatom and radiolarian

565   assemblages (Sancetta, 1983; Katsuki and Takahashi, 2005; Tanaka and Takahashi,

566   2005). Notably, during MIS 3 and 2 the $\delta^{15}N_{bulk}$ values recorded at Shirshov Ridge

567   are on average ~1‰ lower than at Bowers Ridge Site 17JPC and ~0.5‰ lower than

568   at Site PC24A (Figure 7). This might indicate that stratification in the Bering Sea was

569   regionally different and more pronounced in the South, or that influence from

570   denitrification resulted in the heavier $\delta^{15}N$ values.

571

572   **5.3 Deglacial and interglacial conditions**

573   In our records, the deglaciations are characterized by the transition from the glacial

574   situation of pronounced stratification with almost complete nitrate utilization and low

575   export production toward the interglacial situation of reduced stratification, high

576   marine productivity, and reduced terrestrial input. Yet, Termination II and Termination

577   I show some notable differences. During Termination I our $\delta^{15}N_{bulk}$ record is

578   characterized by an initial decrease, which might correspond to the HS1 cold phase,

579   subsequent local maxima during the B/A and the early Holocene warm phases, and

580   an intercalated minimum during the YD (Figure 6). The same temporal evolution was

581   reported for sedimentary and diatom-bound $\delta^{15}N$ at Bering Sea sites 17JPC (Brunelle

582   et al., 2007, 2010) and PC24A (Kim et al., 2011). The B/A-peak, occurring

583   simultaneously with a rise in export production, is found in several other records from

584   the North Pacific realm and related to enhanced denitrification (Keigwin et al., 1992;

585 Emmer and Thunell, 2000; Ternois et al., 2001; Kienast et al., 2002; Galbraith et al.,

586 2008a; Kao et al., 2008; Brunelle et al., 2007, 2010; Addison et al., 2012; Khim et al.,

587 2012; Schlung et al., 2013).

588 Notably, m_{terr} shows a local maximum during HS1 at Site 85KL. Hence, the initial

589 decrease in $\delta^{15}N_{bulk}$ might be related to higher terrestrial input or to lower nitrate

590 utilization due to weakened stratification. However, the latter should have resulted in

591 higher export production, which is not observed at our site. This drop is not fully

592 understood and alternative explanations include changes in $\delta^{15}N_{nitrate}$, iron limitation,

593 and light limitation (Brunelle et al., 2007, 2010; Lam et al., 2013). Light limitation by

594 expanded sea-ice coverage is supported by the qualitative detection of IP_{25} in

595 western Bering Sea sediments during HS1 and the YD (Max et al., 2012).

596 During the B/A and the early Holocene our $\delta^{15}N_{bulk}$ values exceeded the modern

597 $\delta^{15}N_{nitrate}$ value, supporting an increase in denitrification. At the same time a shift

598 toward oxygen-depleted bottom water conditions is inferred from benthic foraminiferal

599 assemblages (Kim et al., 2011; Ovsepyan et al., 2013), which is in agreement with

600 the proposed expansion of the OMZ and the occurrence of laminated sediments

601 during warm intervals (e.g., Zheng et al., 2000; van Geen et al., 2003; Cook et al.,

602 2005). A recent comparison between alkenone- and Mg/Ca-based paleotemperature

603 estimates suggests enhanced thermal mixed-layer stratification in the western Bering

604 Sea during the B/A (Riethdorf et al., 2013b), implying that at least some of the

605 recorded $\delta^{15}N_{bulk}$ increase is due to stronger surface nitrate utilization. Recently, Lam

606 et al. (2013) suggested two stepwise events starting with deep convection initialized

607 at ~18 ka BP increasing the nutrient supply but inducing light limitation, and

608 subsequent meltwater-induced stratification resulting in bloom conditions and leaving

609 surface waters enriched in nutrients. The drop in $\delta^{15}N_{bulk}$ observed during the YD in

610 hand with decreasing SSTs and the presence of IP_{25} (Max et al., 2012) argues for a

611 similar situation as recorded during HS1.

612 Termination II differs from Termination I at Site 85KL in such that the $\delta^{15}N_{bulk}$ values

613 are lower and presumably not affected by denitrification. An early deglacial $\delta^{15}N_{bulk}$

614 minimum at ~133 ka BP, followed by a local maximum at ~131 ka BP might hint

615 toward analogs of the HS1 and the B/A, respectively. The subsequent drop in $\delta^{15}N_{bulk}$

616 into MIS 5.5 reflecting the 'breakdown' of glacial stratification is sudden and
617 accompanied by the decrease in both, m_{terr} and bottom water oxygenation then
618 prevailing during the penultimate interglacial (Figure 6). Notably, this drop from
619 relatively high MIS 6 values occurred ~5 kyr before the maximum in insolation was
620 reached, but its timing is comparable to that recorded in the NW Pacific. In the
621 Okhotsk Sea it seems to have occurred significantly earlier at ~147-141 ka BP
622 (Figure 7). A respective drop in $\delta^{15}N_{bulk}$ during Termination I was not recorded at Site
623 85KL until ~9.1 ka BP, while at Bowers Ridge sites U1340 (Schlung et al., 2013) and
624 17JPC (Brunelle et al., 2007) it occurred directly after the B/A maximum.

625

626 **5.4 Millennial-scale oscillations**

627 Riethdorf et al. (2013a) reported on millennial-scale oscillations in core 85KL, thought
628 to reflect increased export production and sudden sea-ice melt, which might be
629 connected to GI (Dansgaard-Oeschger events; e.g., Dansgaard et al., 1993). Similar
630 observations are reported for other sediment cores from the Bering Sea (Gorbarenko
631 et al., 2005, 2010b; Kim et al., 2011; Rella et al., 2012; Schlung et al., 2013) and the
632 Okhotsk Sea (Gorbarenko et al., 2007, 2010a, 2012), indicating warmer SSTs,
633 enhanced marine productivity, weak ventilation of intermediate waters, and poor
634 (dysoxic) dissolved oxygen conditions during interstadials. NE Pacific sediments
635 related to GI are in part laminated and suggested to reflect phases of weak
636 ventilation of NPIW and fluctuations in the strength of the OMZ (Behl and Kennett,
637 1996; Cannariato and Kennett, 1999; Hendy and Kennett, 2000, 2003). Results of
638 Ortiz et al. (2004) from a core off Baja California implied that elevated marine
639 productivity was caused by enhanced nutrient flux to surface waters.

640 In the Bering Sea higher interstadial $\delta^{15}N_{bulk}$ values were explained by Kim et al.
641 (2011) by increased marine productivity as a result of reduced sea-ice influence and
642 a strengthened BSC. They also suggested that stronger inflow of water masses from
643 the Gulf of Alaska (Gorbarenko et al., 2005) resulted in enhanced nutrient supply to
644 Bering Sea surface waters. Schlung et al. (2013) attributed higher $\delta^{15}N_{bulk}$ and
645 concurrent minima in planktonic $\delta^{13}C$ to amplified local upwelling of subsurface
646 nitrate rather than to increased nitrate utilization. Our data show short-lived maxima

647 in $\delta^{15}N_{bulk}$ and concurrent minima in m_{terr} during some, but not all, GI (1, 7, 8, 12, 17-
648 20), and during MIS 6 (at ~133, ~148, and ~173 ka BP) when export production was
649 high (Figure 6). The opposite pattern was recorded when export production was low
650 during some GS (2, 7, 18, 20) and also during MIS 6 (at ~151, ~157, and ~170 ka
651 BP). Despite the low reproducibility of our $\delta^{15}N_{bulk}$ results our data support the view of
652 increased interstadial marine productivity which led to stronger utilization of the
653 available nitrate in a still stratified upper water column. During interstadials warmer
654 SSTs most probably resulted in less sea-ice influence and reduced supply of
655 terrestrial organic matter. Conversely, during stadials strengthened sea-ice formation
656 and coverage is likely to have restricted marine productivity, led to enhanced
657 terrestrial organic matter supply to Bering Sea sediments, and resulted in better
658 ventilation of NPIW, potentially making the Bering Sea a proximate source of this
659 water mass as suggested by Rella et al. (2012).

660

661 **6. Summary**

662 We determined TN, TOC, $\delta^{13}C_{bulk}$ and $\delta^{15}N_{bulk}$ in a core from the western Bering Sea
663 in high-resolution to reconstruct changes in surface nitrate utilization (stratification)
664 over the last 180 kyr. A linear endmember model was applied to assess the
665 contributions of marine- and terrestrial-derived organic matter. Besides the expected
666 difference between glacial and interglacial conditions reported for the subarctic NW
667 Pacific and its marginal seas, our results suggest a more complex evolution of
668 stratification with enhanced vertical mixing during warm intervals (MIS 5.5, 5.3, 5.1,
669 1), and stratification becoming fostered during cold intervals (MIS 6, 5.4, 5.2, 4-2).
670 This development is explained by insolation forcing and a feedback in sea-ice
671 formation and the strength of winter mixing. In addition, sea-level changes might
672 have further influenced the extent of stratification when the Bering Strait was closed
673 and relatively fresh waters pooled in the Bering Sea. During warm intervals,
674 variations in seasonal contrasts, sea-ice influence, and stratification resulted in
675 enhanced export production and dominantly marine-derived organic matter, but less
676 nitrate utilization due to better vertical mixing. Conversely, enhanced terrestrial-
677 derived organic matter, most probably associated with sea-ice formation, low export

678 production, and enhanced stratification characterized cold intervals of the past 180
679 kyr. Moreover, we present supporting evidence that millennial-scale climate
680 oscillations connected with Greenland interstadials occurred in the Bering Sea
681 environment, and that sea-ice formation there influenced the ventilation of North
682 Pacific Intermediate Water.

683

697

698 **References**

699 Addison, J.A., Finney, B.P., Dean, W.E., Davies, M.H., Mix, A.C., Stoner, J.S.,
700 Jaeger, J.M., 2012. Productivity and sedimentary $\delta^{15}N$ variability for the last 17,000
701 years along the northern Gulf of Alaska. Paleoceanography 27, PA1206.
702 doi:10.1029/2011PA002161.

703 Altabet, M.A., Francois, R., 1994. Sedimentary nitrogen isotopic ratio as a recorded
704 for surface ocean nitrate utilization. Global Biogeochem. Cy. 8 (1), 103-116.

705 Altabet, M.A., Pilskaln, C., Thunell, R., Pride, C., Sigman, D., Chavez, F., Francois,
706 R., 1999. The nitrogen isotope biogeochemistry of sinking particles from the margin
707 of the Eastern North Pacific. Deep-Sea Res. Pt. I 46, 655-679.

708 Asahara, Y., Takeuchi, F., Nagashima, K., Harada, N., Yamamoto, K., Oguri, K.,
709 Tadai, O., 2012. Provenance of terrigenous detritus of the surface sediments in the
710 Bering and Chukchi Seas as derived from Sr and Nd isotopes: Implications for recent
711 climate change in the Arctic regions. Deep-Sea Res. Pt. II 61-64, 155-171.
712 doi:10.1016/j.dsr2.2011.12.004.

713 Barford, C.C., Montoya, J.P., Altabet, M.A., Mitchell, R., 1999. Steady-state nitrogen
714 isotope effects of N_2 and N_2O production in *Paracoccus denitrificans*. Appl. Environ.
715 Microb. 65 (3), 989-994.

716 Behl, R.J., Kennett, J.P., 1996. Brief interstadial events in the Santa Barbara basin,
717 NE Pacific, during the past 60 kyr. Nature 379, 243-246.

718 Blockley, S.P.E., Lane, C.S., Hardiman, M., Rasmussen, S.O., Seierstad, I.K.,
719 Steffensen, J.P., Svensson, A., Lotter, A.F., Turney, C.S.M., Ramsey, C.B.,
720 INTIMATE members, 2012. Synchronisation of palaeoenvironmental records over the
721 last 60,000 years, and an extended INTIMATE event stratigraphy to 48,000 b2k.
722 Quaternary Sci. Rev. 36, 2-10. doi:10.1016/j.quascirev.2011.09.017.

723 Brodie, C.R., Casford, J.S.L., Lloyd, J.M., Leng, M.J., Heaton, T.H.E., Kendrick, C.P.,
724 Yongqiang, Z., 2011a. Evidence for bias in C/N, $\delta^{13}C$ and $\delta^{15}N$ values of bulk organic
725 matter, and on environmental interpretation, from a lake sedimentary sequence by
726 pre-analysis acid treatment methods. Quaternary Sci. Rev. 30, 3076-3087.
727 doi:10.1016/j.quascirev.2011.07.003.

728 Brodie, C.R., Heaton, T.H.E., Leng, M.J., Kendrick, C.P., Casford, J.S.L., Lloyd, J.M.,
729 2011b. Evidence for bias in measured $\delta^{15}N$ values of terrestrial and aquatic organic
730 materials due to pre-analysis acid treatment methods. Rapid Commun. Mass Sp. 25,
731 1089-1099. doi:10.1002/rcm.4970.

732 Brunelle, B.G., Sigman, D.M., Cook, M.S., Keigwin, L.D., Haug, G.H., Plessen, B.,
733 Schettler, G., Jaccard, S.L., 2007. Evidence from diatom-bound nitrogen isotopes for
734 subarctic Pacific stratification during the last ice age and a link to North Pacific
735 denitrification changes. Paleoceanography 22, PA1215. doi:10.1029/2005PA001205.

736 Brunelle, B.G., Sigman, D.S., Jaccard, S.L., Keigwin, L.D., Plessen, B., Schettler, G.,
737 Cook, M.S., Haug, G.H., 2010. Glacial/interglacial changes in nutrient supply and

stratification in the western subarctic North Pacific since the penultimate glacial maximum. Quaternary Sci. Rev. 29, 2579-2590. doi:10.1016/j.quascirev.2010.03.010.

Cannariato, K.G., Kennett, J.P., 1999. Climatically related millennial-scale fluctuations in strength of California margin oxygen-minimum zone during the past 60 k.y.. Geology 27 (11), 975-978.

Carpenter, E.J., Harvey, H.R., Fry, B., Capone, D.G., 1997. Biogeochemical tracers of the marine cyanobacterium *Trichodesmium*. Deep-Sea Res. Pt. I 44 (1), 27-38.

Chikamoto, M.O., Menviel, L., Abe-Ouchi, A., Ohgaito, R., Timmermann, A., Okazaki, Y., Harada, N., Oka, A., Mouchet, A., 2012. Variability in North Pacific intermediate and deep water ventilation during Heinrich events in two coupled climate models. Deep-Sea Res. Pt. II 61-64, 114-126. doi:10.1016/j.dsr2.2011.12.002.

Codispoti, L.A., Brandes, J.A., Christensen, J.P., Devol, A.H., Naqvi, S.W.A., Paerl, H.W., Yoshinari, T., 2001. The oceanic fixed nitrogen and nitrous oxide budgets: Moving targets as we enter the anthropocene?. Sci. Mar. 65 (Suppl. 2), 85-105.

Cook, M.S., Keigwin, L.D., Sancetta, C.A., 2005. The deglacial history of surface and intermediate water of the Bering Sea. Deep-Sea Res. Pt. II 52, 2163-2173. doi:10.1016/j.dsr2.2005.07.004.

Crusius, J., Pedersen, T.F., Kienast, S., Keigwin, L., Labeyrie, L., 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 (7), 633-363. doi: 10.1130/G20508.1.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjörnsdottir, A.E., Jouzel, J., Bond, G., 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. Nature 364, 218-220.

Debret, M., Desmet, M., Balsam, W., Copard, Y., Francus, P., Laj, C., 2006. Spectrophotometer analysis of Holocene sediments from an anoxic fjord: Saanich Inlet, British Columbia, Canada. Mar. Geol. 229, 15-28. doi:10.1016/j.margeo.2006.01.005.

768　Dullo, W.-C., Baranov, B., van den Bogaard, C. (Eds.), 2009. FS Sonne
769　Fahrtbericht/Cruise Report SO201-2 KALMAR, Busan/Korea-Tomakomai/Japan,
770　30.08.-08.10.2009. IFM-GEOMAR Report 35, Leibniz Institute of Marine Sciences,
771　Kiel, 233 pp.

772　Elderfield, H., Rickaby, R.E.M., 2000. Oceanic Cd/P ratio and nutrient utilization in
773　the glacial Southern Ocean. Nature 405, 305-310.

774　Emile-Geay, J., Cane, M., Naik, N., Seager, R., Clement, A.C., van Geen, A., 2003.
775　Warren revisited: Atmospheric freshwater fluxes and "Why is no deep water formed
776　in the North Pacific". J. Geophys. Res. 108 (C6), 3178. doi:10.1029/2011JC001058.

777　Emmer, E., Thunell, R.C., 2000. Nitrogen isotope variations in Santa Barbara Basin
778　sediments: Implications for denitrification in the eastern tropical North Pacific during
779　the last 50,000 years. Paleoceanography 15 (4), 377-387.
780　doi:10.1029/1999PA000417.

781　Francois, R., Altabet, M.A., Yu, E.-F., Sigman, D.M., Bacon, M.P., Frank, M.,
782　Bohrmann, G., Bareille, G., Labeyrie, L.D., 1997. Contribution of Southern Ocean
783　surface-water stratification to low atmosphere CO_2 concentrations during the last
784　glacial period. Nature 389, 929-935.

785　Galbraith, E.D., Kienast, M., Pedersen, T.F., Calvert, S.E., 2004. Glacial-interglacial
786　modulation of the marine nitrogen cycle by high-latitude O_2 supply to the global
787　thermocline. Paleoceanography 19, PA4007. doi:10.1029/2003PA001000.

788　Galbraith, E.D., Kienast, M., Jaccard, S.L., Pedersen, T.F., Brunelle, B.G., Sigman,
789　D.M., Kiefer, T., 2008a. Consistent relationship between global climate and surface
790　nitrate utilization in the western subarctic Pacific throughout the last 500 ka.
791　Paleoceanography 23, PA2212. doi:10.1029/2007PA001518.

792　Galbraith, E.D., Sigman, D.M., Robinson, R.S., Pedersen, T.F., 2008b. Nitrogen in
793　past marine environments, in: Capone, D.G., Bronk, D.A., Mulholland, M.R.,
794　Carpenter, E.J. (Eds.), Nitrogen in the Marine Environment. Second Edition, Elsevier
795　Inc., pp. 1497-1535. doi:10.1016/B978-0-12-372522-6.00034-7.

796　Garcia, H.E., Locarnini, R.A., Boyer, T.P., Antonov, J.I., 2010. World Ocean Atlas
797　2009, Volume 4: Nutrients (phosphate, nitrate, and silicate), in: Levitus, S. (Ed.),

798 NOAA Atlas NESDIS 71. U.S. Government Printing Office, Washington, D.C., 184
799 pp.

800 Gaye-Haake, B., Lahajnar, N., Emeis, K.-Ch., Unger, D., Rixen, T., Suthhof, A.,
801 Ramaswamy, V., Schulz, H., Paropkari, A.L., Guptha, M.V.S., Ittekkot, V., 2005.
802 Stable nitrogen isotopic ratios of sinking particles and sediments from the northern
803 Indian Ocean. Mar. Chem. 96, 243-255. doi:10.1016/j.marchem.2005.02.001.

804 Geider, R.J., La Roche, J., 2002. Redfield revisited: variability of C:N:P in marine
805 microalgae and its biochemical basis. Eur. J. Phycol. 37, 1-17.
806 doi:10.1017/S0967026201003456.

807 Goñi, M.A., Ruttenberg, K.C., Eglinton, T.I., 1998. A reassessment of the sources
808 and importance of land-derived organic matter in surface sediments from the Gulf of
809 Mexico. Geochim. Cosmochim. Ac. 62 (18), 3055-3075.

810 Gorbarenko, S.A., Khusid, T.A., Basov, I.A., Oba, T., Southon, J.R., Koizumi, I.,
811 2002. Glacial Holocene environment of the southeastern Okhotsk Sea: Evidence
812 from geochemical and palaeontological data. Palaeogeogr. Palaeocl. 177, 237-263.
813 doi:10.1016/S0031-0182(01)00335-2.

814 Gorbarenko, S.A., Basov, I.A., Chekhovskaya, M.P., Southon, J., Khusid, T.A.,
815 Artemova, A.V., 2005. Orbital and millennium scale environmental changes in the
816 southern Bering Sea during the last glacial-Holocene: Geochemical and
817 paleontological evidence. Deep-Sea Res. Pt. II 52, 2174-2185.
818 doi:10.1016/j.dsr2.2005.08.005.

819 Gorbarenko, S.A., Goldberg, E.L., Kashgarian, M., Velivetskaya, T.A., Zakharkov,
820 S.P., Pechnikov, V.S., Bosin, A.A., Psheneva, O.Y., Ivanova, E.D., 2007. Millennium
821 scale environment changes of the Okhotsk Sea during last 80 kyr and their phase
822 relationship with global climate changes. J. Oceanogr. 63, 609-623.

823 Gorbarenko, S.A., Harada, N., Malakhov, M.I., Vasilenko, Y.P., Bosin, A.A.,
824 Goldberg, E.L., 2010a. Orbital and millennial-scale environmental and
825 sedimentological changes in the Okhotsk Sea during the last 350 kyr. Global Planet.
826 Change 72, 79-85. doi:10.1016/j.gloplacha.2010.03.002.

827   Gorbarenko, S.A., Wang, P., Wang, R., Cheng, X., 2010b. Orbital and suborbital
828   environmental changes in the southern Bering Sea during the last 50 kyr.
829   Palaeogeogr. Palaeocl. 286, 97-106. doi:10.1016/j.palaeo.2009.12.014.

830   Gorbarenko, S.A., Harada, N., Malakhov, M.I., Velivetskaya, T.A., Vasilenko, Y.P.,
831   Bosin, A.A., Derkachev, A.N., Goldberg, E.L., Ignatiev, A.V., 2012. Responses of the
832   Okhotsk Sea environment and sedimentology to global climate changes at the orbital
833   and millennial scale during the last 350 kyr. Deep-Sea Res. Pt. II 61-64, 73-84.
834   doi:10.1016/j.dsr2.2011.05.016.

835   Guo, L., Macdonald, R.W., 2006. Source and transport of terrigenous organic matter
836   in the upper Yukon River: Evidence, from isotopic (δ^{13}C, Δ^{14}C, and δ^{15}N) composition
837   of dissolved, colloidal, and particulate phases. Global Biogeochem. Cy. 20, GB2011.
838   doi:10.1029/2005GB002593.

839   Guo, L., Tanaka, T., Wang, D., Tanaka, N., Murata, A., 2004. Distributions,
840   speciation and stable isotope composition of organic matter in the southeastern
841   Bering Sea. Mar. Chem. 91, 211-226. doi:10.1016/j.marchem.2004.07.002.

842   Haug, G.H., Sigman, D.M., Tiedemann, R., Pedersen, T.F., Sarnthein, M., 1999.
843   Onset of permanent stratification in the subarctic Pacific Ocean. Nature 401, 779-
844   782.

845   Hendy, I.L., Kennett, J.P., 2000. Dansgaard-Oeschger cycles and the California
846   Current System: Planktonic foraminiferal response to rapid climate change in Santa
847   Barbara Basin, Ocean Drilling Program hole 893A. Paleoceanography 15 (1), 30-42.

848   Hendy, I.L., Kennett, J.P., 2003. Tropical forcing of North Pacific intermediate water
849   distribution during Late Quaternary rapid climate change? Quaternary Sci. Rev. 22,
850   673-689. doi:10.1016/S0277-3791(02)00186-5.

851   Honda, M.C., Imai, K., Nojiri, Y., Hoshi, F., Sugawara, T., Kusakabe, M., 2002. The
852   biological pump in the northwestern North Pacific based on fluxes and major
853   components of particulate matter obtained by sediment-trap experiments (1997-
854   2000). Deep-Sea Res. Pt. II 49, 5595-5625. doi:10.1016/S0967-0645(02)00201-1.

855   Horikawa, K., Asahara, Y., Yamamoto, K., Okazaki, Y., 2010. Intermediate water
856   formation in the Bering Sea during glacial periods: Evidence from neodymium isotope
857   ratios. Geology 38 (5), 435-438. doi:10.1130/G30225.1.

858   Hu, A., Meehl, G.A., Otto-Bliesner, B.L., Waelbroeck, C., Han, W., Loutre, M.-F.,
859   Lambeck, K., Mitrovica, J.X., Rosenbloom, N., 2010. Influence of Bering Strait flow
860   and North Atlantic circulation on glacial sea-level changes. Nat. Geosci. 3, 118-121.
861   doi:10.1038/ngeo0729.

862   Itaki, T., Uchida, M., Kim, S., Shin, H.-S., Tada, R., Khim, B.-K., 2009. Late
863   Pleistocene stratigraphy and palaeoceanographic implications in northern Bering Sea
864   slope sediments: Evidence from the radiolarian species *Cycladophora davisiana*. J.
865   Quaternary Sci. 24 (8), 856-865. doi:10.1002/jqs.1356.

866   Jaccard, S.L., Galbraith, E.D., 2012. Large climate-driven changes of oceanic oxygen
867   concentrations during the last deglaciation. Nat. Geosci. 5 (2), 151-156.
868   doi:10.1038ngeo1352.

869   Jaccard, S.L., Galbraith, E.D., 2013. Direct ventilation of the North Pacific did not
870   reach the deep ocean during the last deglaciation. Geophys. Res. Lett. 40, 199-203.
871   doi:10.1029/2012GL054118.

872   Jaccard, S.L., Haug, G.H., Sigman, D.M., Pedersen, T.F., Thierstein, H.R., Röhl, U.,
873   2005. Glacial/interglacial changes in subarctic North Pacific stratification. Science
874   308, 1003-1006. doi:10.1126/science.1108696.

875   Jaccard, S.L., Galbraith, E.D., Sigman, D.M., Haug, G.H., 2010. A pervasive link
876   between Antarctic ice core and subarctic Pacific sediment records over the past 800
877   kyrs. Quaternary Sci. Rev. 29, 206-212. doi:10.1016/j.quascirev.2009.10.007.

878   Kao, S.J., Liu, K.K., Hsu, S.C., Chang, Y.P., Dai, M.H., 2008. North Pacific-wide
879   spreading of isotopically heavy nitrogen during the last deglaciation: Evidence from
880   the western Pacific. Biogeosciences 5, 1641-1650. doi:10.5194/bg-5-1641-2008.

881   Kashiyama, Y., Ogawa, N.O., Shiro, M., Tada, R., Kitazato, H., Ohkouchi, N., 2008.
882   Reconstruction of the biogeochemistry and ecology of photoautotrophs based on the
883   nitrogen and carbon isotopic compositions of vanadyl porphyrins from Miocene
884   siliceous sediments. Biogeosciences 5, 797-816. doi:10.5194/bg-5-797-2008.

885 Katsuki, K., Takahashi, K., 2005. Diatoms as paleoenvironmental proxies for
886 seasonal productivity, sea-ice and surface circulation in the Bering Sea during the
887 late Quaternary. Deep-Sea Res. Pt. II 52, 2110-2130.
888 doi:10.1016/j.dsr2.2005.07.001.

889 Keigwin, L.D., Jones, G.A., Froelich, P.N., 1992. A 15,000 year paleoenvironmental
890 record from Meiji Seamount, far northwestern Pacific. Earth Planet. Sc. Lett. 111,
891 425-440.

892 Keigwin, L.D., Donnelly, J.P., Cook, M.S., Driscoll, N.W., Brigham-Grette, J., 2006.
893 Rapid sea-level rise and Holocene climate in the Chukchi Sea. Geology 34 (10), 861-
894 864. doi:10.1130/G22712.1.

895 Khim, B.-K., Sakamoto, T., Harada, N., 2012. Reconstruction of surface water
896 conditions in the central region of the Okhotsk Sea during the last 180 kyrs. Deep-
897 Sea Res. Pt. II 61-64, 63-72. doi:10.1016/j.dsr2.2011.05.014.

898 Kienast, S.S., Calvert, S.E., Pedersen, T.F., 2002. Nitrogen isotope and productivity
899 variations along the northeast Pacific margin over the last 120 kyr: Surface and
900 subsurface paleoceanography. Paleoceanography 17 (4), 1055.
901 doi:10.1029/2001PA000650.

902 Kienast, S.S., Hendy, I.L., Crusius, J., Pedersen, T.F., Calvert, S.E., 2004. Export
903 production in the subarctic North Pacific over the last 800 kyrs: No evidence for iron
904 fertilization? J. Oceanogr. 60 (1), 189-203.

905 Kim, S., Khim, B.K., Uchida, M., Itaki, T., Tada, R., 2011. Millennial-scale
906 paleoceanographic events and implication for the intermediate-water ventilation in
907 the northern slope area of the Bering Sea during the last 71 kyrs. Global Planet.
908 Change 79, 89-98. doi:10.1016/j.gloplacha.2011.08.004.

909 Lam, P.J., Robinson, L.F., Blusztajn, J., Li, C., Cook, M.S., McManus, J.F., Keigwin,
910 L.D., 2013. Transient stratification as the cause of the North Pacific productivity spike
911 during deglaciation. Nat. Geosci. 6, 622-626. doi:10.1038/ngeo1873.

912 Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A.C.M., Levrard, B., 2004.
913 A long-term numerical solution for the insolation quantities of the Earth. Astron.
914 Astrophys. 428, 261-285. doi:10.1051/0004-6361:20041335.

915 Lehmann, M.F., Sigman, D.M., McCorkle, D.C., Brunelle, B.G., Hoffmann, S.,
916 Kienast, M., Cane, G., Clement, J., 2005. Origin of the deep Bering Sea nitrate
917 deficit: Constraints from the nitrogen and oxygen isotopic composition of water
918 column nitrate and benthic nitrate fluxes. Global Biogeochem. Cy. 19, GB4005.
919 doi:10.1029/2005GB002508.

920 Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally
921 distributed benthic $\delta^{18}O$ records. Paleoceanography 20, PA1003.
922 doi:10.1029/2004PA001071.

923 Liu, K.-K., Kaplan, I.R., 1989. The eastern tropical Pacific as a source of ^{15}N-enriched
924 nitrate in seawater off southern California. Limnol. Oceanogr. 34 (5), 820-830.

925 Macdonald, A.M., Suga, T., Curry, R.G., 2001. An isopycnally averaged North Pacific
926 climatology. J. Atmos. Ocean. Tech. 18 (3), 394-420.

927 Mariotti, A., Germon, J.C., Hubert, P., Kaiser, P., Letolle, R., Tardieux, A., Tardieux,
928 P., 1981. Experimental determination of nitrogen kinetic isotope fractionation: Some
929 principles; Illustration for the denitrification and nitrification processes. Plant Soil 62
930 (3), 413-430. doi:10.1007/BF02374138.

931 Matsumoto, K., Oba, T., Lynch-Stieglitz, J., Yamamoto, H., 2002. Interior
932 hydrography and circulation of the glacial Pacific Ocean. Quaternary Sci. Rev. 21,
933 1693-1704. doi:10.1016/S0277-3791(01)00142-1.

934 Max, L., Riethdorf, J.-R., Tiedemann, R., Smirnova, M., Lembke-Jene, L., Fahl, K.,
935 Nürnberg, D., Matul, A., Mollenhauer, G., 2012. Sea surface temperature variability
936 and sea-ice extent in the subarctic northwest Pacific during the past 15,000 years.
937 Paleoceanography 27, PA3213. doi:10.1029/2012PA002292.

938 Max, L., Lembke-Jene, L., Riethdorf, J.-R., Tiedemann, R., Nürnberg, D., Kühn, H.,
939 Mackensen, A., 2014. Pulses of enhanced North Pacific Intermediate Water
940 ventilation from the Okhotsk Sea and Bering Sea during the last deglaciation. Clim.
941 Past 10, 591-605. doi:10.5194/cp-10-591-2014.

942 McQuoid, M.R., Whiticar, M.J., Calvert, S.E., Pedersen, T.F., 2001. A post-glacial
943 isotope record of primary production and accumulation in the organic sediments of
944 Saanich Inlet, ODP Leg 169S. Mar. Geol. 174, 273-286.

945 Méheust, M., Fahl, K., Stein, R., 2013. Variability in modern sea surface temperature, sea ice and terrigenous input in the sub-polar North Pacific and Bering Sea: Reconstruction from biomarker data. Org. Geochem. 57, 54-64. doi:10.1016/j.orggeochem.2013.01.008i.

949 Menviel, L., Timmermann, A., Elison Timm, O., Mouchet, A., Abe-Ouchi, A., Chikamoto, M.O., Harada, N., Ohgaito, R., Okazaki, Y., 2012. Removing the North Pacific halocline: Effects on global climate, ocean circulation and the carbon cycle. Deep-Sea Res. Pt. II 61-64, 106-113. doi:10.1016/j.dsr2.2011.03.005.

953 Meyers, P.A., 1994. Preservation of elemental and isotopic source identification of sedimentary organic matter. Chem. Geol. 114, 289-302.

955 Müller, P.J., Schneider, R., 1993. An automated leaching method for the determination of opal in sediments and particulate matter. Deep-Sea Res. Pt. I 40 (3), 425-444.

958 Nagashima, K., Asahara, Y., Takeuchi, F., Harada, N., Toyoda, S., Tada, R., 2012. Contribution of detrital materials from the Yukon River to the continental shelf sediments of the Bering Sea based on the electron spin resonance signal intensity and crystallinity of quartz. Deep-Sea Res. Pt. II 61-64, 145-154. doi:10.1016/j.dsr2.2011.12.001.

963 Nakatsuka, T., Watanabe, K., Handa, N., Matsumoto, E., 1995. Glacial to interglacial surface nutrient variations of Bering deep basins recorded by δ^{13}C and δ^{15}N of sedimentary organic matter. Paleoceanography 10 (6), 1047-1061.

966 Narita, H., Sato, M., Tsunogai, S., Murayama, M., Ikehara, M., Nakatsuka, T., Wakatsuchi, M., Harada, N., Ujiié, Y., 2002. Biogenic opal indicating less productive northwestern North Pacific during the glacial ages. Geophys. Res. Lett. 29 (15), 1732. doi:10.1029/2001GL014320.

970 Needoba, J.A., Waser, N.A., Harrison, P.J., Calvert, S.E., 2003. Nitrogen isotope fractionation in 12 species of marine phytoplankton during growth on nitrate. Mar. Ecol.-Prog. Ser. 255, 81-91. doi:10.3354/meps255081.

973 Niebauer, H.J., Bond, N.A., Yakunin, L.P., Plotnikov, V.V., 1999. An update on the climatology and sea ice of the Bering Sea, in: Loughlin, T.R., Ohtani, K. (Eds.),

975   Dynamics of the Bering Sea. University of Alaska Sea Grant, Fairbanks, Alaska, pp.
976   29-59.

977   North Greenland Ice Core Project members, 2004. High-resolution record of Northern
978   Hemisphere climate extending into the last interglacial period. Nature 431, 147-151.
979   doi:10.1038/nature02848.

980   Nürnberg, D., Tiedemann, R., 2004. Environmental change in the Sea of Okhotsk
981   during the last 1.1 million years. Paleoceanography 19, PA4011.
982   doi:10.1029/2004PA001023.

983   Nürnberg, D., Dethleff, D., Tiedemann, R., Kaiser, A., Gorbarenko, S.A., 2011.
984   Okhotsk Sea ice coverage and Kamchatka glaciation over the last 350 ka - Evidence
985   from ice-rafted debris and planktonic δ^{18}O. Palaeogeogr. Palaeocl. 310, 191-205.
986   doi:10.1016/j.palaeo.2011.07.011.

987   Ohkushi, K., Itaki, T., Nemoto, N., 2003. Last Glacial-Holocene change in
988   intermediate-water ventilation in the Northwestern Pacific. Quaternary Sci. Rev. 22,
989   1477-1484. doi:10.1016/S0277-3791(03)00082-9.

990   Okazaki, Y., Takahashi, K., Asahi, H., Katsuki, K., Hori, J., Yasuda, H., Sagawa, Y.,
991   Tokuyama, H., 2005a. Productivity changes in the Bering Sea during the late
992   Quaternary. Deep-Sea Res. Pt. II 52, 2150-2162. doi:10.1016/j.dsr2.2005.07.003.

993   Okazaki, Y., Takahashi, K., Katsuki, K., Ono, A., Hori, J., Sakamoto, T., Uchida, M.,
994   Shibata, Y., Ikehara, M., Aoki, K., 2005b. Late Quaternary paleoceanographic
995   changes in the southwestern Okhotsk Sea: Evidence from geochemical, radiolarian,
996   and diatom records. Deep-Sea Res. Pt. II 52, 2332-2350.
997   doi:10.1016/j.dsr2.2005.07.007.

998   Okazaki, Y., Timmermann, A., Menviel, L., Harada, N., Abe-Ouchi, A., Chikamoto,
999   M.O., Mouchet, A., Asahi, H., 2010. Deepwater formation in the North Pacific during
1000   the last glacial termination. Science 329, 200-204. doi:10.1126/science.1190612.

1001   Okazaki, Y., Sagawa, T., Asahi, H., Horikawa, K., Onodera, J., 2012. Ventilation
1002   changes in the western North Pacific since the last glacial period. Clim. Past 8, 17-
1003   24. doi:10.5194/cp-8-17-2012.

1004   Ortiz, J.D., O'Connell, S.B., DelViscio, J., Dean, W., Carriquiry, J.D., Marchitto, T.,
1005   Zheng, Y., van Geen, A., 2004. Enhanced marine productivity off western North
1006   America during warm climate intervals of the past 52 k.y.. Geology 32 (6), 521-524.
1007   doi:10.1130/G20234.1.

1008   Ovsepyan, E.A., Ivanova, E.V., Max, L., Riethdorf, J.-R., Nürnberg, D., Tiedemann,
1009   R., 2013. Late Quaternary oceanographic conditions in the western Bering Sea.
1010   Oceanology 53 (2), 211-222. doi:10.1134/S0001437013020136.

1011   Pennock, J.R., Velinsky, D.J., Ludlam, J.M., Sharp, J.H., Fogel, M.L., 1996. Isotopic
1012   fractionation of ammonium and nitrate during uptake by *Skeletonema costatum*:
1013   Implications for $\delta^{15}N$ dynamics under bloom conditions. Limnol. Oceanogr. 41 (3),
1014   451-459.

1015   Perdue, E.M., Koprivnjak, J.-F., 2007. Using the C/N ratio to estimate terrigenous
1016   inputs of organic matter to aquatic environments. Estuar. Coast. Shelf S. 73, 65-72.
1017   doi:10.1016/j.ecss.2006.12.021.

1018   Rasmussen, S.O., Andersen, K.K., Svensson, A.M., Steffensen, J.P., Vinther, B.M.,
1019   Clausen, H.B., Siggaard-Andersen, M.-L., Johnsen, S.J., Larsen, L.B., Dahl-Jensen,
1020   D., Bigler, M., Röthlisberger, R., Fischer, H., Goto-Azuma, K., Hansson, M.E., Ruth,
1021   U., 2006. A new Greenland ice core chronology for the last glacial termination. J.
1022   Geophys. Res. 111, D06102. doi:10.1029/2005JD006079.

1023   Rella, S.F., Tada, R., Nagashima, K., Ikehara, M., Itaki, T., Ohkushi, K., Sakamoto,
1024   T., Harada, N., Uchida, M., 2012. Abrupt changes of intermediate water properties on
1025   the northeastern slope of the Bering Sea during the last glacial and deglacial period.
1026   Paleoceanography 27, PA3203. doi:10.1029/2011PA002205.

1027   Riethdorf, J.-R., Nürnberg, D., Max, L., Tiedemann, R., Gorbarenko, S.A., Malakhov,
1028   M.I., 2013a. Millennial-scale variability of marine productivity of marine productivity
1029   and terrigenous matter supply in the western Bering Sea over the past 180 kyr. Clim.
1030   Past 9, 1345-1373. doi:10.5194/cp-9-1345-2013.

1031   Riethdorf, J.-R., Max, L., Nürnberg, D., Lembke-Jene, L., Tiedemann, R., 2013b.
1032   Deglacial development of (sub) sea surface temperature and salinity in the subarctic

northwest Pacific: Implications for upper-ocean stratification. Paleoceanography 28, 91-104. doi:10.1002/palo.20014.

Robinson, R.S., Kienast, M., Albuquerque, A.L., Altabet, M., Contreras, S., De Pol Holz, R., Dubois, N., Francois, R., Galbraith, E., Hsu, T.-C., Ivanochko, T., Jaccard, S., Kao, S.-J., Kiefer, T., Kienast, S., Lehmann, M.F., Martinez, P., McCarthy, M., Möbius, J., Pedersen, T., Quan, T.M., Ryabenko, E., Schmittner, A., Schneider, R., Schneider-Mor, A., Shigemitsu, M., Sinclair, D., Somes, C., Studer, A., Thunell, R., Yang, J.-Y., 2012. A review of nitrogen isotopic alteration in marine sediments. Paleoceanography 27, PA4203. doi:10.1029/2012PA002321.

Roden, G.I., 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.

Roden, G.I., 2000. Flow and water property structures between the Bering Sea and Fiji in the summer of 1993. J. Geophys. Res. 105 (C12), 28595-28612. doi:10.1029/1999JC000030.

Sancetta, C., 1983. Effect of Pleistocene glaciation upon oceanographic characteristics of the North Pacific Ocean and Bering Sea. Deep-Sea Res 30 (8A), 851-869.

Sarnthein, M., Stattegger, K., Dreger, D., Erlenkeuser, H., Grootes, P., Haupt, B.J., Jung, S., Kiefer, T., Kuhnt, W., Pflaumann, U., Schäfer-Neth, C., Schulz, H., Schulz, M., Seidov, D., Simstich, J., van Kreveld, S., Vogelsang, E., Völker, A., Weinelt, M., 2001. Fundamental modes and abrupt changes in North Atlantic circulation and climate over the last 60 ky - Concepts, reconstruction and numerical modeling, in: Schäfer, P., Ritzrau, W., Schlüter, M., Thiede, J. (Eds.),The Northern North Atlantic: A Changing Environment, Springer, Berlin, pp. 365-410.

Schlitzer, R., 2013. Ocean Data View, http://odv.awi.de (last access: 4 February 2013).

Schlung, S.A., Ravelo, A.C., Aiello, I.W., Andreasen, D.H., Cook, M.S., Drake, M., Dyez, K.A., Guilderson, T.P., LaRiviere, J., Stroynowski, Z., Takahashi, K., 2013. Millennial-scale climate change and intermediate water circulation in the Bering Sea

1063   from 90 ka: A high-resolution record from IODP Site U1340. Paleoceanography 28,

1064   1-14. doi:10.1029/2012PA002365.

1065   Seki, O., Ikehara, M., Kawamura, K., Nakatsuka, T., Ohnishi, K., Wakatsuchi, M.,

1066   Narita, H., Sakamoto, T., 2004. Reconstruction of paleoproductivity in the Sea of

1067   Okhotsk over the last 30 kyr. Paleoceanography 19, PA1016.

1068   doi:10.1029/2002PA000808.

1069   Shigemitsu, M., Narita, H., Watanabe, Y.W., Harada, N., Tsunogai, S., 2007. Ba, Si,

1070   U, Al, Sc, La, Th, C and $^{13}C/^{12}C$ in a sediment core in the western subarctic Pacific

1071   as proxies of past biological production. Mar. Chem. 106, 442-455.

1072   doi:10.1016/j.marchem.2007.04.004.

1073   Shigemitsu, M., Watanabe, Y.W., Narita, H., 2008. Time variations of $\delta^{15}N$ of organic

1074   nitrogen in deep western subarctic Pacific sediment over the last 145 ka. Geochem.

1075   Geophy. Geosy. 9 (10), Q10012. doi:10.1029/2008GC001999.

1076   Sigman, D.M., Altabet, M.A., McCorkle, D.C., Francois, R., Fischer, G., 1999. The

1077 $\delta^{15}N$ of nitrate in the Southern Ocean: Consumption of nitrate in surface waters.

1078   Global Biogeochem. Cy. 13 (4), 1149-1166. doi:10.1029/1999GB900038.

1079   Sigman, D.M., Altabet, M.A., McCorkle, D.C., Francois, R., Fischer, G., 2000. The

1080 $\delta^{15}N$ of nitrate in the Southern Ocean: Nitrogen cycling and circulation in the ocean

1081   interior. J. Geophys. Res. 105 (C8), 19599-19614.

1082   Sigman, D.M., Lehman, S.J., Oppo, D.W., 2003. Evaluating mechanisms of nutrient

1083   depletion and ^{13}C enrichment in the intermediate-depth Atlantic during the last ice

1084   age. Paleoceanography 18 (3), 1072. doi:10.1029/2002PA000818.

1085   Sigman, D.M., Hain, M.P., Haug, G.H., 2010. The polar ocean and glacial cycles in

1086   atmospheric CO_2 concentration. Nature 466, 47-55. doi:10.1038/nature09149.

1087   Smith, S.L., Henrichs, S.M., Rho, T., 2002. Stable C and N isotopic composition of

1088   sinking particles and zooplankton over the southeastern Bering Sea shelf. Deep-Sea

1089   Res. Pt. II 49, 6031-6050. doi:10.1016/S0967-0645(02)00332-6.

1090   Somes, C.J., Schmittner, A., Galbraith, E.D., Lehmann, M.F., Altabet, M.A., Montoya,

1091   J.P., Letelier, R.M., Mix, A.C., Bourbonnais, A., Eby, M., 2010. Simulating the global

1092   distribution of nitrogen isotopes in the ocean. Global Biogeochem. Cy. 24, GB4019.
1093   doi:10.1029/2009GB003767.

1094   Springer, A.M., McRoy, C.P., Flint, M.V., 1996. The Bering Sea Green Belt: Shelf-
1095   edge processes and ecosystem production. Fish. Oceanogr. 5, 205-223.

1096   Stabeno, P.J., Schumacher, J.D., Ohtani, K., 1999. The physical oceanography of
1097   the Bering Sea, in: Loughlin, T.R., Ohtani, K. (Eds.), Dynamics of the Bering Sea.
1098   University of Alaska Sea Grant, Fairbanks, Alaska, pp. 1-28.

1099   Takahashi, K., Fujitani, N., Yanada, M., 2002a. Long term monitoring of particle
1100   fluxes in the Bering Sea and the central subarctic Pacific Ocean, 1990-2000. Progr.
1101   Oceanogr. 55, 95-112. doi:10.1016/S0079-6611(02)00072-1.

1102   Takahashi, T., Sutherland, S.C., Sweeney, C., Poisson, A., Metzl, N., Tilbrook, B.,
1103   Bates, N., Wanninkhof, R., Feely, R.A., Sabine, C., Olafsson, J., Nojiri, Y., 2002b.
1104   Global sea-air CO_2 flux based on climatological surface ocean pCO_2, and seasonal
1105   biological   and   temperature   effects.   Deep-Sea   Res..   Pt.   II   49,   1601-1622.
1106   doi:10.1016/S0967-0645(02)00003-6.

1107   Tanaka, S., Takahashi, K., 2005. Late Quaternary paleoceanographic changes in the
1108   Bering Sea and the western subarctic Pacific based on radiolarian assemblages.
1109   Deep-Sea Res. Pt. II 52, 2131-2149. doi:10.1016/j.dsr2.2005.07.002.

1110   Ternois, Y., Kawamura, K., Keigwin, L., Ohkouchi, N., Nakatsuka, T., 2001. A
1111   biomarker approach for assessing marine and terrigenous inputs to the sediments of
1112   Sea of Okhotsk for the last 27,000 years. Geochim. Cosmochim. Ac. 65 (5), 791-802.

1113   Thunell, R.C., Sigman, D.M., Muller-Karger, F., Astor, Y., Varela, R., 2004. Nitrogen
1114   isotope dynamics of the Cariaco Basin, Venezuela. Global Biogeochem. Cy. 18,
1115   GB3001. doi:10.1029/2003GB002185.

1116   Tomczak, M., Godfrey, J.S., 1994. Regional oceanography: An introduction. Elsevier
1117   Science Ltd., Oxford, 391 pp.

1118   Tsuda, A., Takeda, S., Saito, H., Nishioka, J., Nojiri, Y., Kudo, I., Kiyosawa, H.,
1119   Shiomoto, A., Imai, K., Ono, T., Shimamoto, A., Tsumune, D., Yoshimura, T., Aono,
1120   T., Hinuma, A., Kinugasa, M., Suzuki, K., Sohrin, Y., Noiri, Y., Tani, H., Deguchi, Y.,
1121   Tsurushima, N., Ogawa, H., Fukami, K., Kuma, K., Saino, T., 2003. A mesoscale iron

1122 enrichment in the western subarctic Pacific induces a large centric diatom bloom.
1123 Science 300, 958-961. doi:10.1126/science.1082000.

1124 Tyrrell, T., Merico, A., Waniek, J.J., Wong, C.S., Metzl, N., Whitney, F., 2005. Effect
1125 of seafloor depth on phytoplankton blooms in high-nitrate, low-chlorophyll (HNLC)
1126 regions. J. Geophys. Res. 110, G02007. doi:10.1029/2005JG000041.

1127 van Geen, A., Zheng, Y., Bernhard, J.M., Cannariato, K.G., Carriquiry, J., Dean,
1128 W.E., Eakins, B.W., Ortiz, J.D., Pike, J., 2003. On the preservation of laminated
1129 sediments along the western margin of North America. Paleoceanography 18 (4),
1130 1098. doi:10.1029/2003PA000911.

1131 VanLaningham, S., Pisias, N.G., Duncan, R.A., Clift, P.D., 2009. Glacial-interglacial
1132 sediment transport to the Meiji Drift, northwest Pacific Ocean: Evidence for timing of
1133 Beringian outwashing. Earth Planet. Sc. Lett. 277, 64-72.
1134 doi:10.1016/j.epsl.2008.09.033.

1135 Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J.C., McManus, J.F., Lambeck,
1136 K., Balbon, E., Labracherie, M., 2002. Sea-level and deep water temperature
1137 changes derived from benthic foraminifera isotopic records. Quaternary Sci. Rev. 21,
1138 295-305. doi:10.1016/S0277-3791(01)00101-9.

1139 Walinsky, S.E., Prahl, F.G., Mix, A.C., Finney, B.P., Jaeger, J.M., Rosen, G.P., 2009.
1140 Distribution and composition of organic matter in surface sediments of coastal
1141 Southeast Alaska. Cont. Shelf Res. 29, 1565-1579. doi:10.1016/j.csr.2009.04.006.

1142 Walsh, E.M., Ingalls, A.E., Keil, R.G., 2008. Sources and transport of terrestrial
1143 organic matter in Vancouver Island fjords and the Vancouver-Washington Margin: A
1144 multiproxy approach using $\delta^{13}C_{org}$, lignin phenols, and the ether lipid BIT index.
1145 Limnol. Oceanogr. 53 (3), 1054-1063. doi:10.4319/lo.2008.53.3.1054.

1146 Wang, Y., Cheng, H., Edwards, R.L., Kong, X., Shao, X., Chen, S., Wu, J., Jiang, X.,
1147 Wang, X., An, Z., 2008. Millennial- and orbital-scale changes in the East Asian
1148 monsoon over the past 224,000 years. Nature 451, 1090-1093.
1149 doi:10.1038/nature06692.

1150 Warren, B.A., 1983. Why is no deep water formed in the North Pacific? J. Mar. Res.
1151 41, 327-347.

1152 Waser, N.A.D., Harrison, P.J., Nielsen, B., Calvert, S.E., Turpin, D.H., 1998. Nitrogen
1153 isotope fractionation during the uptake and assimilation of nitrate, nitrite, ammonium,
1154 and urea by a marine diatom. Limnol. Oceanogr. 43 (2), 215-224.

1155 Yamamoto, M., Tanaka, N., Tsunogai, S., 2001. Okhotsk Sea intermediate water
1156 formation deduced from oxygen isotope systematics. J. Geophys. Res. 106 (C12),
1157 31075-31084.

1158 Yasuda, I., 1997. The origin of the North Pacific Intermediate Water. J. Geophys.
1159 Res. 102 (C1), 893-909.

1160 Yokoyama, Y., Esat, T.M., 2011. Global climate and sea level: Enduring variability
1161 and rapid fluctuations over the past 150,000 years. Oceanography 24 (2), 54-69.
1162 doi:10.5670/oceanog.2011.27.

1163 Zhang, J., Woodgate, R., Moritz, R., 2010. Sea ice response to atmospheric and
1164 oceanic forcing in the Bering Sea. J. Phys. Oceanogr. 40 (8), 1729-1747.
1165 doi:10.1175/2010JPO4323.1.

1166 Zheng, Y., van Geen, A., Anderson, R.F., Gardner, J.V., Dean, W.E., 2000.
1167 Intensification of the northeast Pacific oxygen minimum zone during the Bölling-
1168 Alleröd warm period. Paleoceanography 15 (5), 528-536.

1169 Ziegler, M., Jilbert, T., de Lange, G.J., Lourens, L.J., and Reichart, G.-J. (2008),
1170 Bromine counts from XRF scanning as an estimate of the marine organic carbon
1171 content of sediment cores, Geochemistry Geophysics Geosystems, 9(5), Q05009,
1172 doi:10.1029/2007GC001932.

1173

1174 **Figure captions**

1175 Figure 1: (a) Surface circulation pattern (red arrows; after Tomczak and Godfrey,
1176 1994; Stabeno et al., 1999) and bathymetry of the subarctic North Pacific realm. The
1177 red dot marks the location of sediment core SO201-2-85KL studied here. Published
1178 reference records are marked by yellow dots. Bering Sea: MR06-04-PC24A (Kim et
1179 al., 2011), KH99-3-BOW-8A (Horikawa et al., 2010), HLY02-02-17JPC (Brunelle et
1180 al., 2007, 2010), IODP Site U1340 (Schlung et al., 2013). Okhotsk Sea: YK0712-
1181 GC9A (Khim et al., 2012), GGC27 (Brunelle et al., 2010). NW Pacific: MD01-2416

1182 (Galbraith et al., 2008a), PC13 (Brunelle et al., 2010). NE Pacific: ODP Site 887

1183 (Galbraith et al., 2008a). The modern average maximum sea-ice extent during March

1184 is indicated by the dashed black line (after Niebauer et al., 1999; Zhang et al., 2010;

1185 IRI/LDEO Climate Data Library, http://iridl.ldeo.columbia.edu/). Surface currents:

1186 ANSC = Aleutian North Slope Current, BSC = Bering Slope Current, EKC = East

1187 Kamchatka Current, ESC = East Sakhalin Current, NOC = North Okhotsk Current,

1188 SC = Soya Current, WKC = West Kamchatka Current. Straits: bus = Bussol Strait,

1189 kss = Kruzenshtern Strait, ks = Kamchatka Strait, ns = Near Strait, bp = Buldir Pass,

1190 as = Amchitka Strait, ap = Amukta Pass, up = Unimak Pass, bs = Bering Strait. (b)

1191 Surface nitrate concentration during modern summer (July-September; in µmol l^{-1})

1192 from World Ocean Atlas 2009 data (Garcia et al., 2010). Maps produced with "Ocean

1193 Data View" (Schlitzer, 2013).

1194

1195 Figure 2: (a) Relationship between concentrations of total nitrogen (TN) and total

1196 organic carbon (TOC) in samples from core SO201-2-85KL. For the calculation of

1197 molar N/C ratios a linear regression between TOC and TN was used to assess the

1198 fraction of inorganic nitrogen, represented by the intercept of the regression line at

1199 TOC = 0. (b) Comparison with $\delta^{15}N_{bulk}$ indicates that there is only a weak linear

1200 relationship between the isotopic signal and TN concentrations ($R^2 = 0.24$; $p < 10^{-4}$).

1201

1202 Figure 3: Linear sedimentation rate (LSR) vs. bulk accumulation rate (AR Bulk), and

1203 comparison of concentration and accumulation rate (AR) records of Siliciclastics,

1204 $CaCO_3$, TOC and opal for core SO201-2-85KL.

1205

1206 Figure 4: Records reflecting changes in export production and terrigenous matter

1207 supply in core SO201-2-85KL over the past 180 kyr in comparison with Northern

1208 Hemisphere summer (July-September) insolation at 65°N (after Laskar et al., 2004).

1209 Logging data (underlying grey lines), %Siliciclastics, as well as $CaCO_3$ and opal

1210 concentrations are from Max et al. (2012) and Riethdorf et al. (2013a, 2013b). Note

1211 inverted axes of %Siliciclastics and XRF Al count rates. The $\delta^{18}O$ records from the

1212 NGRIP ice core in Greenland (NGRIP members, 2004; GICC05 timescale,
1213 Rasmussen et al., 2006) and from the Sanbao stalagmites in China (Wang et al.,
1214 2008) are shown for reference. Greenland interstadials (GI) are highlighted by pale
1215 red vertical bars. Boundaries of Marine Isotope Stages (MIS) after Lisiecki and
1216 Raymo (2005).

1217

1218 Figure 5: Comparison of $\delta^{13}C_{bulk}$ with (a) molar N/C ratios and (b) $\delta^{15}N_{bulk}$ for core
1219 SO201-2-85KL. Samples from warm intervals (MIS 5.5, 5.3, 5.1, 3, and 1) are
1220 marked by red dots, while blue dots mark those from cold intervals (MIS 6, 5.4, 5.2,
1221 4, 2), and green triangles indicate Holocene (<11.7 ka BP) samples. Grey-shaded
1222 boxes represent geochemical provenances (after literature data; see text for
1223 references). The dashed lines indicate the applied linear mixing model for the
1224 estimation of m_{terr}. Marine phytoplankton and vascular plant detritus (VPD) are
1225 considered as potential marine and terrestrial organic matter sources (endmembers).

1226

1227 Figure 6: Proxy records from core SO201-2-85KL in comparison with published
1228 reference records covering the last 180 kyr: (a) Northern Hemisphere summer (65°;
1229 July-September) insolation after Laskar et al. (2004), (b) sedimentary $\delta^{13}C$ and (c)
1230 molar N/C ratios used to estimate the fraction of terrestrial organic matter (m_{terr};
1231 respective axes apply to those of $\delta^{13}C$ and N/C), (d) color b* assumed to reflect
1232 export production, (e) $\delta^{15}N_{bulk}$ reflecting surface nitrate utilization, (f) neodymium
1233 isotope ratios from core KH99-3-BOW-8A (Horikawa et al., 2010; cf. Figure 1)
1234 considered to approximate intermediate water formation, and (g) relative sea-level
1235 (Waelbroeck et al., 2002) normalized to the sill depth (~50 m; dashed line) of the
1236 Bering Strait (ΔRSL). MIS boundaries after Lisiecki and Raymo (2005), GI
1237 highlighted by pale red vertical bars.

1238

1239 Figure 7: Comparison of $\delta^{15}N_{bulk}$ from SO201-2-85KL (black line) with other
1240 sedimentary (solid lines) and diatom-bound (dotted lines) $\delta^{15}N$ records from the

1241   subarctic North Pacific and its marginal seas (cf. Figure 1). The timing of Greenland

1242   interstadials and MIS boundaries are indicated.