1   **Late Quaternary depositional and glacial history of the Arliss Plateau off the East**

2   **Siberian margin in the western Arctic Ocean**

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Abstract   Sedimentary stratigraphy and facies analysis along with seismostratigraphic and multibeam bathymetry data are used to reconstruct the last glacial impact on the Arliss Plateau (AP) and attendant sedimentation in the adjacent Chukchi Basin (CB) in the western Arctic Ocean off the East Siberian margin. Sediment core ARA02B/16B-GC from the AP lower slope captures glacier-related depositional history during the last estimated ca. 100 ka (Marine Isotope Stage (MIS) 1 to 5c) based on regional lithostratigraphic correlation. The sedimentary record shows distinguishable interglacial (interstadial) and glacial (stadial) patterns. The identified sedimentary facies reflect several modes of glaciogenic deposition by drifting icebergs, suspension settling from turbid meltwater plumes and/or detached underflows, and turbidity currents. Based on strong seismic reflectors related to lithological boundaries, a downslope subbottom profile from AP to CB is divided into seismostratigraphic units (SSU) 1 and 2 corresponding in the core record to MIS 1-3 and MIS 3-5c, respectively. An acoustically transparent lens within SSU 2 correlates on the upper slope to debris lobes downslope from the AP top covered by megascale glacial lineations. This geomorphic/sedimentary pattern indicates a glacial erosional impact on the AP and proglacial deposition of eroded sediments on the slope and in the basin. Based on the developed sediment stratigraphy and facies analysis, the last debris lobe horizon was deposited in glacial/deglacial environments during late MIS 4 to early MIS 3. The absence of similar glaciogenic debris lobes within SSU 1 indicates no direct glacial impact on the AP during the Last Glacial Maximum (MIS 2). These results suggest that the last glacial erosion of the AP occurred during or immediately after MIS 4, possibly related to major glaciation in northern Siberia at ~50-70 ka.

1. Introduction

Over last two decades, multiple geophysical investigations along with sedimentary records revealed that the Arctic East Siberian and Chukchi continental margins have been repeatedly glaciated during the Pleistocene (Jakobsson et al., 2008, 2014; Polyak et al., 2007; Niessen et al., 2013; Dove et al., 2014; O'Regan et al., 2017). In particular, streamlined glacial lineations and related diamicton deposits identified by swath seafloor and subbottom profiling on the East Siberian margin, including the adjacent part of the Mendeleev Ridge known as the Arliss Plateau (AP) (Fig. 1C), indicate the repeated development of up to 1 km-thick marine ice sheets/ice shelves (Niessen et al., 2013). However, the age framework for these glacial events, and thus, their relation to a broader paleoclimatic context, is not well understood. Recently, Schreck et al. (2018) provided lithostratigraphic insight on glacial impacts on the East Siberian margin in the Late Pleistocene, concluding that the last major regional ice grounding occurred during MIS 4 to 3 (60-70 ka). This conclusion differs from the notion that this event was coeval with the deepest glacial erosion on the Lomonosov Ridge in the central Arctic Ocean (Jakobsson et al., 2016; O'Regan et al., 2017), which has been constrained to MIS 6 (Jakobsson et al., 2001).

To better constrain the timing of ice grounding events and their spatial pattern, it is important to link the geophysical records and sediment core stratigraphy. This task requires analyzing representative sedimentary records accompanied by sub-bottom seismic profiles and detailed multibeam bathymetry from the areas affected by glaciations. However, most areas of continental margins, as well as bathymetric highs on the oceanic ridges and plateaus impacted by glacial erosion or iceberg turbation, are unsuitable for recovering well preserved, stratigraphically coherent records (Polyak et al., 2007; Jakobsson et al., 2008; Dove et al., 2014). In comparison, the adjacent slopes and basins may hold more expanded, better-preserved stratigraphies, provided their depositional context, which is sufficiently constrained by geophysical data. In particular, the slope of the AP and the adjacent Chukchi Basin (CB) was chosen for this study as a relatively well surveyed area of deposition for glaciogenic material originating from the East Siberian continental margin (Niessen et al., 2013; Jakobsson et al., 2014; Schreck et al., 2018). We are using a sediment core collected on a CHIRP sub-bottom profile (SBP) from the AP top to the CB (Fig. 1) to reconstruct the depositional history of the East Siberian margin during and since the last glacial grounding event.

2. Regional setting

2.1. Seafloor morphology and sedimentary cover

The CB is bounded by the Mendeleev Ridge/AP and the Chukchi Plateau and shelf and is connected to the north with the Canada Basin (Fig. 1). The CB covers ~43,000 km² (Jakobsson et al., 2003) and has a relatively flat and smooth seafloor ranging from 1,800 m to 2,300 m water depth and comprised of well-stratified deposits (Dove et al., 2014). The tops of both the Arliss and the Chukchi plateaus are covered by streamlined morphological features interpreted as megascale glacial lineations (MSGL), complicated in some areas with transverse ridges (Niessen et al., 2013; Dove et al., 2014; Jakobsson et al., 2014, 2016). The outer parts of the East Siberian and Chukchi margins at water depths to 700 m feature extensive till wedges and recessional moraines related to repeated glacial advances and retreats (Niessen et al., 2013; Dove et al., 2014; Jakobsson et al., 2014). Further downslope these margins as well as adjacent borderlands, including the AP, hold large, lens-shaped debris lobes (Fig. 2), similar to seismostratigraphic features observed in the glacial trough mouth fans in front of large ice streams (Batchelor and Dowdeswell, 2014; Jakobsson et al., 2014; O'Regan et al., 2017).

2.2. Sediment stratigraphy in the western Arctic Ocean

The Quaternary western Arctic sediments are generally characterized by the cyclical pattern of dark brownish and olive/greyish units (Phillips and Grantz, 1997; Polyak et al., 2004, 2009; Adler et al., 2009; Stein et al., 2010; Schreck et al., 2018). This color contrast was shown to be controlled by the content of manganese (Mn) oxy(hydr)oxides, with higher values in brown units attributed to Mn inputs from the continental margins at higher sea levels during interglacial/major interstadial periods (Jakobsson et al., 2000; Polyak et al., 2004, 2009; März et al., 2011; Löwemark et al., 2014; Ye et al., 2019). While Mn-enriched intervals can be affected by diagenetic remobilization and redistribution of this redox-sensitive element, their stratigraphic pattern is consistent throughout the central Arctic Ocean

101 (März et al., 2011; Löwemark et al., 2014; Schreck et al., 2018). Their correlation is corroborated in

102 multiple cores by the distribution of other lithostratigraphic parameters, such as sediment density, ice-

103 rafted debris (IRD), and biogenic proxies. The brown, Mn-enriched units (layers), interpreted as

104 interglacial/major interstadial sedimentary intervals, are typically characterized also by low to moderate

105 amounts of coarse IRD, intense bioturbation, and relatively abundant microfossils (Polyak et al., 2004,

106 2009; Adler et al., 2009; Stein et al., 2010; Löwemark et al., 2012). In comparison, olive/greyish, Mn-

107 depleted units, related to glacial periods, are composed of fossil-poor sediments with strongly variable

108 IRD amounts (Polyak et al., 2004, 2009; Adler et al., 2009; Stein et al., 2010). Fine-grained and IRD-

109 rich lithological types of grey sediments have been attributed to glacial maxima and deglacial

110 depositional environments, respectively. At sites proximal to glaciated areas, deglacial sediments may

111 have expanded thicknesses indicative of high depositional rates (Polyak et al., 2007; Wang et al., 2013;

112 Schreck et al., 2018). Some of the IRD peaks have characteristic sediment compositions used as

113 provenance indicators. In particular, detrital carbonates (mostly dolomites) are related to the Paleozoic

114 platform of the Canadian Arctic eroded by the Laurentide Ice Sheet (LIS) (Bischof and Darby, 1997;

115 England et al., 2009; Polyak et al., 2009; Stein et al., 2010; Bazhenova et al., 2017; Dong et al., 2017).

116 Peaks of this material forming characteristic pinkish-white (PW) layers in sediment cores (Clark et al.,

117 1980) can be used to track the distribution of icebergs in the Arctic Ocean during the LIS collapse events.

118

3. Material and methods

3.1. Material

121 We use a 378-cm-long, 12-cm in diameter gravity sediment core ARA02B/16B-GC (hereafter

122 16B-GC) and high frequency (2.5-6.5 kHz) subbottom CHIRP profiler data collected on the AP slope

123 (Fig. 1) during the 2nd RV *Araon* Arctic Expedition (ARA02B) in 2011. Multibeam bathymetry

124 mapping of the AP top (Fig. 1C) was subsequently conducted in 2012 (ARA03B). The 16B-GC coring

125 site (76° 24' N, 175° 58'W, 1841 meter water depth (mwd)) was selected based on the AWI Parasound

126 data acquired previously on the 2008 RV *Polarstern* Arctic expedition ARK-24/3 (Jokat, 2008). The

127   site is located ~1 km downslope from the foot of glaciogenic debris lobes and features relatively flat

128   and smooth seafloor with acoustically well stratified subbottom reflectors (Fig. 2). A 38-cm-long box

129   core 16B-BC was taken at the same site to recover the surficial sediment. Based on the comparison with

130   16B-BC, about 10 cm of top sediment is lost in 16B-GC, including the surficial brown lithological unit

131   B1, which was spliced for a composite record of 16B-BC (Figs. 3, 4).

132

133   **3.2. Methods**

134   After a visual description of the split core, its line-scan images and elemental geochemistry were

135   acquired with AVAATECH XRF-core scanner at Korea Institute of Geoscience and Mineral Resources

136   (KIGAM). Line-scanning was conducted using a Jai CV L107 camera with red-green-blue (RGB)

137   channels. L* (lightness) and a* (red-green color space) data were used for correlation with previously

138   published sediment records (Schreck et al. 2018). Sediment elemental composition was measured at 5-

139   mm intervals using X-ray fluorescence (XRF) AVAATECH scanner settings reported in Schreck et al.

140   (2018). Mn and calcium (Ca) XRF data used in lithostratigraphic correlation were normalized to

141   aluminum (Al) to eliminate the potential dilution effect of background sedimentation (e.g., März et al.,

142   2011). We note, however, that Mn and Ca variations in Arctic sediments are largely independent of the

143   background lithogenic elements as shown in prior studies (e.g., Löwemark et al., 2014; Dong et al.,

144   2017)

145   X-radiographs were performed for detailed sediment structure on 1 x 10 x 30 cm slabs. X-ray

146   images were taken with SOFTEX m-100w digital x-ray scanner (EZ-320) and instantly processed using

147   iX-Pect X-Ray imaging software. Darker objects in X-radiographs presented in this study display dense

148   material, such as lithic fragments. Coarse debris >2 mm in diameter was counted at every 1 cm,

149   according to Grobe (1987). Samples for sediment texture were taken at 10 cm intervals and treated with

150   35% H_2O_2 to decompose organic matter. Sand-sized and coarser grains were separated by wet sieving

151   at 63 μm and dry sieving at 2 mm standard sieves. The grain size of <63-μm sediment was measured

152   with a Micromeritics Sedigraph 5000D.

153   The multibeam bathymetric survey was performed using a hull-mounted EM122 echosounder.

154   Acquired bathymetric data were processed onboard using a specialized software CARIS. Sub-bottom

155   seismic-reflection data were acquired with the CHIRP SBP-120 profiler with the vertical resolution on

156   the 10 cm scale in acoustically stratified sediments. The SBP data were logged in the TOPAS raw

157   format, which was converted to an SEG-Y format and allowed for post-processing with standard

158   software packages. The estimated sound velocity of 1500 m/s was used to convert the signal travel time

159   into sediment depth for comparison with core lithostratigraphy.

160

161   **4. Results**

162   **4.1. Lithostratigraphy**

163   Core 16B is composed of sandy to fine-grained mud displaying a general cyclic depositional

164   pattern characterized by an alternation of brown to dark brown and yellow to olive greyish sedimentary

165   units (Figs. 3, 4). The concentration of coarse IRD (>2 mm) increases at lithological boundaries and

166   within some brown units, whereas IRD within yellow to olive greyish units is typically rare.

167   A total of five Mn-rich brown units, B1 to B5, are identified in the recovered stratigraphy based

168   on sediment color, and Mn/Al ratio (Fig. 3) along with sedimentary structure and texture (Fig. 4),

169   including the upper unit B1 spliced from the box core 16B-BC. The brown units show a characteristic

170   low-L* (darker) and high-a* (more reddish) color pattern with a relatively symmetrical distribution

171   across each unit (Fig. 3). High Mn/Al ratios in surficial sediments may be partially due to high water

172   content (Thallinghii et al., 2007), but that does not affect the stratigraphic distribution of Mn values,

173   which are high in unit B1. The brown units mostly consist of slightly sandy mud with scattered IRD

174   and have relatively sharp upper boundary and uneven, bioturbated (mottled) lower boundary (Fig. 4).

175   Yellowish to olive-grey units are generally thicker than brown units and are mostly composed of fine-

176   grained mud without coarse IRD. In particular, the second grey unit between B2 and B3 is the thickest

177   interval recovered (~150 cm) with a consistent lithology. Two PW carbonate layers, consisting of

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178 sandy-gravelly mud and expressed as relatively solid lenses and adjacent scattered speckles, are

179 identified above B2 and at the base of B3, corresponding to prominent Ca/Al peaks (Figs. 3, 4).

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181 **4.2. Sediment facies**

182   Sediment of core 16B was classified into five major facies based on sedimentary structure,

183 texture, and color (Fig. 4, Table 1): bioturbated sandy mud (Bsm), bioturbated mud (Bm), crudely

184 laminated or layered mud (CLm), thinly laminated mud/muddy sand (TLm/TLs), and homogeneous

185 mud (Hm).

186   Bioturbated sandy mud (Bsm) is mostly identified in brown units and PW carbonaceous layers

187 and is characterized by strong bioturbation overprinting primary sedimentary structures, sharp upper-

188 and uneven, mottled lower-boundaries (Fig. 4). This facies mainly consists of very poorly sorted (2 – 4

189 Φ) sandy mud with coarse particles randomly scattered within the fine matrix (Fig. 3). In comparison

190 with the brown units, Bsm in the PW layers contains higher sand content, up to 22% (Fig. 3), and more

191 coarse particles.

192   Bioturbated mud (Bm) generally occurs at the transitions between brown and olive-grey units

193 and is characteristic for the olive-grey units underlying B1, where Bm is interbedded with the crudely

194 laminated or layered mud facies (CLm), and below B3 (Fig. 4). Similar to Bsm, the Bm facies also has

195 a mottled structure but is composed of more fine-grained muds with gradual facies boundaries. The

196 sediment sorting index in this facies (1 – 1.5 Φ) is lower than in Bsm due to a lack of coarse particles

197 (Figs. 3, 4).

198   Crudely laminated or layered mud (CLm) is characterized by the alternation of planar to sub-

199 parallel silt-and clay-rich laminae or layers with weak bioturbation (Fig. 4). Individual laminae are

200 generally difficult to discern. Clayey laminae/layers are mostly thicker (4 – 12 mm) than their silt-rich

201 counterparts (1 – 3 mm). Vertical changes in lamina thickness and composition appear to be random,

202 and the sediment is overall weakly sorted (1 – 1.5 Φ). Facies CLm is predominant within the second

203 olive-grey unit bounded by B2 and B3 and is also interbedded with the Bm facies in the first olive-grey

204 unit.

205   Thinly laminated mud/muddy sand facies (TLm/TLs) can be subdivided into silt-rich mud (TLm)

206 and muddy sand (TLs). This distinct facies has a laterally continuous lamination (a few millimeters

207 thick), typically identified as couplets with overlying homogeneous mud facies Hm (Fig 4). TLm/TLs

208 laminae show sharp lower and gradational upper boundaries, and relatively regular thicknesses (Fig. 3).

209 TLm/TLs facies occur episodically in the second and third olive-grey intervals, between B2 – B3 and

210 B3 – B4, respectively (Figs. 3, 4).

211   Homogeneous mud facies (Hm) consists of fine-grained muddy sediment characterized by an

212 overall absence of primary structures, bioturbation, or any grading. This facies is mostly interbedded

213 with the thinly laminated facies TLm/TLs on millimeter scales as couplets. Independently occurring

214 Hm facies (3 cm thick) is identified only in the upper part of the third olive-grey interval below B3 (Fig.

215 4).

216

217 **4.3. Geomorphology and seismostratigraphy of the AP basinward slope**

218   The seafloor of the AP top shows several sets of streamlined morphological features (Fig. 1C)

219 interpreted in prior studies as megascale glacial lineations (Niessen et al., 2013; Jakobsson et al., 2016).

220 The youngest set of lineations of SSW-NNE orientation is traced to water depths of ~950 m in the

221 southern part of AP, compared to somewhat shallower depths of 850 m in the northern part. The lineated

222 surface is covered by ~3-m thick, acoustically stratified sediment (Niessen et al., 2013).

223   The subbottom data from the basin-ward middle to the lower slope of the AP (~1600-1800 mwd)

224 show stacked series of debris lobes of ~10-20 km long and up to 25 m thick (Fig. 2A). The debris

225 deposits are covered by acoustically stratified sediments (Figs. 2B, C). Three distinct acoustic reflectors

226 (R1 – R3) were identified by lateral continuity and strong reflectivity. The reflectors divide the upper

227 part of the sediment cover into two major seismostratigraphic units, SSU 1 and 2.

228   The surficial unit SSU 1 bounded by reflectors R1 (seafloor) and R2 is laterally continuous and

229   has bedding parallel to R2 (Figs 2B and C). SSU 1 is characterized by transparent to slightly fuzzy sub-

230   bottom echoes, with a laterally discontinuous, weak reflector also recognized within this unit (Fig. 2B).

231   SSU 1 thins out from the core 16B site towards the steep middle slope (Figs. 2B and C) and is

232   indistinguishable further up-slope near the AP top.

233   As seen on records from the basin and lower slope (Fig. 2B), the underlying unit SSU 2, bounded

234   by R2 and R3, is generally characterized by parallel lamination with variable amplitudes and transparent

235   acoustic signature in the middle part of the unit. Further up-slope, two coeval debris lobes are

236   recognized in the lower part of SSU 2, covered by acoustically stratified sediment comprised of the

237   remaining part of SSU 2 and overlying SSU 1 (Figs. 2B, C). Two thin reflectors with low to moderate

238   reflectivity are identified in the lowermost part of SSU 2 in the basin, but cannot be found near the

239   debris lobes on the upper slope (Fig. 2B).

240

241   **5. Discussion**

242   **5.1. Age model**

243   As core 16B-GC lacks calcareous material for ^{14}C dating, its age model is constrained by

244   lithostratigraphic correlation with the earlier investigated core ARA03B/28B-GC from the AP top

245   (Schreck et al., 2018) and core HLY0503-8JPC from the foot of the Mendeleev Ridge ~350 km north

246   (Adler et al., 2009). The latter core record has the best regional age control constrained by multiple [14]C

247   datings in the upper part of the stratigraphy (MIS 1 to 3, units B1-B3), ^{14}C-calibrated amino-acid

248   racemization (AAR) rates extending to MIS 5 (Kaufman et al., 2008; Adler et al., 2009), and coccolith

249   occurrences (Backman et al., 2009). The correlation is primarily based on sediment color and

250   distribution of Mn and Ca content showing a regionally consistent pattern (Fig. 5) (Polyak et al., 2009;

251   Stein et al., 2010; Schreck et al., 2018). According to ^{14}C dating, brown, Mn-rich units B1 and B2,

252   including the detrital carbonate layer W3, represent MIS 1 (~ 9 ka) and MIS 3 (34-45 ka)

253   interglacial/interstadial environments, respectively (Fig. 5; Table 2). The intermittent grey unit G1 thus

254   corresponds to the last glaciation/deglaciation (MIS 2 to 1) and the youngest, pre-glacial part of MIS 3.

255   Brown units B3 to B5 recovered by core 16B-GC have been attributed in the reference cores to

256   pre-MIS 3 interstadials ranging to MIS 5c, with some variance. Based on coccolith distribution in

257   HLY0503-8JPC, Backman et al. (2009) suggested that B5 represents MIS 5a, while Adler et al. (2009)

258   assigned its age to MIS 5c, consistent with AAR ages (Kaufman et al., 2008). This age model was also

259   used by Schreck et al. (2018), who pointed out that the coccolith stratigraphy cannot pinpoint a specific

260   time interval within MIS 5. In any case, core 16B-GC does not reach the prominent detrital carbonate

261   layer PW2 identified in multiple cores across the western Arctic Ocean, including HLY0503-8JPC (Fig.

262   5), and attributed to MIS 5d (e.g., Stein et al., 2010; Bazhenova et al., 2017; Dong et al., 2017). Overall,

263   we consider MIS 5c, with the approximate age of ca. 100 ka, as a reasonable estimate for the bottom of

264   16B-GC.

265   The developed lithostratigraphic framework indicates that a ~150-cm thick grey unit G2 pre-

266   dates the MIS 3 interstadial, likely being contained in the time interval from MIS 4 to early MIS 3. This

267   prominent lithostratigraphic unit can be traced in multiple cores from the Siberian margin to the Arctic

268   Ocean interior (e.g., Spielhagen et al., 2004; Polyak et al., 2009; Dong et al., 2017; Wang et al., 2018),

269   and has been proposed to correlate to the last pre-LGM glaciation in northern Eurasia with a glacial

270   maximum around 50-60 ka (Svendsen et al., 2004).

271   In comparison to this, regionally consistent age model, the B2 unit in a very similar nearby record

272   was assigned to MIS 5 based solely on the presence of rare coccoliths, including *E. huxleyi* (core

273   SWERUS-L2-13PC; https://bolin.su.se/data/swerus/physical.php; Fig. 1) (Jakobsson et al., 2016). The

274   occurrence of *E. huxleyi* clearly indicates that the sediments of this unit are younger than MIS 8/7

275   (Thiersten et al., 1977; Backman et al., 2009). However, due to an overall limited occurrence of

276   coccoliths, *E. huxleyi* presence may not be used to identify specific substages within MIS 5 as

277   attempted previously (e.g., Jakobsson et al., 2001; Sphielhagen et al., 2004) because this species can

278   potentially occur in other relatively warm intervals of Arctic sediments. In particular, MIS 3 is a very

279   pronounced interstadial unit in Arctic paleoceanographic records, as exemplified by a high content of

280 Mn and high productivity indicated by abundant planktic and benthic microfossils (Nørgaard-Pedersen

281 et al., 1998; Adler et al., 2009; Hanslik et al., 2010; Schreck et al., 2018). Some authors further inferred

282 reduced sea ice and enhanced the advection of Atlantic waters during this time (Cronin et al., 2012;

283 Poirier et al., 2012). We note that in another SWERUS core from the Siberian margin further west, *E.*

284 *huxleyi* was also found in B2 with a ^{14}C age of 33 ka just a little above this finding (O'Regan et al.,

285 2017).

286

287 **5.2. Core-seismic correlation**

288   Distinct acoustic reflectors on seismic records can be formed by vertical changes in acoustic

289 impedance controlled by sediment density and p-wave velocity. Deep-sea Arctic Ocean sediments

290 generally consist of relatively soft fine-grained muds, which are occasionally intercalated with denser,

291 coarse sediment layers mostly attributed to iceberg sedimentation during deglacial events (Clark et al.,

292 1980; Polyak et al., 2009; Stein et al., 2010; Dong et al., 2017; Schreck et al., 2018). In particular,

293 detrital carbonate IRD layers, such as W3 and PW2 captured by western Arctic Ocean sediments, are

294 characterized in cores from the study region by a remarkably high density and p-wave velocity

295 compared to the enclosing fine-grained muds (Matthiessen et al., 2010; Stein et al., 2010; Schreck et

296 al., 2018). These layers, marking the LIS iceberg discharge from the Canadian Arctic, thus have the

297 potential to form strong reflectors.

298   Comparison of the subbottom acoustic reflection data and core 16B record (Fig. 6) confirms that

299 the surficial reflector R1 represents the seafloor, the interface between bottom water and relatively dense

300 surficial sediment of unit B1. A fine-grained unit G1 characterized by a relatively constant density and

301 p-wave velocity in multiple records from the study region (Matthiessen et al., 2010; Stein et al., 2010;

302 Schreck et al., 2018) defines a mostly transparent echo characteristic of SSU 1. A strong reflector R2

303 probably indicates a pronounced lithological change from the IRD-rich interval, including the W3 layer

304 to the overlying fine-grained unit G1 (Fig. 6). A relatively strong reflector in the uppermost part of SSU

305 2, partially divided into two layers, appears to be associated with a relatively coarse unit B2 (Fig. 6). A

306 transparent echo signature similar to that in SSU 1 characterizes the upper part of SSU 2 corresponding

307 to the fine-grained, thickest unit G2. Two thin reflectors of low to moderate amplitude in the lower part

308 of SSU 2 appear to correspond to a relatively sandy unit B3, especially its detrital carbonate-rich base

309 layer, and an intermittent sandy layer within G3 (Fig. 6). While core 16B-GC does not extend to the

310 depth of R3, a likely candidate for this strong reflector is a prominent, IRD-rich, detrital carbonate layer

311 PW2 that underlies unit B6 and is barely reached by 16B-GC (Fig. 5). PW2 is characterized by a

312 significantly higher wet bulk density and p-wave velocity than other Upper Quaternary sediments in

313 the study region (Mattiessen et al., 2010; Schreck et al., 2018).

314

5.3. Glaciogenic debris lobes on the AP slope

316 The presence of debris lobes in the lower part of SSU 2 up-slope from core 16B (Fig. 2) is likely

317 related to glacial scouring on the AP crest. While debris flows can potentially originate from various

318 slope instability processes, the consistent occurrence of these lobes in the Arctic Ocean downslope from

319 continental margins and plateaus characterized by glacial erosion/deposition indicates their glaciogenic

320 origin. In particular, the CB features debris lobes on both the AP and Chukchi Plateau slopes (Niessen

321 et al., 2013; Dove et al., 2014 and reference therein), indicating glaciogenic inputs from both sides. The

322 mechanisms for debris flows could include sediment bulldozing on the tops by grounded ice and/or

323 rapid deposition of deglacial sediments, e.g., from glacial underflows (e.g., Powell, 1990; Batchelor

324 and Dowdeswell, 2015). The timing of debris lobe emplacement would thus correspond to glacial

325 advance or early deglaciation stages.

326 The SSW-NNE orientation and southward inclination of the youngest set of glacial lineations

327 on the AP top indicate that the eroding ice was sourced from the East Siberian margin (Niessen et al.,

328 2013). According to the age constraints developed for core 16B-GC and a correlative core 28B-GC

329 from the AP top (Fig. 5), this event occurred prior to the MIS 3 interstadial and was possibly related to

330 a major glaciation in northern Siberia dated to the time interval between ~70 and 50 ka (Svendsen et

331 al., 2004; Möller et al., 2015).

13

The upper seismostratigraphic unit SSU 1 bounded by the seafloor, and the W3 layer represents deposition during the time interval from late MIS 3 to the Holocene (Fig. 6). The absence of glaciogenic debris lobes within SSU 1 indicates that glacial erosion of the AP was unlikely during the LGM, as suggested in previous studies (Niessen et al., 2013; Schreck et al., 2018).

5.4. Glacier-induced depositional environments

Based on the developed age model, sedimentary unit G2 (MIS 4 to early MIS 3) that is characterized by laminated mud facies, has a relatively high sedimentation rate of ~6 cm/kyr as compared to more bioturbated units G3 and G4 (Fig. 4, Table 2). Laminated structures of glaciogenic sediments in the high-latitude oceans can be potentially formed under various depositional settings such as permanent sea ice, turbid meltwater plumes, distal fine-grained turbidity currents, and contour currents (Hesse et al., 1996; Darby et al., 1997; Kleiber et al., 2000; ÓCofaigh et al., 2003; Matthiessen et al., 2010). However, contour currents are unlikely to be strong in a semi-enclosed Arctic deep-sea basin, while sedimentation rates under perennial sea ice are too low to form a laminated sedimentary sequence. Couplets of TLm/TLs and Hm facies similar to those observed in 16B-GC (Fig. 4) have been interpreted in previous studies as deposition from underflow turbidity currents in combination with rapid settling from fine-grained suspension plumes in the attenuation stage of the carrying current (Piper, 1978; Chough, 1984; Yoon et al., 1991). The occurrence of TLm/TLs facies within G2 and G3 suggests that turbidities could be associated with the glaciogenic debris flows emanating from their front on a relatively steep AP slope (Fig. 2B). The development of CLm facies may reflect deposition from turbid meltwater plumes repeatedly discharged from ice-sheet termini (Hesse et al., 1996; Kleiber et al., 2000; ÓCofaigh et al., 2003).

Glacial erosion on glaciated continental margins plays an important role in delivering large amounts of sediments to the continental slope and the adjacent basins, and thus leading to the formation of debris flows (Vorren et al., 1998; Kleiber et al., 2000; Dowdeswell and Elverhøi, 2002; Jakobsson et al., 2008). Based on our seismostratigraphic and sediment core data, this scenario applies to the AP

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during MIS 4, when large volumes of eroded sediment may have been transported as proglacial under- and overflows into the adjacent deep basin. This rapid sediment deposition likely caused slope instability and thus triggered debris flows. Suspended fine sediments could have been detached from the slope and transported into the interior of the Arctic Ocean at the density boundaries between different water masses with subsequent deposition. Prolonged glacial erosion could have also formed repeatedly detached turbid layers, leading to the formation of crudely laminated mud (CLm) facies (Fig. 7A). As the grounded ice retreated, turbid meltwater underflow plumes were likely repeatedly discharged from the grounding line, thereby depositing CLm facies (Hesse et al., 1996; ÓCofaigh et al., 2003) (Fig. 7B). These glacier-induced processes could lead to the deposition of fine-grained, fully laminated sediments with fairly high sedimentation rates (unit G2) during the MIS 4 deglaciation that may have extended to early MIS 3.

Bioturbated sediments of B2 and the overlying W3 layer are interpreted to represent hemipelagic depositional environments with drifting icebergs during a relatively warm period of MIS 3 (Fig. 7C). In general, Mn-enriched, brown sediments are believed to reflect elevated sea levels and/or reduced sea-ice conditions, under which sediment from the Siberian shelves could be transported to the central Arctic basins (Jakobsson et al., 2000; Polyak et al., 2004, 2009; Stein et al., 2010; Löwemark et al., 2014; Ye et al., 2019). Overall, sedimentary properties of G2 and overlying facies indicate depositional environments affected by the advance and retreat of a marine-based ice sheet grounded on the AP top.

The absence of glaciogenic debris lobe within SSU 1 indicates that the AP has not encountered glacial erosion during MIS 2, consistent with previous studies from the Arctic Ocean off the Siberian margin (Niessen et al., 2013; Schreck et al., 2018) as well as terrestrial studies from adjacent islands and the East Siberian mainland (Romanovskii et al., 2004; Gualtieri et al., 2005). Nevertheless, the presence of CLm facies within unit G1, most of which is attributed to MIS 2 (Figs. 3, 4), reflects glacier-induced depositional processes, such as the suspension settling of turbid meltwater plumes and detached turbid layers (Figs 7D, E). This lithology indicates the influence of a limited glaciation in the study region, although its location and distribution pattern (e.g., grounded ice vs. ice shelf) needs to be further

384 investigated. There is no evidence for the LGM glaciation at the East Siberian margin, but data from

385 the eastern Chukchi shelf margin show glaciogenic bedforms indicative of grounded ice, age-

386 constrained to MIS 2 (Polyak et al., 2007). These bedforms are restricted to water depths of less than

387 ~450 m, but we cannot exclude that this grounded ice extended over the Chukchi shelf to its western

388 margin bordering the Chukchi Basin.

389 The predominance of the bioturbated sandy mud (Bsm) facies in the uppermost unit B1 reflects

390 hemipelagic deposition with sea-ice and/or iceberg debris rafting during much of the Holocene (Fig.

391 7F). Intense bioturbation indicates a relatively elevated export of primary production to the seafloor,

392 and thus moderate sea-ice conditions, and/or enhanced oxygenation of surface sediments.

393

394 **6. Conclusions**

395 A combination of litho- and seismostratigraphy on the basinward slope of the Arliss Plateau

396 depicts a glacier-induced depositional history in front of the East Siberian continental margin since

397 estimated MIS 5c (~100 ka). The age constraints were obtained by correlation of the studied core

398 ARA02B/16B-GC with earlier developed stratigraphies (Adler et al., 2009; Schreck et al., 2018). Based

399 on this age model and the core-seismic correlation, deposition of the youngest glaciogenic debris lobes

400 on the slope, indicative of the last glacial erosion of the AP top, occurred within seismostratigraphic

401 unit SSU 2 during MIS 4 to early MIS 3. This event may be correlative to a major glaciation in northern

402 Siberia that has been constrained to the time interval between ~70 and 50 ka (Svendsen et al., 2004;

403 Möller et al., 2015). Laminated sedimentary facies and sedimentation rates as high as ~6 cm/kyr in the

404 respective core interval (lithological unit G2) reflect glacier-induced sediment deposition. This

405 enhanced sedimentation probably resulted from detached turbid layers, turbidity currents, and

406 meltwater discharge pulses during glaciation/deglaciation.

407 The absence of glaciogenic debris lobes within the upper seismostratigraphic unit SSU 1 (late

408 MIS 3 to 1) indicates that glacial erosion unlikely impacted the AP during the LGM, consistent with

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previous inferences (Niessen et al., 2013; Schreck et al., 2018). Nevertheless, the presence of laminated facies within the corresponding core interval G1 may indicate a limited glaciation at a nearby continental margin. Further investigation by geophysical seafloor surveys verified by sediment core records is needed to better understand the glacial history of the western Arctic Ocean.

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570   **Figure and Table captions**

571   **Figure 1** (A) Index physiographic map of the study area showing the location of cores ARA02B/16B-

572   GC (this study), ARA03B/28B-GC (Schreck et al., 2018) and HLY0503-8JPC (Adler et al., 2009). MR,

573   Mendeleev Ridge; AP, Arliss Plateau; CB, Chukchi Basin; CP, Chukchi Plateau; NR, Northwind Ridge.

574   Solid line tracks sub-bottom profiler data. Bathymetric contour interval is 250 m. (B) Broader

575   physiographic/paleogeographic context with inferred ice-sheet maxima shown by semi-transparent

576   white fill (Svendsen et al., 2004; Niessen et al., 2013; Stokes et al., 2016). (C) Multibeam bathymetry

577   of the Arliss Plateau showing several generations of Mega-Scale Glacial Lineations. Bathymetric

578   contour interval is 100 m.

579   **Figure 2** Sub-bottom profiler from Arliss Plateau to Chukchi Basin across the site of core

580   ARA02B/16B-GC (red vertical bar in panel B). Seismostratigraphic units SSU 1 and 2 are bounded by

581   distinct reflectors R1 to 3. The uppermost glaciogenic debris lobes on the slope are correlated with the

582   lower part of SSU 2. Black box shows the position of Fig. 6.

583   **Figure 3** Down-core distribution of sediment lightness (L*), Mn/Al and Ca/Al ratios, IRD numbers,

584   and grain size in core ARA02B/16B-GC. Top 10 cm are estimated to be missing due to overpenetration;

585   XRF data for this interval (blue-colored solid lines) are spliced from the box core 16B-BC. High Mn/Al

586   and Ca/Al ratios in surficial sediment may be partially due to high water content (Thallinghii et al.,

587   2007). Lithostratigraphic indices are shown in Fig. 3.

588   **Figure 4** Split-core photos, X-radiograph images, and sedimentary facies (see Table 1 for details).

589   Indices for lithological units (B1-B5 and G1-G4) and the detrital carbonate layer W3 are shown on the

590   core photo.

591   **Figure 5** Lithostratigraphic correlation of core ARA02B/16B-GC with ARA03B/28B-GC (Schreck et

592   al., 2018) and HLY0503-8JPC (Adler et al., 2009) based on sediment color and variations of Mn/Al

593   and Ca/Al. The tops of major brown units (B2-B5) are correlated by dotted lines; the bottom of core

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594   ARA03B/28B-GC is projected onto ARA02B/16B-GC by the dashed line. Estimated Marine Isotope

595   Stages and radiocarbon/AAR ages (Adler et al., 2009) are shown next to core HLY0503-8JPC.

596   **Figure 6** Correlation of the subbottom data with core ARA02B/16B-GC. Surficial reflector R1 is the

597   seafloor. Strong reflectors R2 and R3 correspond to prominent lithological changes at the boundaries

598   of fine-grained sediments and IRD-rich layers W3 and PW2. Estimated Marine Isotope Stages are

599   shown on the right.

600   **Figure 7** Schematic model of sedimentary processes on the Arliss Plateau and adjacent Chukchi Basin

601   during major depositional regimes from the MIS 4 glaciation to the Holocene.

602

603   **Table 1** Summary of sedimentary facies in core ARA02B/16B-GC. Black arrows indicate the intervals

604   of TLm, TLs, and Hm facies. See Fig. 4 for the stratigraphic distribution of sedimentary facies.

605   **Table 2** Estimated duration and sedimentation rates of lithostratigraphic units in core ARA02B/16B-

606   GC based on the age model developed from lithostratigraphic correlations (Fig. 5).

Late Quaternary depositional and glacial history of the Arliss Plateau off the East

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Siberian margin in the western Arctic Ocean

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27 **Abstract**   Sedimentary stratigraphy and facies analysis along with seismostratigraphic and

28 multibeam bathymetry data are used to reconstruct the last glacial impact on the Arliss Plateau (AP)

29 and attendant sedimentation in the adjacent Chukchi Basin (CB) in the western Arctic Ocean off the

30 East Siberian margin. Sediment core ARA02B/16B-GC from the AP lower slope captures glacier-

31 related depositional history during the last estimated ca. 100 ka (Marine Isotope Stage (MIS) 1 to 5c)

32 based on regional lithostratigraphic correlation. The sedimentary record shows distinguishable

33 interglacial (interstadial) and glacial (stadial) patterns. The identified sedimentary facies reflect several

34 modes of glaciogenic deposition by drifting icebergs, suspension settling from turbid meltwater plumes

35 and/or detached underflows, and turbidity currents. Based on strong seismic reflectors related to

36 lithological boundaries, a downslope subbottom profile from AP to CB is divided into

37 seismostratigraphic units (SSU) 1 and 2 corresponding in the core record to MIS 1-3 and MIS 3-5c,

38 respectively. An acoustically transparent lens within SSU 2 correlates on the upper slope to debris lobes

39 downslope from the AP top covered by megascale glacial lineations. This geomorphic/sedimentary

40 pattern indicates a glacial erosional impact on the AP and proglacial deposition of eroded sediments on

41 the slope and in the basin. Based on the developed sediment stratigraphy and facies analysis, the last

42 debris lobe horizon was deposited in glacial/deglacial environments during late MIS 4 to early MIS 3.

43 The absence of similar glaciogenic debris lobes within SSU 1 indicates no direct glacial impact on the

44 AP during the Last Glacial Maximum (MIS 2). These results suggest that the last glacial erosion of the

45 AP occurred during or immediately after MIS 4, possibly related to major glaciation in northern Siberia

46 at ~50-70 ka.

1. Introduction

Over last two decades, multiple geophysical investigations along with sedimentary records revealed that the Arctic East Siberian and Chukchi continental margins have been repeatedly glaciated during the Pleistocene (Jakobsson et al., 2008, 2014; Polyak et al., 2007; Niessen et al., 2013; Dove et al., 2014; O'Regan et al., 2017). In particular, streamlined glacial lineations and related diamicton deposits identified by swath seafloor and subbottom profiling on the East Siberian margin, including the adjacent part of the Mendeleev Ridge known as the Arliss Plateau (AP) (Fig. 1C), indicate the repeated development of up to 1 km-thick marine ice sheets/ice shelves (Niessen et al., 2013). However, the age framework for these glacial events, and thus, their relation to a broader paleoclimatic context, is not well understood. Recently, Schreck et al. (2018) provided lithostratigraphic insight on glacial impacts on the East Siberian margin in the Late Pleistocene, concluding that the last major regional ice grounding occurred during MIS 4 to 3 (60-70 ka). This conclusion differs from the notion that this event was coeval with the deepest glacial erosion on the Lomonosov Ridge in the central Arctic Ocean (Jakobsson et al., 2016; O'Regan et al., 2017), which has been constrained to MIS 6 (Jakobsson et al., 2001).

To better constrain the timing of ice grounding events and their spatial pattern, it is important to link the geophysical records and sediment core stratigraphy. This task requires analyzing representative sedimentary records accompanied by sub-bottom seismic profiles and detailed multibeam bathymetry from the areas affected by glaciations. However, most areas of continental margins, as well as bathymetric highs on the oceanic ridges and plateaus impacted by glacial erosion or iceberg turbation, are unsuitable for recovering well preserved, stratigraphically coherent records (Polyak et al., 2007; Jakobsson et al., 2008; Dove et al., 2014). In comparison, the adjacent slopes and basins may hold more expanded, better-preserved stratigraphies, provided their depositional context, which is sufficiently constrained by geophysical data. In particular, the slope of the AP and the adjacent Chukchi Basin (CB) was chosen for this study as a relatively well surveyed area of deposition for glaciogenic material originating from the East Siberian continental margin (Niessen et al., 2013; Jakobsson et al., 2014; Schreck et al., 2018). We are using a sediment core collected on a CHIRP sub-bottom profile (SBP) from the AP top to the CB (Fig. 1) to reconstruct the depositional history of the East Siberian margin during and since the last glacial grounding event.

2. Regional setting

2.1. Seafloor morphology and sedimentary cover

The CB is bounded by the Mendeleev Ridge/AP and the Chukchi Plateau and shelf and is connected to the north with the Canada Basin (Fig. 1). The CB covers ~43,000 km^2 (Jakobsson et al., 2003) and has a relatively flat and smooth seafloor ranging from 1,800 m to 2,300 m water depth and comprised of well-stratified deposits (Dove et al., 2014). The tops of both the Arliss and the Chukchi plateaus are covered by streamlined morphological features interpreted as megascale glacial lineations (MSGL), complicated in some areas with transverse ridges (Niessen et al., 2013; Dove et al., 2014; Jakobsson et al., 2014, 2016). The outer parts of the East Siberian and Chukchi margins at water depths to 700 m feature extensive till wedges and recessional moraines related to repeated glacial advances and retreats (Niessen et al., 2013; Dove et al., 2014; Jakobsson et al., 2014). Further downslope these margins as well as adjacent borderlands, including the AP, hold large, lens-shaped debris lobes (Fig. 2), similar to seismostratigraphic features observed in the glacial trough mouth fans in front of large ice streams (Batchelor and Dowdeswell, 2014; Jakobsson et al., 2014; O'Regan et al., 2017).

2.2. Sediment stratigraphy in the western Arctic Ocean

The Quaternary western Arctic sediments are generally characterized by the cyclical pattern of dark brownish and olive/greyish units (Phillips and Grantz, 1997; Polyak et al., 2004, 2009; Adler et al., 2009; Stein et al., 2010; Schreck et al., 2018). This color contrast was shown to be controlled by the content of manganese (Mn) oxy(hydr)oxides, with higher values in brown units attributed to Mn inputs from the continental margins at higher sea levels during interglacial/major interstadial periods (Jakobsson et al., 2000; Polyak et al., 2004, 2009; März et al., 2011; Löwemark et al., 2014; Ye et al., 2019). While Mn-enriched intervals can be affected by diagenetic remobilization and redistribution of this redox-sensitive element, their stratigraphic pattern is consistent throughout the central Arctic Ocean

101 (März et al., 2011; Löwemark et al., 2014; Schreck et al., 2018). Their correlation is corroborated in

102 multiple cores by the distribution of other lithostratigraphic parameters, such as sediment density, ice-

103 rafted debris (IRD), and biogenic proxies. The brown, Mn-enriched units (layers), interpreted as

104 interglacial/major interstadial sedimentary intervals, are typically characterized also by low to moderate

105 amounts of coarse IRD, intense bioturbation, and relatively abundant microfossils (Polyak et al., 2004,

106 2009; Adler et al., 2009; Stein et al., 2010; Löwemark et al., 2012). In comparison, olive/greyish, Mn-

107 depleted units, related to glacial periods, are composed of fossil-poor sediments with strongly variable

108 IRD amounts (Polyak et al., 2004, 2009; Adler et al., 2009; Stein et al., 2010). Fine-grained and IRD-

109 rich lithological types of grey sediments have been attributed to glacial maxima and deglacial

110 depositional environments, respectively. At sites proximal to glaciated areas, deglacial sediments may

111 have expanded thicknesses indicative of high depositional rates (Polyak et al., 2007; Wang et al., 2013;

112 Schreck et al., 2018). Some of the IRD peaks have characteristic sediment compositions used as

113 provenance indicators. In particular, detrital carbonates (mostly dolomites) are related to the Paleozoic

114 platform of the Canadian Arctic eroded by the Laurentide Ice Sheet (LIS) (Bischof and Darby, 1997;

115 England et al., 2009; Polyak et al., 2009; Stein et al., 2010; Bazhenova et al., 2017; Dong et al., 2017).

116 Peaks of this material forming characteristic pinkish-white (PW) layers in sediment cores (Clark et al.,

117 1980) can be used to track the distribution of icebergs in the Arctic Ocean during the LIS collapse events.

118

119 **3. Material and methods**

120 **3.1. Material**

121 We use a 378-cm-long, 12-cm in diameter gravity sediment core ARA02B/16B-GC (hereafter

122 16B-GC) and high frequency (2.5-6.5 kHz) subbottom CHIRP profiler data collected on the AP slope

123 (Fig. 1) during the 2nd RV *Araon* Arctic Expedition (ARA02B) in 2011. Multibeam bathymetry

124 mapping of the AP top (Fig. 1C) was subsequently conducted in 2012 (ARA03B). The 16B-GC coring

125 site (76° 24' N, 175° 58'W, 1841 meter water depth (mwd)) was selected based on the AWI Parasound

126 data acquired previously on the 2008 RV *Polarstern* Arctic expedition ARK-24/3 (Jokat, 2008). The

127 site is located ~1 km downslope from the foot of glaciogenic debris lobes and features relatively flat

128 and smooth seafloor with acoustically well stratified subbottom reflectors (Fig. 2). A 38-cm-long box

129 core 16B-BC was taken at the same site to recover the surficial sediment. Based on the comparison with

130 16B-BC, about 10 cm of top sediment is lost in 16B-GC, including the surficial brown lithological unit

131 B1, which was spliced for a composite record of 16B-BC (Figs. 3, 4).

132

133 **3.2. Methods**

134   After a visual description of the split core, its line-scan images and elemental geochemistry were

135 acquired with AVAATECH XRF-core scanner at Korea Institute of Geoscience and Mineral Resources

136 (KIGAM). Line-scanning was conducted using a Jai CV L107 camera with red-green-blue (RGB)

137 channels. L* (lightness) and a* (red-green color space) data were used for correlation with previously

138 published sediment records (Schreck et al. 2018). Sediment elemental composition was measured at 5-

139 mm intervals using X-ray fluorescence (XRF) AVAATECH scanner settings reported in Schreck et al.

140 (2018). Mn and calcium (Ca) XRF data used in lithostratigraphic correlation were normalized to

141 aluminum (Al) to eliminate the potential dilution effect of background sedimentation (e.g., März et al.,

142 2011). We note, however, that Mn and Ca variations in Arctic sediments are largely independent of the

143 background lithogenic elements as shown in prior studies (e.g., Löwemark et al., 2014; Dong et al.,

144 2017)

145   X-radiographs were performed for detailed sediment structure on 1 x 10 x 30 cm slabs. X-ray

146 images were taken with SOFTEX m-100w digital x-ray scanner (EZ-320) and instantly processed using

147 iX-Pect X-Ray imaging software. Darker objects in X-radiographs presented in this study display dense

148 material, such as lithic fragments. Coarse debris >2 mm in diameter was counted at every 1 cm,

149 according to Grobe (1987). Samples for sediment texture were taken at 10 cm intervals and treated with

150 35% H_2O_2 to decompose organic matter. Sand-sized and coarser grains were separated by wet sieving

151 at 63 μm and dry sieving at 2 mm standard sieves. The grain size of <63-μm sediment was measured

152 with a Micromeritics Sedigraph 5000D.

6

153   The multibeam bathymetric survey was performed using a hull-mounted EM122 echosounder.

154   Acquired bathymetric data were processed onboard using a specialized software CARIS. Sub-bottom

155   seismic-reflection data were acquired with the CHIRP SBP-120 profiler with the vertical resolution on

156   the 10 cm scale in acoustically stratified sediments. The SBP data were logged in the TOPAS raw

157   format, which was converted to an SEG-Y format and allowed for post-processing with standard

158   software packages. The estimated sound velocity of 1500 m/s was used to convert the signal travel time

159   into sediment depth for comparison with core lithostratigraphy.

160

4. Results

4.1. Lithostratigraphy

163   Core 16B is composed of sandy to fine-grained mud displaying a general cyclic depositional

164   pattern characterized by an alternation of brown to dark brown and yellow to olive greyish sedimentary

165   units (Figs. 3, 4). The concentration of coarse IRD (>2 mm) increases at lithological boundaries and

166   within some brown units, whereas IRD within yellow to olive greyish units is typically rare.

167   A total of five Mn-rich brown units, B1 to B5, are identified in the recovered stratigraphy based

168   on sediment color, and Mn/Al ratio (Fig. 3) along with sedimentary structure and texture (Fig. 4),

169   including the upper unit B1 spliced from the box core 16B-BC. The brown units show a characteristic

170   low-L* (darker) and high-a* (more reddish) color pattern with a relatively symmetrical distribution

171   across each unit (Fig. 3). High Mn/Al ratios in surficial sediments may be partially due to high water

172   content (Thallinghii et al., 2007), but that does not affect the stratigraphic distribution of Mn values,

173   which are high in unit B1. The brown units mostly consist of slightly sandy mud with scattered IRD

174   and have relatively sharp upper boundary and uneven, bioturbated (mottled) lower boundary (Fig. 4).

175   Yellowish to olive-grey units are generally thicker than brown units and are mostly composed of fine-

176   grained mud without coarse IRD. In particular, the second grey unit between B2 and B3 is the thickest

177   interval recovered (~150 cm) with a consistent lithology. Two PW carbonate layers, consisting of

178   sandy-gravelly mud and expressed as relatively solid lenses and adjacent scattered speckles, are

179   identified above B2 and at the base of B3, corresponding to prominent Ca/Al peaks (Figs. 3, 4).

180

181   **4.2. Sediment facies**

182   Sediment of core 16B was classified into five major facies based on sedimentary structure,

183   texture, and color (Fig. 4, Table 1): bioturbated sandy mud (Bsm), bioturbated mud (Bm), crudely

184   laminated or layered mud (CLm), thinly laminated mud/muddy sand (TLm/TLs), and homogeneous

185   mud (Hm).

186   Bioturbated sandy mud (Bsm) is mostly identified in brown units and PW carbonaceous layers

187   and is characterized by strong bioturbation overprinting primary sedimentary structures, sharp upper-

188   and uneven, mottled lower-boundaries (Fig. 4). This facies mainly consists of very poorly sorted (2 – 4

189   Φ) sandy mud with coarse particles randomly scattered within the fine matrix (Fig. 3). In comparison

190   with the brown units, Bsm in the PW layers contains higher sand content, up to 22% (Fig. 3), and more

191   coarse particles.

192   Bioturbated mud (Bm) generally occurs at the transitions between brown and olive-grey units

193   and is characteristic for the olive-grey units underlying B1, where Bm is interbedded with the crudely

194   laminated or layered mud facies (CLm), and below B3 (Fig. 4). Similar to Bsm, the Bm facies also has

195   a mottled structure but is composed of more fine-grained muds with gradual facies boundaries. The

196   sediment sorting index in this facies (1 – 1.5 Φ) is lower than in Bsm due to a lack of coarse particles

197   (Figs. 3, 4).

198   Crudely laminated or layered mud (CLm) is characterized by the alternation of planar to sub-

199   parallel silt-and clay-rich laminae or layers with weak bioturbation (Fig. 4). Individual laminae are

200   generally difficult to discern. Clayey laminae/layers are mostly thicker (4 – 12 mm) than their silt-rich

201   counterparts (1 – 3 mm). Vertical changes in lamina thickness and composition appear to be random,

202   and the sediment is overall weakly sorted (1 – 1.5 Φ). Facies CLm is predominant within the second

203 olive-grey unit bounded by B2 and B3 and is also interbedded with the Bm facies in the first olive-grey

204 unit.

205   Thinly laminated mud/muddy sand facies (TLm/TLs) can be subdivided into silt-rich mud (TLm)

206 and muddy sand (TLs). This distinct facies has a laterally continuous lamination (a few millimeters

207 thick), typically identified as couplets with overlying homogeneous mud facies Hm (Fig 4). TLm/TLs

208 laminae show sharp lower and gradational upper boundaries, and relatively regular thicknesses (Fig. 3).

209 TLm/TLs facies occur episodically in the second and third olive-grey intervals, between B2 – B3 and

210 B3 – B4, respectively (Figs. 3, 4).

211   Homogeneous mud facies (Hm) consists of fine-grained muddy sediment characterized by an

212 overall absence of primary structures, bioturbation, or any grading. This facies is mostly interbedded

213 with the thinly laminated facies TLm/TLs on millimeter scales as couplets. Independently occurring

214 Hm facies (3 cm thick) is identified only in the upper part of the third olive-grey interval below B3 (Fig.

215 4).

216

217 **4.3. Geomorphology and seismostratigraphy of the AP basinward slope**

218   The seafloor of the AP top shows several sets of streamlined morphological features (Fig. 1C)

219 interpreted in prior studies as megascale glacial lineations (Niessen et al., 2013; Jakobsson et al., 2016).

220 The youngest set of lineations of SSW-NNE orientation is traced to water depths of ~950 m in the

221 southern part of AP, compared to somewhat shallower depths of 850 m in the northern part. The lineated

222 surface is covered by ~3-m thick, acoustically stratified sediment (Niessen et al., 2013).

223   The subbottom data from the basin-ward middle to the lower slope of the AP (~1600-1800 mwd)

224 show stacked series of debris lobes of ~10-20 km long and up to 25 m thick (Fig. 2A). The debris

225 deposits are covered by acoustically stratified sediments (Figs. 2B, C). Three distinct acoustic reflectors

226 (R1 – R3) were identified by lateral continuity and strong reflectivity. The reflectors divide the upper

227 part of the sediment cover into two major seismostratigraphic units, SSU 1 and 2.

228   The surficial unit SSU 1 bounded by reflectors R1 (seafloor) and R2 is laterally continuous and

229   has bedding parallel to R2 (Figs 2B and C). SSU 1 is characterized by transparent to slightly fuzzy sub-

230   bottom echoes, with a laterally discontinuous, weak reflector also recognized within this unit (Fig. 2B).

231   SSU 1 thins out from the core 16B site towards the steep middle slope (Figs. 2B and C) and is

232   indistinguishable further up-slope near the AP top.

233   As seen on records from the basin and lower slope (Fig. 2B), the underlying unit SSU 2, bounded

234   by R2 and R3, is generally characterized by parallel lamination with variable amplitudes and transparent

235   acoustic signature in the middle part of the unit. Further up-slope, two coeval debris lobes are

236   recognized in the lower part of SSU 2, covered by acoustically stratified sediment comprised of the

237   remaining part of SSU 2 and overlying SSU 1 (Figs. 2B, C). Two thin reflectors with low to moderate

238   reflectivity are identified in the lowermost part of SSU 2 in the basin, but cannot be found near the

239   debris lobes on the upper slope (Fig. 2B).

240

241   **5. Discussion**

242   **5.1. Age model**

243   As core 16B-GC lacks calcareous material for ^{14}C dating, its age model is constrained by

244   lithostratigraphic correlation with the earlier investigated core ARA03B/28B-GC from the AP top

245   (Schreck et al., 2018) and core HLY0503-8JPC from the foot of the Mendeleev Ridge ~350 km north

246   (Adler et al., 2009). The latter core record has the best regional age control constrained by multiple [14]C

247   datings in the upper part of the stratigraphy (MIS 1 to 3, units B1-B3), ^{14}C-calibrated amino-acid

248   racemization (AAR) rates extending to MIS 5 (Kaufman et al., 2008; Adler et al., 2009), and coccolith

249   occurrences (Backman et al., 2009). The correlation is primarily based on sediment color and

250   distribution of Mn and Ca content showing a regionally consistent pattern (Fig. 5) (Polyak et al., 2009;

251   Stein et al., 2010; Schreck et al., 2018). According to ^{14}C dating, brown, Mn-rich units B1 and B2,

252   including the detrital carbonate layer W3, represent MIS 1 (~ 9 ka) and MIS 3 (34-45 ka)

253   interglacial/interstadial environments, respectively (Fig. 5; Table 2). The intermittent grey unit G1 thus

254   corresponds to the last glaciation/deglaciation (MIS 2 to 1) and the youngest, pre-glacial part of MIS 3.

255   Brown units B3 to B5 recovered by core 16B-GC have been attributed in the reference cores to

256   pre-MIS 3 interstadials ranging to MIS 5c, with some variance. Based on coccolith distribution in

257   HLY0503-8JPC, Backman et al. (2009) suggested that B5 represents MIS 5a, while Adler et al. (2009)

258   assigned its age to MIS 5c, consistent with AAR ages (Kaufman et al., 2008). This age model was also

259   used by Schreck et al. (2018), who pointed out that the coccolith stratigraphy cannot pinpoint a specific

260   time interval within MIS 5. In any case, core 16B-GC does not reach the prominent detrital carbonate

261   layer PW2 identified in multiple cores across the western Arctic Ocean, including HLY0503-8JPC (Fig.

262   5), and attributed to MIS 5d (e.g., Stein et al., 2010; Bazhenova et al., 2017; Dong et al., 2017). Overall,

263   we consider MIS 5c, with the approximate age of ca. 100 ka, as a reasonable estimate for the bottom of

264   16B-GC.

265   The developed lithostratigraphic framework indicates that a ~150-cm thick grey unit G2 pre-

266   dates the MIS 3 interstadial, likely being contained in the time interval from MIS 4 to early MIS 3. This

267   prominent lithostratigraphic unit can be traced in multiple cores from the Siberian margin to the Arctic

268   Ocean interior (e.g., Spielhagen et al., 2004; Polyak et al., 2009; Dong et al., 2017; Wang et al., 2018),

269   and has been proposed to correlate to the last pre-LGM glaciation in northern Eurasia with a glacial

270   maximum around 50-60 ka (Svendsen et al., 2004).

271   In comparison to this, regionally consistent age model, the B2 unit in a very similar nearby record

272   was assigned to MIS 5 based solely on the presence of rare coccoliths, including *E. huxleyi* (core

273   SWERUS-L2-13PC; https://bolin.su.se/data/swerus/physical.php; Fig. 1) (Jakobsson et al., 2016). The

274   occurrence of *E. huxleyi* clearly indicates that the sediments of this unit are younger than MIS 8/7

275   (Thiersten et al., 1977; Backman et al., 2009). However, due to an overall limited occurrence of

276   coccoliths, *E. huxleyi* presence may not be used to identify specific substages within MIS 5 as

277   attempted previously (e.g., Jakobsson et al., 2001; Sphielhagen et al., 2004) because this species can

278   potentially occur in other relatively warm intervals of Arctic sediments. In particular, MIS 3 is a very

279   pronounced interstadial unit in Arctic paleoceanographic records, as exemplified by a high content of

280   Mn and high productivity indicated by abundant planktic and benthic microfossils (Nørgaard-Pedersen

281   et al., 1998; Adler et al., 2009; Hanslik et al., 2010; Schreck et al., 2018). Some authors further inferred

282   reduced sea ice and enhanced the advection of Atlantic waters during this time (Cronin et al., 2012;

283   Poirier et al., 2012). We note that in another SWERUS core from the Siberian margin further west, *E.*

284   *huxleyi* was also found in B2 with a ^{14}C age of 33 ka just a little above this finding (O'Regan et al.,

285   2017).

286

5.2. Core-seismic correlation

288   Distinct acoustic reflectors on seismic records can be formed by vertical changes in acoustic

289   impedance controlled by sediment density and p-wave velocity. Deep-sea Arctic Ocean sediments

290   generally consist of relatively soft fine-grained muds, which are occasionally intercalated with denser,

291   coarse sediment layers mostly attributed to iceberg sedimentation during deglacial events (Clark et al.,

292   1980; Polyak et al., 2009; Stein et al., 2010; Dong et al., 2017; Schreck et al., 2018). In particular,

293   detrital carbonate IRD layers, such as W3 and PW2 captured by western Arctic Ocean sediments, are

294   characterized in cores from the study region by a remarkably high density and p-wave velocity

295   compared to the enclosing fine-grained muds (Matthiessen et al., 2010; Stein et al., 2010; Schreck et

296   al., 2018). These layers, marking the LIS iceberg discharge from the Canadian Arctic, thus have the

297   potential to form strong reflectors.

298   Comparison of the subbottom acoustic reflection data and core 16B record (Fig. 6) confirms that

299   the surficial reflector R1 represents the seafloor, the interface between bottom water and relatively dense

300   surficial sediment of unit B1. A fine-grained unit G1 characterized by a relatively constant density and

301   p-wave velocity in multiple records from the study region (Matthiessen et al., 2010; Stein et al., 2010;

302   Schreck et al., 2018) defines a mostly transparent echo characteristic of SSU 1. A strong reflector R2

303   probably indicates a pronounced lithological change from the IRD-rich interval, including the W3 layer

304   to the overlying fine-grained unit G1 (Fig. 6). A relatively strong reflector in the uppermost part of SSU

305   2, partially divided into two layers, appears to be associated with a relatively coarse unit B2 (Fig. 6). A

12

306   transparent echo signature similar to that in SSU 1 characterizes the upper part of SSU 2 corresponding

307   to the fine-grained, thickest unit G2. Two thin reflectors of low to moderate amplitude in the lower part

308   of SSU 2 appear to correspond to a relatively sandy unit B3, especially its detrital carbonate-rich base

309   layer, and an intermittent sandy layer within G3 (Fig. 6). While core 16B-GC does not extend to the

310   depth of R3, a likely candidate for this strong reflector is a prominent, IRD-rich, detrital carbonate layer

311   PW2 that underlies unit B6 and is barely reached by 16B-GC (Fig. 5). PW2 is characterized by a

312   significantly higher wet bulk density and p-wave velocity than other Upper Quaternary sediments in

313   the study region (Mattiessen et al., 2010; Schreck et al., 2018).

314

5.3. Glaciogenic debris lobes on the AP slope

316   The presence of debris lobes in the lower part of SSU 2 up-slope from core 16B (Fig. 2) is likely

317   related to glacial scouring on the AP crest. While debris flows can potentially originate from various

318   slope instability processes, the consistent occurrence of these lobes in the Arctic Ocean downslope from

319   continental margins and plateaus characterized by glacial erosion/deposition indicates their glaciogenic

320   origin. In particular, the CB features debris lobes on both the AP and Chukchi Plateau slopes (Niessen

321   et al., 2013; Dove et al., 2014 and reference therein), indicating glaciogenic inputs from both sides. The

322   mechanisms for debris flows could include sediment bulldozing on the tops by grounded ice and/or

323   rapid deposition of deglacial sediments, e.g., from glacial underflows (e.g., Powell, 1990; Batchelor

324   and Dowdeswell, 2015). The timing of debris lobe emplacement would thus correspond to glacial

325   advance or early deglaciation stages.

326   The SSW-NNE orientation and southward inclination of the youngest set of glacial lineations

327   on the AP top indicate that the eroding ice was sourced from the East Siberian margin (Niessen et al.,

328   2013). According to the age constraints developed for core 16B-GC and a correlative core 28B-GC

329   from the AP top (Fig. 5), this event occurred prior to the MIS 3 interstadial and was possibly related to

330   a major glaciation in northern Siberia dated to the time interval between ~70 and 50 ka (Svendsen et

331   al., 2004; Möller et al., 2015).

332   The upper seismostratigraphic unit SSU 1 bounded by the seafloor, and the W3 layer represents

333   deposition during the time interval from late MIS 3 to the Holocene (Fig. 6). The absence of glaciogenic

334   debris lobes within SSU 1 indicates that glacial erosion of the AP was unlikely during the LGM, as

335   suggested in previous studies (Niessen et al., 2013; Schreck et al., 2018).

336

5.4. Glacier-induced depositional environments

338   Based on the developed age model, sedimentary unit G2 (MIS 4 to early MIS 3) that is

339   characterized by laminated mud facies, has a relatively high sedimentation rate of ~6 cm/kyr as

340   compared to more bioturbated units G3 and G4 (Fig. 4, Table 2). Laminated structures of glaciogenic

341   sediments in the high-latitude oceans can be potentially formed under various depositional settings such

342   as permanent sea ice, turbid meltwater plumes, distal fine-grained turbidity currents, and contour

343   currents (Hesse et al., 1996; Darby et al., 1997; Kleiber et al., 2000; ÓCofaigh et al., 2003; Matthiessen

344   et al., 2010). However, contour currents are unlikely to be strong in a semi-enclosed Arctic deep-sea

345   basin, while sedimentation rates under perennial sea ice are too low to form a laminated sedimentary

346   sequence. Couplets of TLm/TLs and Hm facies similar to those observed in 16B-GC (Fig. 4) have been

347   interpreted in previous studies as deposition from underflow turbidity currents in combination with

348   rapid settling from fine-grained suspension plumes in the attenuation stage of the carrying current (Piper,

349   1978; Chough, 1984; Yoon et al., 1991). The occurrence of TLm/TLs facies within G2 and G3 suggests

350   that turbidities could be associated with the glaciogenic debris flows emanating from their front on a

351   relatively steep AP slope (Fig. 2B). The development of CLm facies may reflect deposition from turbid

352   meltwater plumes repeatedly discharged from ice-sheet termini (Hesse et al., 1996; Kleiber et al., 2000;

353   ÓCofaigh et al., 2003).

354   Glacial erosion on glaciated continental margins plays an important role in delivering large

355   amounts of sediments to the continental slope and the adjacent basins, and thus leading to the formation

356   of debris flows (Vorren et al., 1998; Kleiber et al., 2000; Dowdeswell and Elverhøi, 2002; Jakobsson et

357   al., 2008). Based on our seismostratigraphic and sediment core data, this scenario applies to the AP

14

358   during MIS 4, when large volumes of eroded sediment may have been transported as proglacial under-

359   and overflows into the adjacent deep basin. This rapid sediment deposition likely caused slope

360   instability and thus triggered debris flows. Suspended fine sediments could have been detached from

361   the slope and transported into the interior of the Arctic Ocean at the density boundaries between

362   different water masses with subsequent deposition. Prolonged glacial erosion could have also formed

363   repeatedly detached turbid layers, leading to the formation of crudely laminated mud (CLm) facies (Fig.

364   7A). As the grounded ice retreated, turbid meltwater underflow plumes were likely repeatedly

365   discharged from the grounding line, thereby depositing CLm facies (Hesse et al., 1996; ÓCofaigh et al.,

366   2003) (Fig. 7B). These glacier-induced processes could lead to the deposition of fine-grained, fully

367   laminated sediments with fairly high sedimentation rates (unit G2) during the MIS 4 deglaciation that

368   may have extended to early MIS 3.

369   Bioturbated sediments of B2 and the overlying W3 layer are interpreted to represent hemipelagic

370   depositional environments with drifting icebergs during a relatively warm period of MIS 3 (Fig. 7C).

371   In general, Mn-enriched, brown sediments are believed to reflect elevated sea levels and/or reduced

372   sea-ice conditions, under which sediment from the Siberian shelves could be transported to the central

373   Arctic basins (Jakobsson et al., 2000; Polyak et al., 2004, 2009; Stein et al., 2010; Löwemark et al.,

374   2014; Ye et al., 2019). Overall, sedimentary properties of G2 and overlying facies indicate depositional

375   environments affected by the advance and retreat of a marine-based ice sheet grounded on the AP top.

376   The absence of glaciogenic debris lobe within SSU 1 indicates that the AP has not encountered

377   glacial erosion during MIS 2, consistent with previous studies from the Arctic Ocean off the Siberian

378   margin (Niessen et al., 2013; Schreck et al., 2018) as well as terrestrial studies from adjacent islands

379   and the East Siberian mainland (Romanovskii et al., 2004; Gualtieri et al., 2005). Nevertheless, the

380   presence of CLm facies within unit G1, most of which is attributed to MIS 2 (Figs. 3, 4), reflects glacier-

381   induced depositional processes, such as the suspension settling of turbid meltwater plumes and detached

382   turbid layers (Figs 7D, E). This lithology indicates the influence of a limited glaciation in the study

383   region, although its location and distribution pattern (e.g., grounded ice vs. ice shelf) needs to be further

15

384   investigated. There is no evidence for the LGM glaciation at the East Siberian margin, but data from

385   the eastern Chukchi shelf margin show glaciogenic bedforms indicative of grounded ice, age-

386   constrained to MIS 2 (Polyak et al., 2007). These bedforms are restricted to water depths of less than

387   ~450 m, but we cannot exclude that this grounded ice extended over the Chukchi shelf to its western

388   margin bordering the Chukchi Basin.

389   The predominance of the bioturbated sandy mud (Bsm) facies in the uppermost unit B1 reflects

390   hemipelagic deposition with sea-ice and/or iceberg debris rafting during much of the Holocene (Fig.

391   7F). Intense bioturbation indicates a relatively elevated export of primary production to the seafloor,

392   and thus moderate sea-ice conditions, and/or enhanced oxygenation of surface sediments.

393

394   **6. Conclusions**

395   A combination of litho- and seismostratigraphy on the basinward slope of the Arliss Plateau

396   depicts a glacier-induced depositional history in front of the East Siberian continental margin since

397   estimated MIS 5c (~100 ka). The age constraints were obtained by correlation of the studied core

398   ARA02B/16B-GC with earlier developed stratigraphies (Adler et al., 2009; Schreck et al., 2018). Based

399   on this age model and the core-seismic correlation, deposition of the youngest glaciogenic debris lobes

400   on the slope, indicative of the last glacial erosion of the AP top, occurred within seismostratigraphic

401   unit SSU 2 during MIS 4 to early MIS 3. This event may be correlative to a major glaciation in northern

402   Siberia that has been constrained to the time interval between ~70 and 50 ka (Svendsen et al., 2004;

403   Möller et al., 2015). Laminated sedimentary facies and sedimentation rates as high as ~6 cm/kyr in the

404   respective core interval (lithological unit G2) reflect glacier-induced sediment deposition. This

405   enhanced sedimentation probably resulted from detached turbid layers, turbidity currents, and

406   meltwater discharge pulses during glaciation/deglaciation.

407   The absence of glaciogenic debris lobes within the upper seismostratigraphic unit SSU 1 (late

408   MIS 3 to 1) indicates that glacial erosion unlikely impacted the AP during the LGM, consistent with

409 previous inferences (Niessen et al., 2013; Schreck et al., 2018). Nevertheless, the presence of laminated

410 facies within the corresponding core interval G1 may indicate a limited glaciation at a nearby

411 continental margin. Further investigation by geophysical seafloor surveys verified by sediment core

412 records is needed to better understand the glacial history of the western Arctic Ocean.

413

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570 **Figure and Table captions**

571 **Figure 1** (A) Index physiographic map of the study area showing the location of cores ARA02B/16B-

572 GC (this study), ARA03B/28B-GC (Schreck et al., 2018) and HLY0503-8JPC (Adler et al., 2009). MR,

573 Mendeleev Ridge; AP, Arliss Plateau; CB, Chukchi Basin; CP, Chukchi Plateau; NR, Northwind Ridge.

574 Solid line tracks sub-bottom profiler data. Bathymetric contour interval is 250 m. (B) Broader

575 physiographic/paleogeographic context with inferred ice-sheet maxima shown by semi-transparent

576 white fill (Svendsen et al., 2004; Niessen et al., 2013; Stokes et al., 2016). (C) Multibeam bathymetry

577 of the Arliss Plateau showing several generations of Mega-Scale Glacial Lineations. Bathymetric

578 contour interval is 100 m.

579 **Figure 2** Sub-bottom profiler from Arliss Plateau to Chukchi Basin across the site of core

580 ARA02B/16B-GC (red vertical bar in panel B). Seismostratigraphic units SSU 1 and 2 are bounded by

581 distinct reflectors R1 to 3. The uppermost glaciogenic debris lobes on the slope are correlated with the

582 lower part of SSU 2. Black box shows the position of Fig. 6.

583 **Figure 3** Down-core distribution of sediment lightness (L*), Mn/Al and Ca/Al ratios, IRD numbers,

584 and grain size in core ARA02B/16B-GC. Top 10 cm are estimated to be missing due to overpenetration;

585 XRF data for this interval (blue-colored solid lines) are spliced from the box core 16B-BC. High Mn/Al

586 and Ca/Al ratios in surficial sediment may be partially due to high water content (Thallinghii et al.,

587 2007). Lithostratigraphic indices are shown in Fig. 3.

588 **Figure 4** Split-core photos, X-radiograph images, and sedimentary facies (see Table 1 for details).

589 Indices for lithological units (B1-B5 and G1-G4) and the detrital carbonate layer W3 are shown on the

590 core photo.

591 **Figure 5** Lithostratigraphic correlation of core ARA02B/16B-GC with ARA03B/28B-GC (Schreck et

592 al., 2018) and HLY0503-8JPC (Adler et al., 2009) based on sediment color and variations of Mn/Al

593 and Ca/Al. The tops of major brown units (B2-B5) are correlated by dotted lines; the bottom of core

594   ARA03B/28B-GC is projected onto ARA02B/16B-GC by the dashed line. Estimated Marine Isotope

595   Stages and radiocarbon/AAR ages (Adler et al., 2009) are shown next to core HLY0503-8JPC.

596   **Figure 6** Correlation of the subbottom data with core ARA02B/16B-GC. Surficial reflector R1 is the

597   seafloor. Strong reflectors R2 and R3 correspond to prominent lithological changes at the boundaries

598   of fine-grained sediments and IRD-rich layers W3 and PW2. Estimated Marine Isotope Stages are

599   shown on the right.

600   **Figure 7** Schematic model of sedimentary processes on the Arliss Plateau and adjacent Chukchi Basin

601   during major depositional regimes from the MIS 4 glaciation to the Holocene.

602

603   **Table 1** Summary of sedimentary facies in core ARA02B/16B-GC. Black arrows indicate the intervals

604   of TLm, TLs, and Hm facies. See Fig. 4 for the stratigraphic distribution of sedimentary facies.

605   **Table 2** Estimated duration and sedimentation rates of lithostratigraphic units in core ARA02B/16B-

606   GC based on the age model developed from lithostratigraphic correlations (Fig. 5).

Figure 1
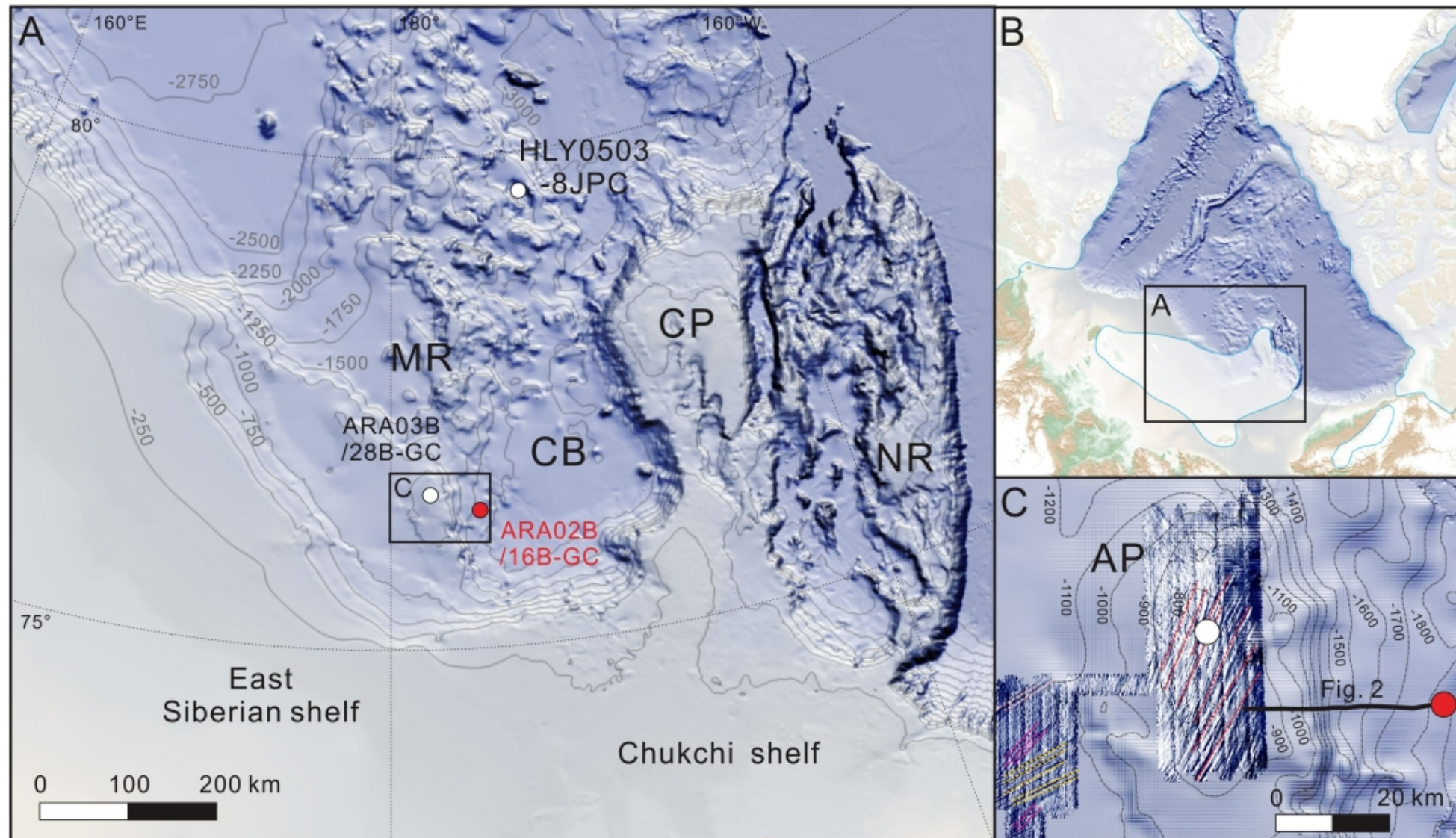
Figure 2
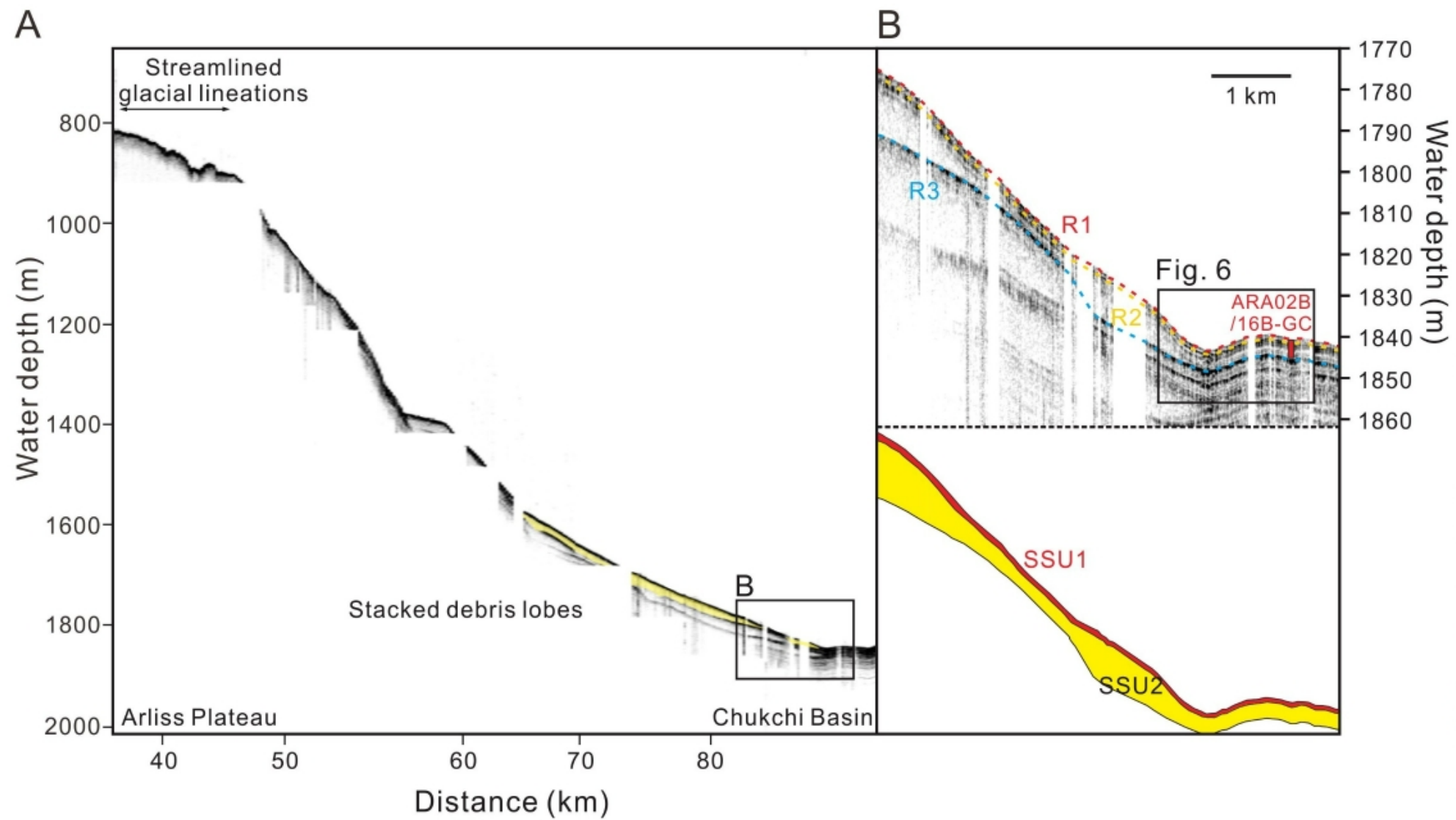
Figure 3
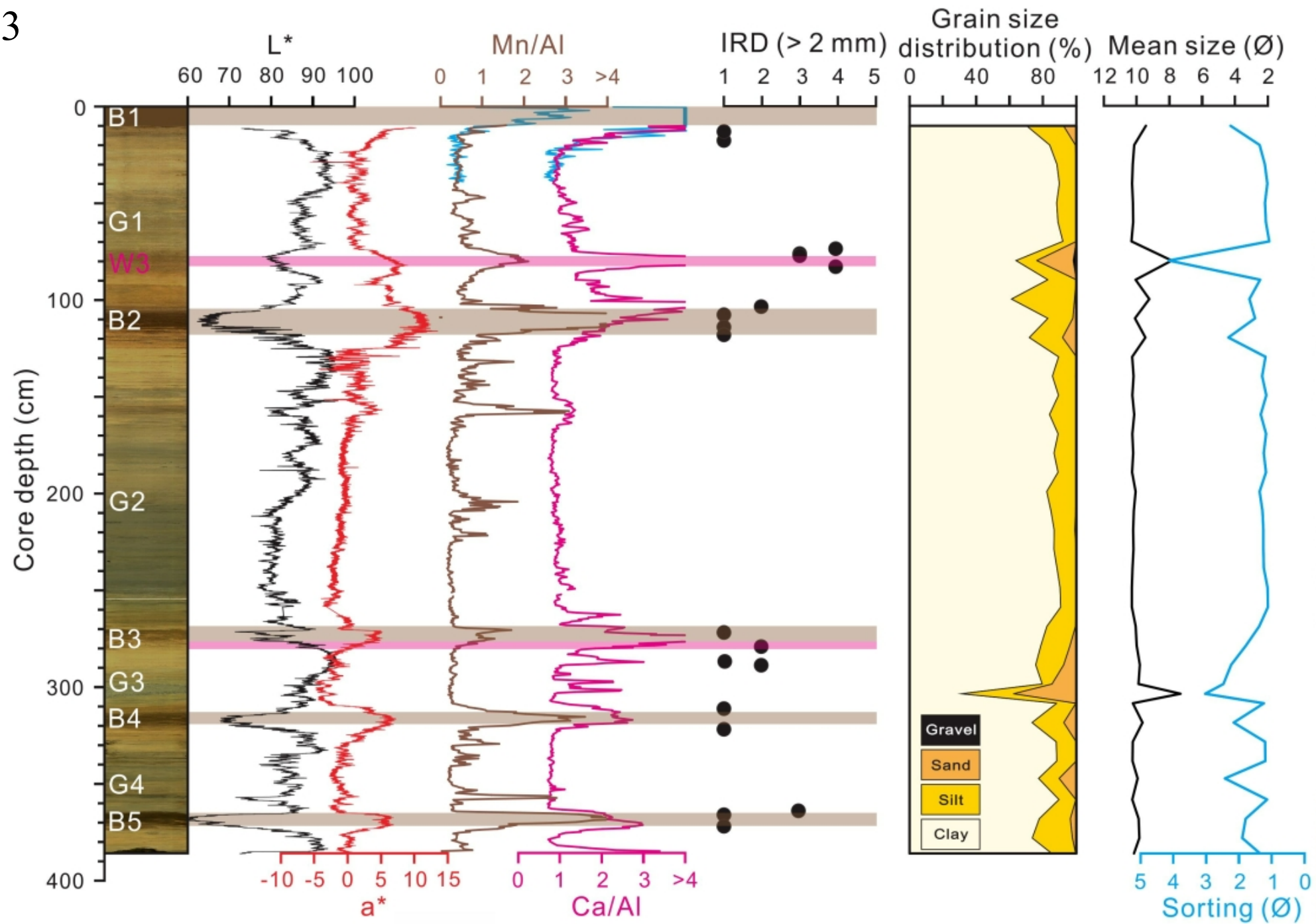
Figure 4
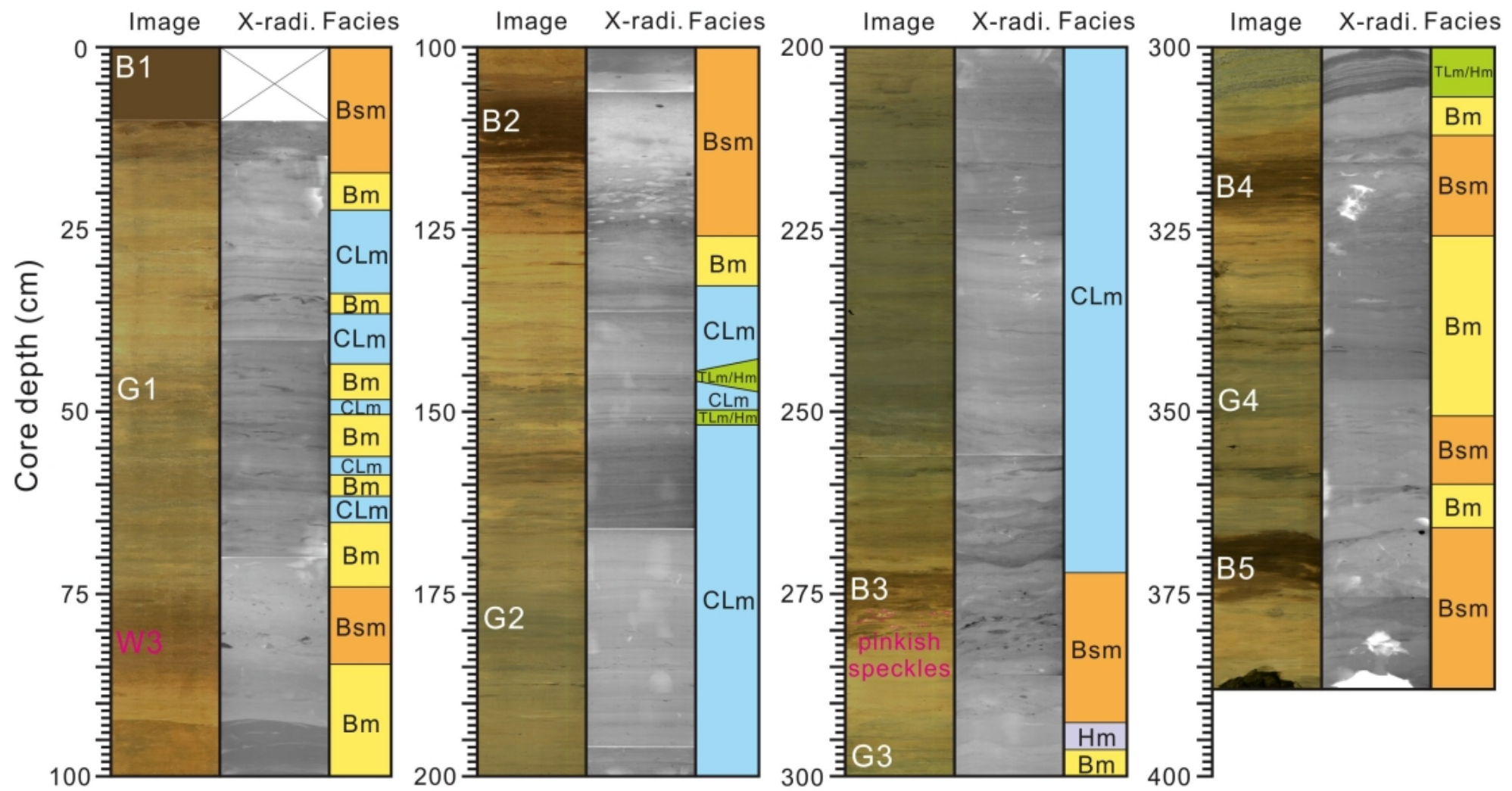
Figure 5
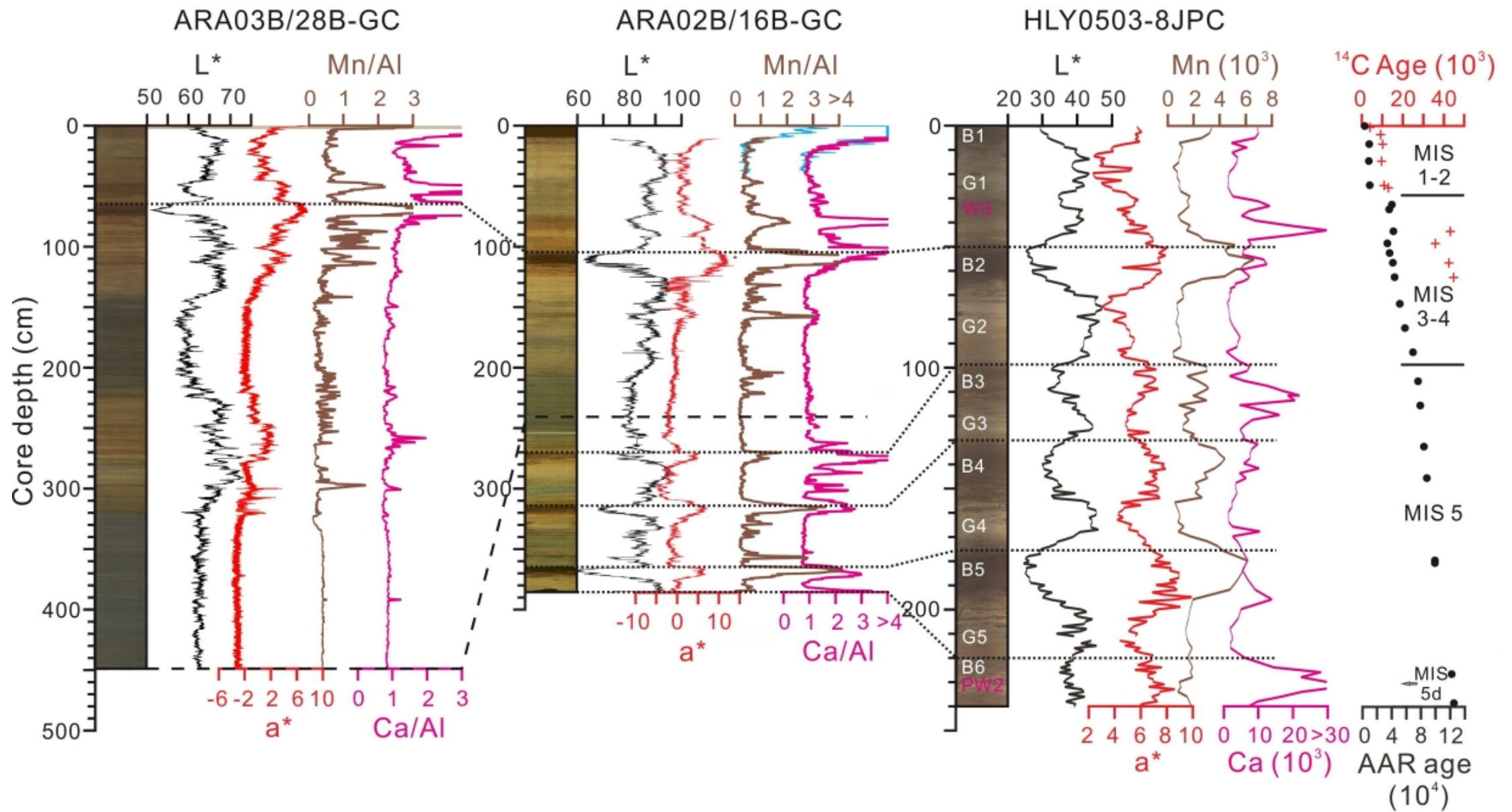
Figure 6

Figure 7

A. Stadial (MIS 4)
Present sea level
Glacial erosion
Arliss Plateau
Atlantic layer
Detached turbid layers
Deep water
Glaciogenic debris flow & turbidity current
Stacked glaciogenic debris flow deposits

B. Deglaciation (MIS 3-4)
Present sea level
Glacial lineations
Arlis Plateau
Turbid meltwater plume
Turbidity current
Stacked glaciogenic debris flow deposits

C. Interstadial (MIS 3)
Present sea level
Laurentide icebergs
Hemipelagic settling
Glacial lineations
Arliss Plateau
Stacked glaciogenic debris flow deposits

D. Glacial (MIS 2)
Present sea level
Glacial lineations
Arliss Plateau
Detached turbid layers
Stacked glaciogenic debris flow deposits

E. Deglaciation (MIS 1-2)
Present sea level
Glacial lineations
Arliss Plateau
Turbid meltwater plume
Stacked glaciogenic debris flow deposits

F. Interglacial (MIS 1)
Present sea level
Hemipelagic settling
Glacial lineations
Arliss Plateau
Stacked glaciogenic debris flow deposits

Table 1

Facies		Description	Depositional process
Bioturbated sandy mud (Bsm)		Brown to dark brown and pinkish; (gravelly) sandy mud, IRD-rich; very poorly sorted; intensely bioturbated, absence of primary structure; relatively sharp and flat upper boundaries and mottled lower basal contact	Hemipelagic settling affected by drifting ice (iceberg/sea-ice)
Bioturbated mud (Bm)		Yellow to olive grey; fine-grained mud, IRD is rare or absent; poorly sorted; located near brown and pinkish layers, repeatedly alternated with poorly defined laminated facies; gradual facies boundaries	Hemipelagic settling
Crudely laminated or layered mud (CLm)		Yellow to olive grey; silty mud to clayey mud, IRD is absent; poorly sorted; weakly bioturbated, planner and sub-parallel lamina or layers, lamination is laterally discontinuous and poorly defined, clayey lamina or layer is thicker (4-12 mm) than silty lamina (1-3 mm), non-systematic vertical changes, gradual facies boundaries	Repeated settling of suspension from turbid melt water plumes and/or detached turbid layers
Thinly laminated mud/sand (TLm/TLs)		Olive grey; silt-rich mud/sandy mud; IRD is absent; distinct and laterally continuous laminae are coupled with homogenous muds, relatively regular lamina thickness (1-3 mm), sharp lower boundaries and gradational upper boundaries	Bottom-hugging turbidity current
Homogeneous mud (Hm)		Olive grey; fine-grained mud, IRD is absent; poorly sorted; absence of primary structure as well as any grading and bioturbation; generally coupled with TLm/TLs facies	Rapid settling from fine-grained suspension plumes in the weakening stage of turbidity current

Black arrows indicate each facies (TLm, TLs, and Hm)

Table 2

Lithostratigraphy	Thickness (cm)	Duration (kyr)	Sedimentation Rate (cm/kyr)
B1	10 (0-10)	9 (0-9)	0.83
G1	68 (10-78)	25 (9-34)	2.72
W3-B2	37 (78-115)	11 (34-45)	3.36
G2	157 (115-272)	26 (45-71)	6.04
B3-B5	103 (272-375)	25 (71-96)	4.12