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1 **Ice sheet retreat and glacio-isostatic adjustment in Lützow-**
2 **Holm Bay, East Antarctica**

3
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21 **Abstract**

22 The East Antarctic Ice Sheet has relatively few field data to constrain its
23 past volume and contribution to global sea-level change since the Last
24 Glacial Maximum. We provide new data on deglaciation history and
25 develop new relative sea-level (RSL) curves along an 80 km transect (from
26 Skallen to Skarsvnes, Langhovde and the Ongul Islands) in Lützow Holm
27 Bay, East Antarctica. The geological constraints were compared with
28 output from two Glacial Isostatic Adjustment (GIA) models. The minimum
29 radiocarbon age for regional deglaciation is c. 11,240 cal. yr BP on West
30 Ongul Island with progressively younger deglaciation ages approaching the
31 main regional ice outflow at Shirase Glacier. Marked regional differences
32 in the magnitude and timing of RSL change were observed. More in
33 particular, in Skarvsnes a minimum marine limit of 32.7 m was inferred,
34 which is c. 12.7 m higher than previously published evidence, and at least
35 15 m higher than that reported in the other three ice-free areas. Current GIA
36 model predictions slightly underestimate the rate of Late Holocene RSL fall
37 at Skallen, Langhovde, and West Ongul, but provide a reasonable fit to the
38 reconstructed minimum marine limit at these sites. GIA model predictions
39 are unable to provide an explanation for the shape of the reconstructed RSL
40 curve at Skarvsnes. We consider a range of possible explanations for the

41 Skarvsnes RSL data and favour an interpretation where the anomalously
42 high marine limit and rate of RSL fall is due to reactivation of a local fault.

43

44 **Key-words:** Sea level changes; Antarctica; Holocene; Coastal
45 geomorphology; Isolation lakes; Raised beaches; Glacial Isostatic
46 Adjustment (GIA) Models; Neotectonics

47

48

49 **1. Introduction**

50 Estimates of the contribution of the continental ice-sheets to past and recent
51 global sea-level change are still relatively imprecise (Bromwich & Nicolas
52 2010, Clark & Tarasov 2014). This is due to an incomplete understanding
53 of changes in continental ice volume, including the maximum extent of
54 glaciation, and the onset and rates of ice retreat. Some of this information
55 can be inferred from radiocarbon dating of organic deposits that have
56 accumulated after ice retreat, and from changes in relative sea-level (RSL)
57 resulting from the glacio-isostatic response of the Earth's crust to ice mass
58 changes. Accurate RSL reconstructions, together with GPS-derived uplift
59 data, can track regional changes in glacial isostatic adjustment (GIA)
60 (Thomas et al. 2011, Hodgson et al. 2016), a process that contaminates
61 satellite gravity measurements of present-day ice sheet mass balance (e.g.,
62 Chen et al. 2009, Shepherd et al. 2012, Williams et al. 2014). In regions
63 where measurements of GIA are sparse, or where modelled estimates are
64 not compared with geological constraints, large errors can be introduced
65 into the GIA correction and hence the mass balance calculations (Velicogna
66 & Wahr 2013). In Antarctica, the paucity of GIA constraints limits the
67 accuracy of estimates of changes in the mass balance of the ice sheets
68 derived from the Gravity Recovery and Climate Experiment (GRACE;

69 Velicogna & Wahr 2013, Clark & Tarasov 2014) as well as predictions of
70 future ice sheet contributions to global sea-level rise (e.g., Adhikari et al.
71 2014). Increasing the spatial resolution of geological data on ice sheet
72 retreat and RSL reconstructions is therefore a recognized research priority
73 (e.g., Watcham et al. 2011, Bentley et al. 2014).

74 Post Last Glacial Maximum (LGM) changes in RSL in previously
75 glaciated regions principally reflect three processes: eustatic sea-level rise,
76 regional GIA, and neotectonic events (Stewart et al. 2000, Bentley et al.
77 2005). The latter are generally assumed to be only important in tectonically
78 active regions (e.g., Pacific coastline of North America (Plafker 1972), the
79 southern part of the Strait of Magellan and southernmost Tierra del Fuego
80 (Bentley & McCulloch 2005)), and can be the dominant forcing of regional
81 variability in RSL changes. However, post-glacial unloading and rebound
82 can also lead to the formation or re-activation of faults in continental
83 shields and hence tectonic activity in otherwise stable areas (e.g.,
84 Lagerbäck 1978, Risberg et al. 2005, Steffen et al. 2014). Therefore, if
85 RSL changes are significantly influenced by neotectonic faulting, this needs
86 to be taken into account when validating GIA models (Watcham et al.
87 2011).

88 Of all ice sheets, the Antarctic ice-sheets probably have the fewest
89 RSL field data (Bentley et al. 2014, Mackintosh et al. 2014). This has
90 resulted in a wide range of model-based estimates of Antarctic Ice Sheet
91 contributions to global sea-level since the LGM, varying from 35 m
92 (Nakada & Lambeck 1988) to 13.6 m (Argus et al. 2014), 9 ± 1.5 m
93 (Whitehouse et al. 2012a), and even 9 to 6 m (Gomez et al. 2013). Given (i)
94 the potential of the EAIS to raise global sea-level by up to 50 m
95 (Huybrechts 2002), and (ii) some studies suggest that the melting of the
96 EAIS might have contributed to the Eemian sea-level high stand, which
97 was 6 to 9 m higher than today (Kopp et al. 2009, Pingree et al. 2011),
98 identifying those areas of the EAIS that respond to Holocene and recent
99 climate changes is critical (Mackintosh et al. 2014).

100 Two complementary approaches are traditionally used to develop
101 RSL curves in Antarctica. The first one relies on radiocarbon dating of
102 marine fossils in raised beaches as direct evidence of former sea-level
103 changes (e.g., Berkman et al. 1998, Miura et al. 1998). The shortcoming of
104 this approach however, is that the organisms producing the shells used for
105 dating occur at different depths in the marine environment (Shennan et al.
106 2015). Dating fossils in raised beaches therefore typically provides
107 minimum constraints on the height of former sea-levels (Shennan et al.

108 2015). The second approach is based on isolation lakes, which are natural
109 depressions in the bedrock that have been inundated by and subsequently
110 isolated from the sea as a result of RSL fall (Verleyen et al. 2004). The
111 isolation event is identified by studying markers of marine and lacustrine
112 phases, such as diatoms, fossil pigments and sedimentological changes
113 (Watcham et al. 2011). The RSL curves are then derived from studying the
114 timing of marine-lacustrine transitions in isolation basins situated at
115 different altitudes (Zwartz et al. 1998). The advantage of isolation basins is
116 that the height of their sills can be measured with precision and that this
117 height therefore provides a better vertical constraint compared with that of
118 fossils in raised beaches (Takano et al. 2012). Moreover, because in
119 isolation lakes the organic matter in the lacustrine sediments that are
120 deposited in equilibrium with atmospheric CO₂ can be dated, problems
121 associated with the marine radiocarbon reservoir effect can be
122 circumvented (Hodgson et al. 2001, Verleyen et al. 2005). One drawback of
123 the isolation basin approach is that during storm over wash events marine
124 diatoms can be transported into the lake, which can complicate to
125 discriminate between lacustrine and marine sediments (Verleyen et al.
126 2004). A second shortcoming is that in saline and brackish lakes in
127 Antarctica, the diatom communities are similar to those in the Southern

128 Ocean (Verleyen et al. 2003), making it difficult to exactly identify the
129 transition from marine to lacustrine sediments based on diatoms alone.
130 However, despite the shortcomings of both approaches, they have been
131 successfully applied to develop RSL curves in parts of the Antarctic
132 Peninsula (Bentley et al. 2005, Hall 2010, Roberts et al. 2011, Watcham et
133 al. 2011) and a few ice-free regions along the East Antarctic coastline, such
134 as the Vestfold Hills (Zwartz et al. 1998), Windmill Islands (Goodwin &
135 Zweck 2000), Rauer Islands (Berg et al. 2010, Hodgson et al. 2016), and
136 Larsemann Hills (Verleyen et al. 2005).

137 Here, we present new RSL constraints for islands and peninsulas in
138 the Lützow-Holm Bay region (Dronning Maud Land, East Antarctica,
139 Fig.1) based on two coastal lakes from Skarvsnes and five lakes from West
140 Ongul Island situated at different elevations, as well as new raised beach
141 data from Skarvsnes. We combined our data with recently published
142 records from an isolation basin on Skallen and one on Skarvsnes (Takano et
143 al. 2012), as well as with radiocarbon dates of fossils incorporated into
144 raised beaches on Skallen, Skarvsnes, Langhovde and West Ongul Island
145 (Miura et al. 1998; Fig.1). These geological constraints were subsequently
146 compared with regional predictions of RSL evolution and high stand from
147 two recently-developed GIA models, namely the ICE-6G_C model (Argus

148 et al. 2014) and the W12 model (Whitehouse et al. 2012a), in order to
149 assess the potential offset between modelling results and the near-field data.

150

151 **2. Site description**

152 Lützow-Holm Bay is part of Antarctic Drainage System 7 based on ICESat
153 data (Fig.1) and is the discharge point of one of the larger East Antarctic
154 glacier systems (Zwally et al. 2012), the Shirase Glacier, as well as of a
155 number of smaller glaciers (Miura et al. 1998). The bay includes several
156 ice-free peninsulas and islands composed of gneisses, metabasites, and
157 granites, together with thin beds of marble and quartzite (Tatsumi & Kizaki
158 1969). Different fault systems have been mapped, including one on
159 Skarvsnes and one between West and East Ongul Island (Ishikawa et al.
160 1976; Fig.1), but there are no records of neotectonic activity.

161 West Ongul Island is the largest ice-free island in the region. It is
162 separated from the Antarctic continent by a c. 600 m deep glacial trough
163 (Mackintosh et al. 2014) in front of the Langhovde and Hazuki Glaciers
164 (Miura et al. 1998), and from East Ongul Island by the 40 m wide Naka-no-
165 seto Strait. ^{14}C dates of *in situ* fossils in raised beaches on the Ongul Islands
166 fall into two age classes; pre-LGM and Holocene. It has therefore been
167 suggested that this part of the region was ice-free during the LGM and

168 Marine Isotope Stage (MIS) 3 (Nakada et al. 2000), or even MIS 6-7
169 (Takada et al. 2003). The maximum Holocene marine limit for the region
170 was estimated to be 17 m (10,590 +/- 160 ¹⁴C yr BP; Miura et al. 1998).

171 Langhovde is one of the two main peninsulas in the region. It is
172 situated to the south west of the Langhovde Glacier and to the north east of
173 the Honnør Glacier (Fig.1). Marine fossils in the raised beaches are either
174 of Late Pleistocene (or older) or Holocene age. The pre-Holocene ages are
175 however only found on the northern part, which has led to the suggestion
176 that this part was ice free during the LGM, whereas the southern part was
177 probably ice-covered (see Mackintosh et al. 2014 for a review). The
178 maximum Holocene marine limit has been estimated at 17 m (6,810 +/- 60
179 ¹⁴C yr BP; Miura et al. 1998).

180 Skarvsnes is the second of the two largest peninsulas and is situated
181 south of Langhovde in between glacial troughs in front of the Honnør and
182 Telen Glaciers (Miura et al. 1998). All but one of the ¹⁴C-dated fossils
183 derived from raised marine deposits on this peninsula are of Holocene age
184 (Miura et al. 1998), suggesting that the region was ice-covered during the
185 LGM. This is confirmed by a recent cosmogenic isotope dating campaign,
186 which revealed that Skarvsnes emerged from at least 350 m of ice cover
187 between 10 and 6 ka BP (Yamane et al. 2011). The maximum Holocene

188 marine limit at 8,440 +/- 140 ¹⁴C yr BP was estimated at c. 20 m based on
189 raised beach data (Miura et al. 1998).

190 Skallen is a smaller peninsula to the south west of Skarvsnes close
191 to the Skallen Glacier (Takano et al. 2012). It lies to the north east of the
192 Shirase Glacier which has created a large glacial trough in Lützow-Holm
193 Bay. All the fossils sampled in raised beach deposits are of Holocene age
194 and relatively recent. The maximum Holocene marine limit is at 12 m and
195 dated at 4,720 +/- 90 ¹⁴C yr BP based on raised beach data (Miura et al.
196 1998).

197

198 **3. Material and methods**

199 3.1. Geomorphological measurements, sampling of raised beaches and lake 200 sediment coring

201 Three specimens of marine macrofossils (*Laternula elliptica*, and
202 polychaete worm tubes) were sampled in raised beaches at different
203 altitudes in Skarvsnes. Sill heights of the lakes and the raised beach
204 deposits were surveyed using a Trimble 5700 base station GPS receiver
205 cross-referenced to the IGS station at Syowa (code SYOG). As a test of the
206 vertical accuracy, Geodetic Station No 39-02 was resurveyed giving an
207 ellipsoidal height error of ± 0.97 cm. Altitudes were referenced to vertical

208 datum WGS84 with the EGM96 geoid separation ranging from 21.14 to
209 22.02 m (mean 21.62 m between the ellipsoidal height and the orthometric
210 height). Where data could not be referenced to the IGS station, spot heights
211 of the sills of the lakes were used from previous mapping surveys (Kimura
212 et al. 2010). A 3 m vertical error bar was used when developing the RSL
213 curves to account for differences between low and high tide in the region.
214 This error bar was based on tidal gauge records measured between April
215 2010 and December 2011 (Aoyama et al. 2016).

216 Sediment cores were extracted from seven lakes at a range of
217 altitudes above sea level. Five lakes were cored on West Ongul Island
218 [Yumi Ike (WO1), Ô-Ike (WO4), Ura Ike (WO5), Higashi Ike (WO6), and
219 Nishi Ike (WO8)] and two lakes on Skarvsnes [(Mago Ike (SK1) and
220 Kobachi Ike (SK4)]; the codes refer to Tavernier et al. (2014) and Verleyen
221 et al. (2012) in which more information on the limnological properties of
222 the cored lakes can be found. All lakes were freshwater, except Kobachi Ike
223 which was brackish (Tavernier et al. 2014). Sediment cores were extracted
224 using a UWITEC gravity corer for surface sediments and a Livingstone
225 square-rod piston sampler (Wright 1967) for intermediate to basal
226 sediments. Bedrock or glacial sediments were present at the base of all the
227 sediment cores.

228

229 3.2. Paleolimnological analyses

230 To identify marine to freshwater transitions in the sediment cores, multiple
231 biological and sedimentological proxies were analysed. Gamma ray density
232 (GRD) and volume-specific magnetic susceptibility (MS), converted to
233 mass-specific MS, were measured using a Bartington 1 ml MS2G sensor
234 for those cores which were transported unsliced. The total carbon (TC)
235 content was quantified using a Flash 2000 Organic Elemental Analyzer.
236 Measurements were carried out by dry combustion at high temperature (left
237 furnace: 950°C and right furnace: 840°C; King et al. 1998). This was then
238 followed by separation and detection of the gaseous products. The data
239 were processed using the Eager Xperience software. Samples were all run
240 at least twice to detect and exclude possible erroneous values. Outliers were
241 excluded and the mean value of replicates was used. Reproducibility within
242 and between different runs was tested using standards. Diatoms were
243 prepared following standardized protocols (Renberg 1990), with absolute
244 abundances calculated following Battarbee & Kneen (1982). Diatoms were
245 counted under oil immersion using a Zeiss axiophot light microscope at a
246 magnification of 1000x. At least 400 valves (>2/3 intact or at least
247 unambiguously containing the middle part of the sternum for pennate

248 diatoms) were counted in each sample, except when concentrations were
249 too low to reach this number. In the latter case samples were first
250 concentrated and then slides were screened in their entirety. Taxonomic
251 identification was mainly based on Sabbe et al. (2003), Ohtsuka et al.
252 (2006), Van de Vijver et al. (2011) and Esposito et al. (2008) for the
253 freshwater diatoms, and Cremer et al. (2003) and Scott & Thomas (2005)
254 for the marine and brackish-water diatoms. Diatoms were grouped into
255 freshwater, brackish and marine species based on their weighted-averaging
256 conductivity optima as calculated in Tavernier et al. (2014). Species were
257 considered as freshwater taxa when their WA-optimum was below 1.5
258 mS/cm. Species were regarded as brackish-water taxa when their WA-
259 optimum fell between 1.5 mS/cm and 4.42 mS/cm (Tavernier et al. 2014).
260 In the sediment cores from the brackish lake (Kobachi Ike, SK4), fossil
261 pigments were additionally analysed, because in brackish and saline lakes
262 identifying the marine-lacustrine transition based on fossil diatoms is
263 sometimes complicated due to the presence of species shared between both
264 environments (Hodgson et al. 2006a). The fossil pigments were extracted
265 and analysed following Van Heukelem & Thomas (2001). The system was
266 calibrated using authentic pigment standards and compounds isolated from
267 reference cultures following Scientific Committee on Oceanic Research

268 (SCOR) protocols (DHI, Denmark). The identification of the pigments was
269 based on Jeffrey et al. (1997) and pigments of unknown affinity were
270 assigned as ‘unknown’ or as derivatives of the pigment with which they
271 showed the closest match based on retention times and absorption spectra.
272 Concentrations of individual pigments in the samples were calculated using
273 the response factors of standard pigments. The abundance of the
274 cyanobacteria pigments zeaxanthin, echinenone, and myxoxanthophyll is
275 reported as a percentage of the total carotenoids (%). Myxoxanthophyll is
276 exclusively produced by cyanobacteria and was therefore considered as the
277 preferred marker pigment for this group, which are the dominant
278 photoautotrophs in lacustrine microbial mat communities in East Antarctica
279 (Hodgson et al. 2004, Verleyen et al. 2010). Hence, the presence of
280 myxoxanthophyll, a dominant pigment in lacustrine Antarctic sediments,
281 was used to diagnose the onset of lacustrine conditions. This is because
282 diatom communities in brackish and saline lakes in Antarctica are highly
283 similar to those occurring in the Southern Ocean. This complicates the
284 delineation between marine and lacustrine sediments based on diatoms
285 alone. The stratigraphic data were plotted using Tilia and Tilia Graph
286 (Grimm 2004).
287

288 3.3. Radiocarbon dating

289 Lake sediment samples and marine macrofossils were dated using AMS ^{14}C
290 by the UK Natural Environment Research Council Radiocarbon Laboratory
291 (NERC) or the Beta Analytic Radiocarbon Dating Laboratory (Table S1).
292 Where possible, discrete macrofossils were dated (i.e. cyanobacterial mats,
293 worm tubes, sponge spicules or shells). The results are reported as
294 conventional radiocarbon years BP with one-sigma (1σ) standard deviation
295 error. The raised beach data were calibrated using the Marine13.14C
296 calibration curve in CALIB (Reimer et al. 2013; Table S1). The dates from
297 the marine sections in the sediment cores were calibrated using the mixed
298 terrestrial SHCal13.14C and the marine13.14C calibration curve, and those
299 of the lacustrine sediments using the terrestrial SHCal13.14C calibration
300 curve (Hogg et al. 2013). No reservoir correction was applied to dates from
301 lacustrine sediments, because surface-sediment dates indicate that ^{14}C in the
302 modern lakes are in near-equilibrium with modern atmospheric CO_2 (Table
303 S1), which is in agreement with results from other East Antarctic oases
304 (e.g., Hodgson et al. 2001, Verleyen et al. 2011). In contrast, the AMS ^{14}C
305 dates of the marine sediments and marine fossils in the raised beaches were
306 calibrated in CALIB 7.1 (Reimer et al. 2013) using a Delta R of 720 years,
307 leading to a total correction of 1120 years as recommended for the region

308 (Yoshida & Moriwaki 1979). An error of ± 100 years for the reservoir
309 effect was calculated based on the Yoshida & Moriwaki (1979) dates. The
310 ^{14}C dates of the sediments in the transition zone between the marine and
311 lacustrine sediments in the isolation lakes were calibrated using the mixed
312 Marine and SH Atmosphere calibration curve, with the percentage of
313 marine carbon taken into account for calculating the Delta R. The
314 percentage of marine carbon was set equal to the total relative abundance of
315 marine diatoms following the procedures detailed in Sterken et al. (2012).
316 The published ^{14}C dates from isolation lakes (Tanako et al. 2012) and raised
317 beach data (Miura et al. 1998) were recalibrated following the procedures
318 described above. Because no diatom data were available for constraining
319 the marine to lacustrine transitions in the cores of Tanako et al. (2012), the
320 amount of marine carbon was set at 100% in the calibration procedure for
321 those samples that were situated in the marine sediments and the transition
322 zone from marine to lacustrine sediments. For developing the RSL curve,
323 calibrated median ages were used and the upper and lower limit of the
324 calibrated ^{14}C dates defined the error bars.

325

326 3.4. Identifying RSL high stands and calculations of RSL fall

327 Minimum RSL high stands and their timing were defined based on the sill
328 height of isolation lakes and ^{14}C dates of their marine sediments, or on the
329 height of marine raised beaches and the ^{14}C ages of incorporated marine
330 fossils. We treat these constraints as minimum marine limits because it is
331 possible that marine sediments are present at higher altitudes, but not
332 surveyed. The maximum RSL limits were identified based on ^{14}C dates of
333 lacustrine sediments in glacial (always above RSL) and isolation lakes
334 (within the range of RSL changes) and their sill heights. The rate of RSL
335 fall was calculated by dividing the difference of the sill heights of two
336 isolation lakes situated above each other in the RSL curve by the difference
337 between the dates since the lakes were isolated. The dates since isolation
338 were determined from the calibrated ^{14}C ages of the first lacustrine sample
339 overlying the marine sediments in these basins. The rate of RSL fall
340 between the lowest lake and the present-day sea level was calculated by
341 dividing the sill height of this lake by its isolation date.

342

343 3.5. Glacial Isostatic Adjustment modelling

344 A GIA model was used to calculate predicted RSL curves for the four ice-
345 free regions. Each of the four peninsula and island sub-areas are small
346 enough (max 16 km across) that the variation in predicted RSL within them

347 would be smaller than the uncertainty in the observations. Therefore, a
348 single RSL prediction is provided for each island and peninsula area, and
349 the sea-level indicators for that location may be combined into a single RSL
350 curve. In contrast, the distance between the outcrops across the whole study
351 area is large enough for there to be a gradient in GIA. This, combined with
352 the differing distances of the islands and peninsulas from former ice loading
353 centres, justifies the need for a different RSL prediction for each outcrop.
354 The GIA model calculates the solid Earth response to ice and ocean loading
355 through time, and the corresponding change in the shape of the geoid
356 (Kendall et al. 2005). The Earth is represented by a three-layer, spherically-
357 symmetric, viscoelastic Maxwell body, while the ice loading history is
358 defined by either the W12 (Whitehouse et al. 2012a) or the ICE-6G_C
359 (Argus et al. 2014) model. The W12 model is combined with the northern
360 hemisphere component of the ICE-5G model (Peltier 2004) such that both
361 ice models define the global change in ice loading throughout the last
362 glacial cycle. Ocean loading is determined by solving the sea-level equation
363 (Farrell and Clark 1976). In combination with the W12 model we use the
364 optimum Earth model of Whitehouse et al. (2012b), which comprises a 120
365 km-thick lithosphere, an upper mantle of viscosity 10^{21} Pa s, and a lower
366 mantle of viscosity 10^{22} Pa s. In contrast, the ICE-6G_C ice loading history

367 should be combined with the VM5a Earth model (Peltier et al. 2015). The
368 VM5a model does not take a uniform viscosity value in the lower mantle
369 (Peltier et al. 2015), so we use an approximation of this model that has a 96
370 km-thick lithosphere, an upper mantle of viscosity 0.5×10^{21} Pa s, and a
371 lower mantle of viscosity 3×10^{21} Pa s. From here onwards we use the
372 terms W12 model and ICE-6G_C model to refer to the combination of the
373 ice and Earth model in each case. RSL predictions are extracted from the
374 models at the four study sites.

375

376 **4. Results**

377 **4.1. Paleolimnological proxy analyses of the sediment cores**

378 4.1.1. Isolation lakes

379 4.1.1.1. Yumi Ike (WO1), West Ongul Island - 10 m above sea-level (a.s.l.)

380 In the Yumi Ike core (Fig.2) a marine zone (WO1-I), a lacustrine
381 freshwater zone (WO1-III), and a transition zone (WO1-II) in between
382 could be identified based on the proxy data. Between 74 and 54 cm core
383 depth, marine diatoms dominated and the total carbon (TC) concentration
384 was relatively low. Mass-specific magnetic susceptibility (MS) values
385 decreased towards the end of this zone whereas gamma ray density (GRD)
386 remained relatively stable. The transition zone between 54 and 46 cm

387 contained a mixture of brackish-water and marine diatom species. The TC
388 concentration remained low. MS values slightly increased, whereas GRD
389 remained stable. From 46 cm until the surface sediments, freshwater
390 diatoms were dominant and brackish and marine diatoms occasionally
391 occurred. The TC concentration was more variable than in the other two
392 zones. MS values further increased to reach a maximum at 37.2 cm,
393 decreased until 14 cm, and rose again. GRD remained relatively stable to
394 become slightly higher in the upper 5 cm of the sediments.

395

396 4.1.1.2. Ô-Ike (WO4), West Ongul Island - 13 m a.s.l.

397 Similar to Yumi Ike, three main zones were identified in the Ô-Ike
398 sediment core (Fig. 3), namely a marine zone (WO4 I), a lacustrine
399 freshwater zone (WO4 III) and a very short transition zone in between
400 (WO4 II). In zone WO4 I, between 176 and 160 cm, TC concentrations
401 were low, while GRD and MS were relatively high. The latter decreased
402 towards the end of this zone. This zone was dominated by marine diatoms,
403 while freshwater species were absent. Between 160 and 158 cm, TC
404 concentrations were still low. This zone was dominated by marine and
405 brackish water diatoms. GRD and MS decreased throughout this zone.
406 Between 158 cm and the top of the core, the TC concentration was

407 relatively high. WO4 III was dominated by freshwater diatoms. GRD
408 remained relatively stable and was lower in this zone compared with zone
409 WO I and WOII until 86.6 cm, above which no measurements were
410 available. MS was low and stable throughout this zone.

411

412 4.1.1.3. Mago Ike (SK1), Skarvsness - 1.5 m a.s.l.

413 Again, three main zones were identified in the core from Mago Ike (Fig.4),
414 namely a marine zone (SK1 I), a lacustrine freshwater zone (SK1 III) and a
415 transition zone in between (SK1 II). Between 254 and 143 cm, the TC
416 concentration was very low. GRD and MS were relatively high and the
417 latter increased towards the end of the zone. Marine diatoms dominated,
418 while brackish-water and particularly freshwater species were only present
419 in low abundances. Between 143 cm and 123 cm TC started to increase.
420 GRD decreased in SK1 II while MS reached a maximum and subsequently
421 dropped sharply. The relative abundance of brackish-water diatoms
422 increased towards the upper part of this zone, while the percentage of
423 marine diatoms decreased. Between 123 cm and the top of the core, TC
424 concentration was relatively high, while GRD and MS were relatively low.
425 This zone was dominated by freshwater diatoms; some brackish-water and
426 marine diatoms occasionally occurred at the beginning of this zone.

427

428 4.1.1.4. Kobachi Ike (SK4), Skarvsness - 28 m a.s.l.

429 The evolution of Kobachi Ike is more complex and the delineation between
430 the different zones in the core was less straight forward compared with the
431 other isolation basins. This is due to the gradual change in the abundance of
432 brackish water versus marine diatoms and the presence of the latter in the
433 entire core, resulting in a slow species turnover in the fossil communities.
434 Based on the diatoms, pigments and sedimentological changes, the
435 sediment core could be subdivided in three main zones (Fig.5), namely a
436 zone consisting of glacial sediments (SK4 I), and a marine zone (SK4 II),
437 which gradually evolved towards a lacustrine zone (SK4 III). Between 280
438 and 245 cm, the total chlorophyll and total carotenoid concentrations as
439 well as the relative abundance of cyanobacterial carotenoids, MS and total
440 diatom concentration were low. From 260 cm onwards, zone SK4 I was
441 further characterized by relatively high TOC concentrations.
442 Myxoxanthophyll, a cyanobacterial marker pigment was absent throughout
443 this zone. Between 245 and 115 cm, the TOC concentrations, and the total
444 chlorophyll and carotenoid concentrations were low. Myxoxanthophyll was
445 almost completely absent in zone SK4 II. This zone was furthermore
446 characterized by relatively high MS values. Marine diatoms were dominant,
447 but brackish-water species became more abundant from c. 165 cm depth. It

448 follows that lake isolation may have started in this zone already. In zone
449 SK4 III, between 115 cm and the top of the core, the TOC, chlorophyll and
450 carotenoid concentrations were relatively high. Myxoxanthophyll became a
451 subdominant pigment which marks the presence of cyanobacteria. From 93
452 cm depth, brackish diatoms generally dominate.

453

454 4.1.2. Glacial lakes

455 All the samples analysed in the cores from Ura Ike (17 m a.s.l.; WO5),
456 Higashi Ike (18 m a.s.l.; WO6) and Nishi Ike (23 m a.s.l.; WO8) in the
457 Ongul Islands were dominated by freshwater lacustrine diatoms. Hence,
458 these lakes were considered to be of glacial origin. The basal ages of the
459 Higashi Ike and Nishi Ike sediment cores are c. 4520 or 4560 and c. 11,240
460 cal. yr BP, respectively. In Ura Ike, age reversals occurred between 73 and
461 59 cm (Table S1), making it difficult to determine the age of the bottom
462 sediments. However, the oldest ^{14}C date obtained suggests that Ura Ike is at
463 least c. 6,290 cal. yr BP old.

464

465 4.2. Initial ice sheet retreat

466 The start of biogenic sedimentation in the lacustrine sediments of glacial
467 lakes and marine sedimentation in isolation basins provides minimum ages

468 of initial ice sheet retreat over the terrestrial and nearshore marine
469 environment respectively (cf. Hodgson et al. 2001 and Verleyen et al. 2004;
470 Table S1). The latter were combined with ^{14}C dates of marine fossils in
471 raised beaches (Miura et al. 1998). In Skallen, Skarvsnes, and Langhovde
472 no glacial lakes were cored. In the most southerly peninsula, Skallen, the
473 oldest marine ^{14}C date was derived from a fragment of a shell in a raised
474 beach at 7 m a.s.l. and is 7,580 cal. yr BP, while the oldest date of marine
475 sediments in the Skallen Ike basin (9.6 m a.s.l.) is 5,810 cal. yr BP (Miura
476 et al. 1998; Fig.6a; Table S2). In Skarvsnes, polychaete tubes in a raised
477 beach at 18 m a.s.l. are 8,670 cal. yr BP old (Fig. 6b) while the oldest date
478 in a marine sediment core sequence comes from the isolation lake Kobachi
479 Ike (28 m a.s.l.), and is 7,430 cal. yr BP old (Fig. 6b; Table S1). The oldest
480 Holocene marine ^{14}C date in Langhovde is 10,390 cal. yr BP and was
481 derived from a shell of *Adamussium colbecki* situated in a raised beach at 6
482 m a.s.l. (Miura et al. 1998; Fig. 6c). The basal age of the freshwater
483 sediment cores from Nishi Ike (23 m a.s.l.) in the Ongul Islands is almost
484 1000 years older (i.e., 11,240 cal. yr BP), which agrees well with the oldest
485 post-LGM date of a marine fossil (shell fragment) in raised beaches at 17 m
486 a.s.l. on these islands (10,810 cal. yr BP; Miura et al. 1998; Fig.6d).
487

488 **4.3. Regional differences in relative sea-level changes**

489 The analyses of fossil diatoms and the sedimentology in all cores, in
490 combination with fossil pigments in Kobachi Ike, revealed that a total of 26
491 radiocarbon dates from the lake sediment cores were of marine or mixed
492 marine-lacustrine origin, while 39 were deposited in a lacustrine
493 environment (Fig.2-5; Table S1). Combined with the ^{14}C dates of the raised
494 beaches, these ages show that the RSL changes of Skallen, Langhovde and
495 the Ongul Islands were broadly similar, but differed markedly with the one
496 from Skarvsnes (Fig.6a-d). In Skallen, the minimum recorded sea-level
497 high stand is 12 m at c. 4,020 cal. yr BP based on the raised beach data.
498 RSL fall equalled no more than 3.7 mm/yr on average and was higher than
499 2.9 mm/yr during the past c. 2,600 cal. yr BP as revealed by the first ^{14}C
500 date in the lacustrine and the last deposited marine sediments respectively
501 in Lake Skallen. In Langhovde, no lake records are available preventing the
502 calculation of a robust rate of RSL fall. Based on the raised beach data
503 alone, the minimum marine limit was estimated to be 17 m at 6,530 cal yr
504 BP. In West Ongul Island, the maximum marine limit was below 17 m after
505 6,288 cal. yr BP as indicated by the absence of ^{14}C dates with a marine
506 origin in Ura Ike, and never exceeded 23 m during the past 11,240 cal. yr
507 BP based on the presence of exclusively lacustrine sediments in the Nishi

508 Ike basin. The raised beach data revealed that the minimum marine limit on
509 the islands is 17 m at 10,813 cal. yr BP (Fig.6d; Table S2). RSL fall
510 equalled on average 2.5 mm/yr during the past c. 5,160 cal. yr BP and 2.3
511 mm/yr during the past c. 4,360 cal. yr BP based on the isolation of Yumi
512 Ike. In Skarvsnes, the minimum RSL high stand is 32.7 m based on a new
513 radiocarbon date of a marine macrofossil (shell) of $5,410 \pm 40$ ^{14}C yr old
514 (5,265 – 4,653 cal. yr BP) preserved in a raised beach in the upper sill of
515 Kobachi Ike (Table S1). The other macrofossils for which new ^{14}C dates are
516 available are from *L. elliptica* and polychaete tubes preserved in raised
517 beaches in the valley which is occupied by L. Suribati to the north east of
518 Kobachi Ike at a height of 8.6 m a.s.l. and they are respectively $4,730 \pm 40$
519 and $6,800 \pm 40$ ^{14}C yr old (Table S1). In Skarvsnes, RSL fall was more
520 rapid during the past 2,410 cal. yr BP than in the Ongul Islands and Skallen,
521 and equalled on average 11.6 mm/yr. The dominance of brackish diatoms at
522 93 cm and the presence of the cyanobacterial pigment myxoxanthophyll
523 (from 115 cm onwards) in the Kobachi Ike sediment core are used to infer
524 lacustrine conditions (Fig.5), and hence lake isolation in this calculation.
525 Between c. 2,410 (first lacustrine ^{14}C date in Kobachi Ike (28 m a.s.l.)) and
526 780 cal. yr BP (first lacustrine ^{14}C date in Mago Ike; 1.5 m.a.s.l.), the mean
527 rate of RSL fall was 16.2 mm/yr, but this dropped to a rate of 1.9 mm/yr

528 from c. 780 cal. yr BP onwards, which is of the same order as that recorded
529 in the other two regions. The inference of the start of freshwater conditions
530 during the Late Holocene in Kobachi Ike also shows that RSL did not fall
531 below 28 m a.s.l. until 2,410 cal. yr BP (Fig.6b).

532

533 **4.4. Ice sheet model outputs and comparison with geological constraints**

534 The maximum RSL high stand in the output of the W12 model is
535 consistently lower and occurs slightly later compared with the ICE-6G_C
536 model, although the difference between the two models decreases with
537 distance from the Shirase Glacier (Fig.6a-d). Along the south to north
538 gradient away from the Shirase Glacier (i.e., between Skallen and the
539 Ongul Islands), the maximum RSL high stand varied between c. 29 and
540 20.3 m and between c. 14.3 and 12.4 m in the output of the ICE-6G_C and
541 W12 models, respectively. The output of the W12 model provides a
542 reasonable fit to the highest radiocarbon date of a marine raised beach
543 sample in Skallen, although this was not necessarily the marine limit. This
544 model also agreed well with the geological constraints on the RSL high
545 stand in the Ongul Islands, but underestimates the RSL high stand in
546 Langhovde and particularly in Skarvsnes. The rate of RSL fall during the
547 Late Holocene is underestimated by this model in all four regions and

548 particularly in Skarvsnes. With the exception of the Ongul Islands, this is
549 also more or less the case with the output from the ICE-6G_C model which
550 underestimates RSL fall in the three other regions. The high stand is
551 predicted by the ICE-6G_C model to lie above the elevation of the highest
552 marine fossils in Skallen and Langhovde, although these fossils were not
553 necessarily sampled at the maximum marine limit. The ICE-6G_C model
554 provides a good fit to the raised beach and lake data in the Ongul Islands
555 and gets closer to matching the highest marine fossils at Skarsvnes.
556 However, in the latter region the timing of the modelled RSL high stand is
557 too early compared with the geological constraints from Kobachi Ike.

558

559 **5. Discussion**

560 **5.1. Interpretation of the proxy results in the lake sediment cores**

561 Delineating between marine and lacustrine sediments in three out of the
562 four isolation basins was relatively straightforward based on the presence of
563 diatom indicator taxa (Fig.2, 3, 4). The occasional occurrence of marine
564 diatoms in the lacustrine zones of the cores from for example Yumi Ike is
565 likely the result of sea spray or the visit of the lake by marine birds or
566 mammals as was observed during sampling in Langhovde. However, in
567 Kobachi Ike, marine diatoms were present in all zones of the sediment

568 cores and the abundance of brackish diatoms gradually increased until 20
569 cm after which they declined again (Fig.5). This gradual change in diatom
570 community structure is likely related to the volume and shape of the basin
571 in relation to the amount of meltwater entering the lake. In the other study
572 lakes, the meltwater input is high compared with the volume of the basin,
573 leading to flushing of the trapped marine water after lake isolation, which in
574 turn resulted in the establishment of freshwater conditions and the
575 colonization by freshwater organisms (including diatoms). By contrast, in
576 Kobachi Ike, the relatively low amount of meltwater entering the lake only
577 slowly diluted the marine water. Moreover, due to the relatively deep water
578 column, the lake is chemically stratified as brackish conditions prevail in
579 the bottom waters (specific conductance below 2.4 m depth equaled 11.4
580 mS/cm at the time of sampling), while low salinity waters (specific
581 conductance of 5.0 mS/cm) were present in the upper 2.4 m of the water
582 column. This freshwater lens at the surface is likely derived from meltwater
583 input from the catchment and/or lake ice (Kimura et al. 2010). The salinity-
584 driven stratified conditions appear to be strong enough to prevent mixing of
585 the bottom water with this meltwater. Furthermore, this situation also
586 provides a mechanism for the passage of large fluxes of meltwater without
587 significantly affecting the salinity of the lake as freshwater can pass through

588 the epilimnion and leave the lake via an outflow stream (which was not
589 active during sampling) without diluting the brackish water stored in the
590 hypolimnion. Hence, instead of the relatively rapid dilution of the lake
591 water in the smaller polymictic freshwater lakes and the subsequent
592 changes in the diatom communities, marine species could probably survive
593 in saline conditions in Kobachi Ike for hundreds of years. This was for
594 example also the case in the saline lakes of the Vestfold Hills (Roberts and
595 McMinn 1999), which are still dominated by marine taxa (Verleyen et al.
596 2003). In turn, this complicates the delineation of the core into marine and
597 lacustrine zones. In Kobachi Ike, we therefore combined fossil diatoms
598 with fossil pigments and changes in the sediment properties to pinpoint the
599 isolation event. At 115 cm depth, myxoxanthophyll becomes a subdominant
600 pigment. Myxoxanthophyll is present in benthic cyanobacteria, which
601 dominate the primary production in microbial mats in the benthos of East
602 Antarctic lakes (Verleyen et al. 2010), as well as Kobachi Ike today
603 (Obbels et al. unpubl. results). However, cyanobacteria are largely absent
604 from the Southern Ocean (Fukuda et al. 1998). We therefore considered the
605 zone between 115 cm and 93 cm as a transition zone, in which benthic
606 cyanobacteria occurred but marine diatoms remained dominant. Hence, the
607 ^{14}C dates at 115 and 107 cm were calibrated using the mixed marine and

608 SH curve (Table S1). From 93 cm depth brackish diatoms generally
609 dominate. We interpret this as the start of the establishment of fully
610 lacustrine conditions. However, spores from marine *Chaetoceros* species
611 remained an important member of the assemblages and even dominated in
612 some samples in the upper 20 cm. These spores can be *in situ* produced,
613 although it is also possible that they were transported to the lake through
614 sea spray, or alternatively that they were washed-in from raised beach
615 deposits within the catchment area. The start of the dominance of the
616 brackish water diatoms also coincided with a decrease in magnetic
617 susceptibility (MS) that further gradually declined from 82 cm. This
618 decrease in MS also suggests a complete isolation of the lake, which was
619 for example similarly observed in Maritime Antarctic lakes and related to
620 differences in the sedimentary infill of the basins during marine versus
621 lacustrine conditions (Watcham et al. 2011). During the latter, mainly local
622 minerals are transported to the basin while during marine conditions
623 sediments from elsewhere might be transported to the site via ice bergs and
624 redistributed sea ice containing wind-blown particles. Hence, we
625 considered the start of the dominance by brackish water diatoms at 93 cm
626 depth as marking the establishment of full lacustrine conditions.

627 The absence of marine sediments in the cores from Ura Ike (17 m
628 a.s.l.; WO5), Higashi Ike (18 m a.s.l.; WO6) and Nishi Ike (23 m a.s.l.;
629 WO8) in the Ongul Islands suggests that these basins were situated above
630 the marine limit throughout the entire Holocene and probably originated
631 from beneath the ice-sheet or permanent snow fields during the Early- to
632 Mid-Holocene.

633

634 **5.2. Initial ice sheet retreat**

635 The finding that all dates obtained from the lake sediment cores are of
636 Holocene age suggests that the regions were ice-covered during the LGM
637 as a result of the expansion of the EAIS, and that they became gradually
638 ice-free during the Early Holocene. This scenario is in general agreement
639 with reconstructions in a large number of the currently ice-free regions in
640 East Antarctica, such as Schirmacher Oasis, the Vestfold Hills (but see
641 Gibson et al. 2009), and the Windmill Islands (see Hall 2009 and
642 Mackintosh et al. 2014 for a review).

643 The ^{14}C dates in the bottom sediments of the lakes also suggest that
644 deglaciation started later near the Shirase Glacier (in Skallen and
645 Skarvsnes) than in the regions further to the north (Langhovde and the
646 Ongul Islands). More in particular, the oldest ^{14}C date in Skallen was 7,580

647 cal yr BP and the oldest date (c. 11,240 cal. yr BP, see Table S1) was
648 obtained in lacustrine sediments overlying glacial sediments in a core from
649 Nishi Ike, a glacial lake in West Ongul Island. This confirms the prediction
650 that regions closer to the main glacier deglaciated more recently than those
651 further to the north. We are however aware that the ages are only minimum
652 ages for deglaciation, and that deglaciation potentially started more or less
653 coincident in the different ice-free regions. However, our lake based
654 estimates of the minimum age of deglaciation in Skarvsnes confirm existing
655 reconstructions of the deglaciation history based on raised beach data
656 (Miura et al. 1998), as well as cosmogenic isotope dates (Yamane et al.
657 2011). More in particular, deglaciation in Skarvsnes seems to have started
658 somewhere around c. 7430 cal. yr BP, as evidenced by the oldest
659 radiocarbon date obtained from the marine sediments in Kobachi Ike. This
660 timing is in agreement with that obtained from the radiocarbon dates in the
661 raised beaches (Miura et al. 1998, Fig.6b; Table S2), where apart from two
662 dates, none is older than c. 8000 cal. yr BP. Moreover, our estimate also
663 corresponds to a cosmogenic isotope dating study which places the
664 deglaciation of Skarvsnes between 10 and 6 ka BP (Yamane et al. 2011).
665 More precisely, the time of deglaciation of the Kobachi Ike basin agrees
666 well with that obtained for nearby Mount Suribati. A relatively late

667 deglaciation in Skarvsnes and Skallen furthermore corroborates recent
668 evidence from regions along the Rayner Glacier (Enderby Land) to the east
669 of Lützow-Holm Bay that became ice-free between 9 and 6 ka (White &
670 Fink 2014).

671 However, the scenario of an early Holocene deglaciation in the
672 Ongul Islands contradicts an alternative interpretation which was based on
673 existing raised beach data (Takada et al. 2003). More in particular, because
674 well-preserved *in situ* fossils of *L. elliptica* in raised beaches from the
675 Ongul Islands and parts of Langhovde predate the LGM (Miura et al.
676 1998), Takada et al. (2003) suggested that the nearshore zone of those
677 regions were ice-free during MIS3 and maybe even during earlier marine
678 isotope stages. One hypothesis to explain the discrepancy between the
679 presence of *in situ* fossils of Late Pleistocene age and the lack of lake
680 sediments predating the Holocene is that terrestrial habitats were covered
681 with permanent snow banks during the LGM. This snow cover would have
682 prevented light penetration and hence primary production in the lakes
683 during the LGM (cf. Gore 1997). In turn, this blanketing by snow would
684 have resulted in the absence of organic carbon in terrestrial habitats and
685 hence the lack of material for ^{14}C dating. In this scenario, the Ongul Islands
686 and parts of Langhovde escaped glacial overriding during the LGM, and the

687 expanding glacier was thus diverted around the regions, possibly through
688 the 600 m deep Fuji Submarine Valley. By contrast, the regions closer to
689 the Shirase Glacier only became ice-free during the Holocene (Mackintosh
690 et al. 2014). These regional differences in deglaciation in Lützow-Holm
691 Bay are furthermore supported by geomorphological evidence and the
692 degree of weathering of the bedrock. Indeed, rocks in the northernmost part
693 of Sôya Coast are deeply weathered, whereas those in the southern part of
694 the coast (i.e. Skarvsnes and Skallen) are relatively unweathered and
695 intensively striated. However, regional differences in the degree of
696 weathering not necessarily require ice-free conditions during the LGM in
697 the Ongul Islands. Instead, these differences can be equally explained by
698 the presence of a cold-based and slow moving ice sheet which was
699 buttressed on the Ongul Islands, while the major ice flow lines diverted into
700 the deep glacial troughs between the islands and the continent. The ice
701 sheet could instead have been more active in the areas closer to the current
702 glacier front leading to intensively striated bedrock. Besides, this could also
703 explain the presence of *in situ* marine fossils of Pleistocene age in the
704 Ongul Islands and parts of Langhovde (Miura et al. 1998). A similar
705 process was proposed by Hodgson et al. (2006b) to invoke the presence of
706 well-preserved Eemian sediments in Progress Lake in the Larsemann Hills,

707 which became ice-free during the Late-Holocene. It is however clear that
708 additional ^{14}C dates of lake sediment cores in combination with cosmogenic
709 isotope dates of landforms are needed to assess whether the Ongul Islands
710 and parts of Langhovde were indeed ice-free during the LGM or rather
711 covered by an inactive ice sheet.

712

713 **5.3. Geological constraints on changes in relative sea-level**

714 Our most significant finding is the striking difference in the RSL high
715 stands and rates of RSL fall between Skallen, Langhovde and the Ongul
716 islands on the one hand, and Skarvsnes on the other (Fig.6a-d). In Skallen,
717 the raised beach data suggest that the RSL high stand was situated at least
718 at 12 m. It is possible that the limit was actually higher, but this needs to be
719 confirmed by additional dating of bottom sediments of glacial lakes (i.e.
720 those that have remained above the Holocene marine limit) and additional
721 surveying of raised beaches in the region at higher altitudes. In the Ongul
722 Islands, RSL was always below 23 m a.s.l. during the Holocene as
723 indicated by the presence of exclusively lacustrine sediments in the glacial
724 lake Nishi Ike between c. 11,240 cal. yr BP until present. The absence of
725 raised beaches 6 m below the sill height of this lake and the absence of
726 marine sediments in the two other glacial lakes (Ura Ike at 17 m a.s.l. and

727 Higashi Ike at 18 m a.s.l.) suggests that the marine limit in the Ongul
728 Islands is probably even lower (i.e., at 17 m a.s.l.). It is however not
729 completely sure whether RSL was below 17 m.a.s.l. during the early
730 Holocene, because the oldest ages obtained in Ura Ike and Higashi Ike were
731 respectively only 6288 cal yr BP and 4596 cal yr BP (Table S1). In
732 Langhovde, the raised beach data suggest that the marine limit is similarly
733 at 17 m a.s.l. Taken together, these marine limits are close to previous
734 estimates based on raised beach data alone (Miura et al. 1998). By contrast,
735 the minimum marine limit in Skarvsnes is at least 9 m higher than the
736 maximum marine limit in the Ongul Islands, and 12 m higher than previous
737 estimates for the peninsula based on raised beach data alone (Miura et al.
738 1998). The rate of RSL fall is also different between Skarvsnes and the
739 other three regions. In Skarvsnes, RSL fall was on average 11.6 mm/yr
740 during the past 2,400 years. This far exceeds the rates in Skallen and the
741 Ongul Islands, which equalled 3.6-2.9 mm/yr during the past c. 2,600 cal.
742 yr BP and 2.5 mm/yr during the past c. 5,160 cal. yr BP, respectively. The
743 shape of the RSL curve is also highly different compared with those in
744 other regions along the East Antarctic coastline (e.g. Zwartz et al. 1998,
745 Verleyen et al. 2005). This difference is mainly related to the rapid RSL fall
746 between 2,400 cal yr BP (isolation of Kobachi Ike) and 780 cal. yr BP

747 (isolation of Mago Ike) in Skarvsnes. These contrasts in the RSL curves in
748 the different regions are potentially underlain by three different, non-
749 mutually exclusive processes, namely regional variation in (i) the timing of
750 deglaciation, (ii) local ice-sheet volume, and (iii) neotectonic processes.
751 The first process is less likely, given the relatively small regional
752 differences in the timing of the start of deglaciation between Skallen and
753 Skarvsnes. Also, the second process can be expected to be negligible,
754 because RSL changes typically reflect regional changes in ice thickness
755 rather than local small-scale differences. GIA could only produce such a
756 spatial contrast in RSL rate if the upper mantle were locally very weak (e.g.
757 Simms et al. 2012) and there had been a short-lived, localised period of
758 significant ice loss in Skarvsnes. There is no evidence for either condition
759 being upheld. We therefore speculate that the third hypothesis, namely that
760 neotectonic processes are involved, is the most likely, given (i) the small
761 distance between the different sites, (ii) the marked difference in the shape
762 of the RSL curve in Skarvsnes with that in the other regions in Lützow-
763 Holm Bay and elsewhere in Antarctica (e.g., Hodgson et al. 2016), (iii) the
764 presence of a mapped fault system on Skarvsnes and other faults in the bay
765 (Ishikawa et al. 1976; Fig.1), and (iv) the well-known tendency for post-
766 glacial crustal stress to result in fault rupture in some locations (Bentley &

767 McCulloch 2005, Steffen et al. 2014). A reactivation of this fault system in
768 response to glacial unloading could explain the sudden difference in RSL
769 fall in Skarvsnes between c. 2400 and 780 cal. yr BP (rate of 16.2 mm/yr)
770 compared with a rate of 1.9 mm/yr from c. 780 cal. yr BP onwards. Short-
771 term tectonic activities along existing fault lines was also hypothesised to
772 explain regional patterns in RSL fall along the Baltic coast of Sweden
773 (Risberg et al. 2005). Similarly, in the Strait of Magellan (South Chile)
774 there is evidence for post-glacial fault movement of at least 30 m, based on
775 the proxy record from a bog near Puerto del Hambre and the regional
776 history of proglacial lakes (Bentley & McCulloch 2005). On account of the
777 differences in RSL changes between the islands and peninsulas in Lützw-
778 Holm Bay we consider that the three similar records (Skallen, Langhovde,
779 Ongul) can be used to constrain GIA models, but that Skarvsnes should be
780 considered an outlier. This could be confirmed by further geological and
781 geomorphological data from either side of the fault lines.

782

783 **5.4. Comparison between geological constraints and monitoring and** 784 **modelling results**

785 The rate of RSL fall in Skallen and the Ongul Islands, which equalled 3.6
786 mm/yr on average during the past c. 2,600 cal. yr BP and 2.5 mm/yr on

787 average during the past c. 5,160 cal. yr BP respectively, is comparable with
788 data obtained from short-term GPS measurements of local crustal
789 deformation between 1999-2003 in Skallen (3.00 +/- 1.9 mm/yr; 69.6710 S,
790 39.3987 E) and between 1998 and 2004 (2.56 +/- 0.24 mm/yr; Ohzono et al.
791 2006) in West Ongul Island (69.0070 S, 39.5833 E). In Skarvsnes the rate
792 of RSL fall is 1.9 mm/yr from c. 780 cal. yr BP onwards, which is in
793 relatively good agreement with the uplift rate measured using GPS
794 monitoring stations in the region (1.12 +/- 1.46 mm/yr, 69.4738 S, 39.6071
795 E; Ohzono et al. 2006). This confirms the robustness of our approach.
796 However, ignoring the anomalous curve before c. 780 cal yr BP at
797 Skarvsnes, the shape and high stand of the RSL curves based on the
798 geological data are not always in agreement with GIA modelling results.
799 For example, the ICE-6G_C model provides a reasonable fit to the recent
800 rate of RSL fall at Skallen but this rate is under-predicted by the W12
801 model. Both models under-predict the recent rate of RSL fall at Langhovde,
802 but fit the data reasonably well in the Ongul Islands. The greater magnitude
803 of the high stand predicted by the ICE-6G_C model at all four locations is
804 due to a combination of two factors: (i) The ICE-6G_C model includes a
805 greater magnitude of regional ice loss since the LGM compared with the
806 W12 model, and (ii) it uses a weaker value for the upper and lower mantle

807 viscosity. The lack of robust, independent constraints on either of these
808 factors makes this an underdetermined problem. Regional RSL data
809 therefore play a vital role in reducing the uncertainty on ice history and
810 Earth rheology around Antarctica.

811

812 **6. Conclusions**

813 The minimum age for deglaciation of the Lützow Holm Bay region is c.
814 11,240 cal. yr BP on West Ongul Island with progressively younger
815 deglaciation ages approaching the main regional ice outflow at Shirase
816 Glacier. Based on our geological evidence, it remains unclear whether parts
817 of the region were ice-free during the LGM, or alternatively covered by
818 permanent snow banks or an inactive ice sheet. Of most significance is the
819 difference in (i) the Holocene RSL high stand and (ii) the shape of the RSL
820 curves in Skarvsnes compared with those in the Ongul Islands, Langhovde
821 and Skallen. We attribute these regional differences to neotectonic events.
822 Current GIA model predictions give a reasonable fit to the reconstructed
823 RSL curves at Skallen, Langhovde, and West Ongul, but they are unable to
824 explain the pattern of RSL recorded at Skarvsnes.

825

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841

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Figure captions

Fig. 1: Overview map of Antarctica with an indication of the study area and the ICESat7 drainage system of the East Antarctic Ice Sheet (Zwally et al. 2012), and a map of Lützow Holm Bay with an indication of the study sites: the Ongul Islands, Langhovde, Skarvsnes and Skallen. The inset shows the location of the lakes used for developing the RSL curves in Fig.6: Yumi Ike (WO1, 10 m a.s.l.), Ô-Ike (WO4, 13 m a.s.l.), Ura Ike (WO5, 17 m a.s.l.), Higashi Ike (WO6, 18 m a.s.l.), Nishi Ike (WO8, 23 m a.s.l.), Mago Ike (SK1, 1.5 m a.s.l.) and Kobachi Ike (SK4, 28 m a.s.l.). The lake codes refer to Tavernier et al. (2014) and Verleyen et al. (2012). The data for Lake Oyako (2.4 m a.s.l.) and Lake Skallen (9.6 m a.s.l.) are based on Takano et al. (2012).

Fig.2: Summary diagram of the Yumi Ike (WO1 – 10 m a.s.l.) sediment core showing the lithology, total carbon content (TC), mass specific magnetic susceptibility (MS), gamma ray density (GRD), and the percentage of lacustrine freshwater, brackish and marine diatoms. The dates are median calibrated ^{14}C ages. Dates in blue were calibrated using the mixed SH marine-terrestrial calibration curve and those in black using the SH Cal13 terrestrial calibration curve.

Fig.3: Summary diagram of the Ô Ike (WO4 – 13 m a.s.l.) sediment core showing the lithology and legend, total carbon content (TC), mass specific magnetic susceptibility (MS), gamma ray density (GRD), and the percentage of lacustrine freshwater, brackish and marine diatoms. GRD and MS were only measured on cores transported intact to the laboratory (between c. 176 and 86 cm depth). The color code for the dates is as in fig.2.

Fig.4: Summary diagram of the Mago Ike (SK1 – 1.5 m a.s.l.) sediment core showing the lithology and legend, total carbon content (TC), gamma ray density (GRD), mass specific magnetic susceptibility (MS), and the percentage of lacustrine freshwater, brackish and marine diatoms. The color code for the dates is as in fig.2. For depths for which two dates are available, the date of the bulk material is on the right and the date of macrofossils on the left. The dates on the marine macrofossils were consistently younger.

Fig.5: Summary diagram of the Kobachi Ike (SK4 – 28 m a.s.l.) sediment core showing the lithology and legend, total carbon content (TC), the total

chlorophyll and carotenoid concentration, the relative abundance of cyanobacteria marker pigments, and the percentage of myxoxanthophyll (%); a pigment exclusively produced by cyanobacteria. Also shown are the gamma ray density (GRD), mass specific magnetic susceptibility (MS), and the percentage of lacustrine freshwater, brackish and marine diatoms. The grey horizontal bar represents a zone of low diatom production. The green line represents the interpreted start of full lacustrine conditions based on the dominance of brackish water diatoms. The color code for the dates is as in fig.2.

Fig.6: Relative sea level curves for (a) Skallen, (b) Skarvsnes, (c) Langhovde and (d) the Ongul Islands; the order of the regions is in increasing distance from the Shirase Glacier. The plots show the height above present sea level (a.s.l.; grey stippled horizontal line) of the median calibrated ^{14}C dates of the marine fossils in the raised beaches (blue circles) extracted from Miura et al. (1998), the marine sediments in the isolation lakes (blue squares), and the lacustrine sediments in the glacial and isolation lakes (red squares), including the data extracted from Takano et al. (2012). The dark blue circles in fig.2b denote the new raised beach data. The red symbols represent the maximum upper limit of the RSL curve,

while the blue symbols are the minimum upper limit. The vertical error bar was set at 3 m corresponding to the maximum tidal range in the region (Aoyama et al. 2016) that exceeds the error of the measurements of the heights of the deposits. The horizontal error bars correspond to the minimum and maximum ranges of the calibrated ^{14}C dates. The green line is the output of the W12 model (Whitehouse et al. 2012a), and the black line is the output from our approximation of the ICE-6G_C model (Argus et al. 2014). The full blue line is a hand-drawn approximation of the minimum RSL based on the available ^{14}C dates of marine sediments in isolation basins or marine raised beaches.