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Oceanic forcing of Holocene ice-sheet retreat in the Amundsen Sea Embayment, West Antarctica

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25 Abstract

Glaciological and oceanographic observations coupled with numerical models indicate 26 that warm Circumpolar Deep Water (CDW) upwelling onto the West Antarctic continental 27 shelf causes melting of the undersides of floating ice shelves. Because the ice shelves are 28 buttressing glaciers feeding into them, their destabilisation through ocean-induced melting is 29 the predominant driver of current ice loss in Antarctica. Here we present the first data-based 30 reconstruction of Holocene (11,700 years ago to present) variability in CDW inflow to the 31 Amundsen Sea sector, the most vulnerable part of the West Antarctic Ice Sheet. The chemical 32 composition of calcareous foraminifer shells and assemblages of benthic foraminifera in 33 marine sediments indicate that enhanced CDW upwelling, controlled by the latitudinal position 34 of the Southern Hemisphere westerly winds, forced deglaciation of the Amundsen Sea sector 35 both during the early Holocene (11,700 to 7,500 years ago), when an ice-shelf collapse may 36 have caused rapid ice-sheet thinning further upstream, and since the 1940s. These results 37 increase confidence in the predictive capability of current ice-sheet models. 38

40 Today ice discharge from glaciers flowing into the Amundsen Sea Embayment (ASE) (Fig. 1) is the main contributor to the negative mass balance of the West Antarctic ice Sheet 41 (WAIS) and global sea-level rise from Antarctica¹. Rapid thinning, flow acceleration, and 42 grounding-line retreat affecting ice streams and ice shelves in this region have increased since 43 the early 1970s and contributed to global sea-level rise by, on average, ~0.23±0.02 mm/yr 44 between 1992 and 2013¹⁻³. Recent modelling suggests that current ice loss of the Thwaites 45 Glacier (Fig. 1), which rests on a bed that is well below sea level and slopes down into the 46 interior of the ASE hinterland, will result in a (partial) WAIS collapse from as early as the 23rd 47 century, thereby raising global sea level over centuries and millennia at rates $\geq 1 \text{ mm/yr}$ (ref.4). 48 A similar study predicts that neighbouring Pine Island Glacier, which drains together with 49 Thwaites Glacier into Pine Island Bay (PIB) in the eastern ASE (Fig. 1), will contribute ~3.5-50 10 mm to eustatic sea-level rise over the next two decades alone¹. Depending on the future 51 extent of atmospheric and oceanic warming, and the dominant processes of ice loss, Antarctica 52 may contribute as much as 1 metre to global sea-level rise by the end of this century⁶. 53

Ocean-driven melting of the undersides of floating ice shelves buttressing the glaciers 54 draining into the ASE has been identified as the main process responsible for the current mass 55 loss. This melting is caused primarily by upwelling of relatively warm Circumpolar Deep 56 Water (CDW) onto the continental shelf⁷⁻¹². Inter-annual to inter-decadal changes in CDW 57 upwelling onto the ASE shelf have been attributed to buoyancy changes⁹ and regional climate 58 variability¹⁰. The latter is assumed to be controlled by latitudinal shifts in the Southern 59 Hemisphere westerly winds (SHWW)¹¹, possibly in response to the Southern Annular Mode¹² 60 and teleconnections to climate in the tropical Pacific¹³. 61

While West Antarctic ice cores provide high-resolution archives of atmospheric 62 changes through the last glacial period to today¹⁴, satellite observations and *in situ* 63 measurements of glaciologic^{2,8} and oceanographic changes^{7,10} are restricted to the last \sim 30-40 64 years^{4,15}. Until now, no data constrain CDW upwelling onto the ASE shelf prior to this time, 65 and its role in driving centennial to millennial retreat of WAIS remains unclear. This 66 information, however, is crucial for verifying the reliability of the current generation of ice-67 sheet models which force WAIS deglaciation following the Last Glacial Maximum (LGM; 19-68 23 kyr BP) predominantly by ocean-induced melting^{16,17} and predict that this process will also 69 drive future Antarctic ice loss in response to global warming^{18,19}, at least during its initial 70 phase⁶. Until now, simulations of ocean forcing have predominantly relied on estimates of 71 deep-water temperatures derived from global climate models that are constrained by far-field 72 oceanic records from outside the Southern Ocean and its complex circulation system^{6,16-18}. 73

74 Foraminifer shell chemistry and ocean circulation

Here we present the first direct evidence for variations in CDW heat supply onto the 75 ASE shelf during the last 10.4 kyr. We measured the magnesium/calcium (Mg/Ca) ratio of 76 calcareous shells of the epifaunal to shallow infaunal benthic foraminifer species Angulogerina 77 angulosa (Methods) in core PS75/160 recovered just offshore from Pine Island Glacier²¹ (Fig. 78 1). The Mg/Ca data are used as a proxy for bottom-water temperature²⁰ and thus inflow of 79 CDW onto the ASE shelf (Methods, see details in Supplementary Information). The low 80 resolution Mg/Ca record of core PS75/160 is supported by a high-resolution stable carbon 81 isotope (δ^{13} C) data set obtained from calcareous shells of A. angulosa and the planktic 82 foraminifer species Neogloboquadrina pachyderma sin. from the same site and from site 83 PS75/167, which is located just offshore from Thwaites Glacier (Fig. 1). We utilize the benthic 84 δ^{13} C ratio as a proxy for the δ^{13} C composition of the dissolved inorganic carbon (δ^{13} C_{DIC}) of 85 the ambient seawater and thus, as a tracer for its source water mass. We also measured $\delta^{13}C$ 86 isotopes on A. angulosa in modern sediments from site PS75/215 in PIB and on N. pachyderma 87 sin. in recent sediments from site PS69/251 on the outer ASE shelf^{21,22} (Fig. 1). 88

For aminiferal $\delta^{13}C$ records from marine sediments are influenced by water-mass 89 circulation in addition to vital effects, biological productivity, air-sea exchange of carbon, or a 90 combination of these factors²³. We corrected the $\delta^{13}C$ data for species-inherent vital effects as 91 follows: The δ^{13} C signal of A. angulosa is known to be on average 0.44-0.60% lower than the 92 $\delta^{13}C_{DIC}$ of ambient seawater, but this offset varies regionally²⁴. Mackensen²⁵ analysed the $\delta^{13}C$ 93 composition of the epibenthic foraminifer genus Cibicidoides spp., a reliable recorder of 94 bottom water $\delta^{13}C_{DIC}$, in core-top sediments from site PS2544 on the outer ASE shelf (Fig. 1). 95 The measured δ^{13} C value of +0.63‰ overlaps with the δ^{13} C_{DIC} range of ~0.3 to 0.6‰ reported 96 for CDW^{26,27} when the analytical errors for δ^{13} C measurements of seawater (±0.1‰) and 97 for aminiferal calcite (±0.06‰) are taken into account²⁵. In contrast, we measured a $\delta^{13}C$ ratio 98 of -0.22‰ on A. angulosa from modern sediments at site PS75/215. Because both core sites 99 are currently bathed by CDW⁷⁻¹¹, we applied an offset correction of +0.85‰ to our benthic 100 δ^{13} C data. Previous studies on N. pachyderma sin. shells from south of the Antarctic Polar 101 Front²⁸ and south of 70°S²⁹, respectively, found a δ^{13} C offset of -1.0‰ between the shells and 102 the ambient surface waters. Accordingly, we corrected our planktic δ^{13} C data by adding +1.0%. 103

To evaluate the influence of past changes in biological productivity on the foraminiferal δ^{13} C down-core record, we measured barium/titanium (Ba/Ti) ratios of cores PS75/160 and PS75/167 (Fig. 2, Extended Data Fig. 5) as the most reliable palaeoproductivity proxy in Antarctic continental margin sediments³⁰. Furthermore, we examined the benthic foraminifer assemblages in sediments of core PS75/160 for evidence of environmental changes, especially whether an ice shelf covered PIB from ~12.3-11.2 kyr BP onwards^{31,32}. Ice-shelf cover would have reduced the foraminiferal δ^{13} C values by restricting air-sea gas exchange^{25,28}. Cosmogenic isotope surface exposure dates obtained from erratic boulders in the Hudson Mountains near Pine Island Glacier (Fig. 1) suggest that such an ice shelf had buttressed grounded ice upstream from PIB and that its thinning or collapse between 8.3 and 7.5 ka triggered rapid ice-stream thinning by >100 m (ref. 32).

Age constraints on cores PS75/160 and PS75/167 are based on previously published^{21,22} 115 and newly calibrated AMS ¹⁴C dates on calcareous microfossils (Extended Data Table 1). The 116 recent chronology for core PS69/251, whose modern surface age has been confirmed by core-117 top AMS ¹⁴C dating²¹, is constrained by a lead isotope (²¹⁰Pb) down-core profile (Extended 118 Data Figs. 2-4, Extended Data Table 2). Core PS75/160, which spans the time period from ~9.2 119 to 1.2 kyr BP (Fig. 2a), was retrieved from a local depression in PIB at a water depth of just 120 337 m. The relatively shallow location of this core explains the abundant occurrence of 121 calcareous microfossils, but another consequence is that the lowermost part of the core, older 122 than ~8.2 kyr BP, is affected by iceberg or ice-shelf keel ploughing²¹. We overcome this issue 123 by splicing the δ^{13} C data from core PS75/160 with those from nearby core PS75/167, which 124 span 10.4 to 8.2 kyr BP (Fig. 2b; Extended Data Fig. 1). Core PS75/167 does not contain 125 calcareous microfossils in sediments younger than ~8.2 kyr BP. In the following we refer to 126 127 the spliced record as the "PIB record".

The benthic Mg/Ca ratios in core PS75/160 imply higher bottom-water temperatures 128 129 before 7.5 kyr BP, indicating that warm CDW flooded PIB during the early Holocene (11.7-7.5 kyr BP) and that this inflow was reduced during the middle Holocene (7.5-4.0 kyr BP) to 130 late Holocene (4.0-0 kyr BP) (Fig. 3c; Extended Data Fig. 6). Intensified early Holocene CDW 131 inflow is corroborated by both the benthic and the planktic foraminiferal δ^{13} C ratios of the PIB 132 record which increase from relatively low values between ~10.4 and 9.0 kyr BP to higher 133 values that are maintained from 7.0 kyr BP until at least 1.2 kyr BP (Figs. 2, 3a,b). The benthic 134 δ^{13} C ratios recorded during the early Holocene match the δ^{13} C_{DIC} range of pure CDW. The 135 planktic δ^{13} C ratios during this time interval reflect the lower range of δ^{13} C_{DIC} values typical 136 for modern Antarctic Surface Water (AASW)³³ which in the ASE usually forms a 200-300 m 137 thick surface layer^{7,10}. This observation suggests that the reduced δ^{13} C ratios of *N. pachyderma* 138 sin., which calcifies within this AASW layer²⁹, are caused by effective admixture of 139 unmodified CDW into AASW. 140

Reduced ¹²C export to the seabed in PIB caused by a decrease in plankton productivity 141 can be ruled out as the reason for the benthic foraminifer $\delta^{13}C$ shift towards heavier values 142 centred at 7.5 kyr BP because this would have resulted in lighter planktic foraminifer δ^{13} C 143 values, i.e. opposite to what is observed. In addition, the Ba/Ti ratios and foraminifer 144 concentrations in core PS75/160 indicate that biological production probably increased rather 145 than decreased after ~7.5 kyr BP (Fig. 2, Extended Data Fig. 5). An onset or intensification of 146 air-sea gas exchange between the atmosphere and surface waters in PIB at 7.5 kyr BP might 147 have resulted from a reduction in ice-shelf coverage. The ice-shelf break-up would be 148 consistent with the productivity increase after 7.5 kyr BP because phytoplankton production 149 on the Antarctic continental margin is mainly controlled by the availability of light³⁰. Such a 150 scenario is in agreement with several early Holocene benthic δ^{13} C values in the PIB record that 151 are even more δ^{13} C-depleted than pure CDW as well as a few planktic δ^{13} C values that are 152 lower than $\delta^{13}C_{DIC}$ of AASW (Figs. 2, 3a,b). This observed $\delta^{13}C$ -depletion could be explained 153 by advection of nearly pure CDW combined with reduced air-sea carbon exchange caused by 154 ice-shelf cover of PIB during the early Holocene (cf. refs. 31, 32). 155

156 Environmental implications from foraminifer assemblages

The benthic foraminifer assemblage in core PS75/160 shows a distinct change at ~7.5 157 kyr BP, when an assemblage dominated by Nonionella bradii and Globocassidulina species 158 (G. subglobosa, G. biora) is replaced by an assemblage dominated by various Angulogerina 159 species, especially A. angulosa (Methods), and when both total foraminifer concentration and 160 arenaceous benthic foraminifer abundance increase (Fig. 2a, Extended Data Fig. 7). While both 161 Globocassidulina species and Angulogerina species have been found in modern^{34,35} and 162 Holocene sub-ice shelf settings³⁶⁻³⁹, an assemblage dominated by Angulogerina observed on 163 the eastern Antarctic Peninsula shelf has been attributed to an ice-shelf edge setting³⁴. Also 164 Nonionella spp. has been reported from modern^{34,35} and Holocene sub-ice shelf sediments^{36,39} 165 but its dominance in benthic foraminifer assemblages is frequently related to high biological 166 productivity⁴⁰. Given that neither the benthic δ^{13} C values nor the Ba/Ti ratios indicate higher 167 productivity during the early Holocene (Figs. 2, 3a), we interpret the Globocassidulina and 168 Nonionella dominated assemblage to indicate ice-shelf presence. Our interpretation is 169 consistent with the dominance of Nonionella species and Globocassidulina species in benthic 170 foraminifer assemblages of early to late Holocene sediments from the part of the eastern 171 Antarctic Peninsula shelf which had been permanently ice-shelf covered from 11.5 kyr BP until 172 AD 2002 (refs. 36, 41). 173

174 We suggest that PIB remained covered by an extensive ice shelf from at least ~9.2 until ~7.5 kyr BP. This conclusion agrees with the high proportion of a distinct non-encrusted shell 175 morphotype of *N. pachyderma* sin. in the early Holocene sediments of core PS75/160 (Fig. 2a), 176 which is indicative of specimens of a 'neanic' (i.e. adolescent) to adult life stage⁴² (Extended 177 Data Fig. 8). In contrast, the middle and late Holocene sediments in core PS75/160 are 178 dominated by an encrusted shell morphotype that is typical for a terminal life stage⁴² (Extended 179 Data Fig. 8). We explain the dominance of the adolescent and adult shells by foraminifer 180 reproduction in seasonal open water offshore from the ice shelf covering PIB, the subsequent 181 182 advection of these foraminifera under the ice shelf by ocean currents, and their demise in this inhospitable environment for zooplankton, i.e. before they could reach the terminal life stage. 183 A contributing factor for the observed down-core change in the abundance of the two 184 morphotypes of *N. pachyderma* sin. may be a deepening of the pycnocline. Under ice shelves 185 the pycnocline lies deeper than in open water^{7,10} and is overlain by a cool mixed layer, in which 186 non-encrusted morphotypes are concentrated⁴². Given the modern water depth of 337 m at site 187 PS75/160 and taking into account the 30-10 m lower global sea level⁴³ and the predicted \sim 50 188 m of seafloor depression in response to glacial loading of the PIB hinterland between 10 and 189 7.5 kyr BP⁴⁴, the thickness of this early Holocene ice shelf cannot have exceeded ~410-430 m, 190 191 which is comparable to the modern thicknesses of Thwaites Glacier Ice Shelf, Ronne Ice Shelf (West Antarctica) and Amery Ice Shelf (East Antarctica) near their calving fronts^{8,45}. 192

Ice-shelf cover of PIB during the early Holocene would have contributed to the low 193 benthic and planktic δ^{13} C values recorded during that time. However, bottom-water 194 temperatures under ice shelves are lower than in seasonal open-marine settings^{10,45}, whereas 195 the benthic Mg/Ca data from PIB imply that these temperatures were higher in the proposed 196 197 early Holocene sub-ice shelf environment than in the middle to late Holocene open-marine setting. Consequently, enhanced CDW supply onto the ASE shelf must have been the dominant 198 factor influencing the early Holocene δ^{13} C record from PIB, even if ice-shelf cover contributed 199 to low benthic and planktic δ^{13} C values. The most plausible scenario for explaining the 200 simultaneous change in both the foraminiferal $\delta^{13}C$ composition and the bottom-water 201 temperature around ~7.5 kyr BP is the reduction of CDW inflow. We conclude that compared 202 to the present-day CDW advection onto the ASE shelf, which can be evaluated from the modern 203 benthic and planktic δ^{13} C ratios, CDW inflow was stronger from 10.4 to 8.0 kyr BP, while it 204 was weaker from 7.5 to 1.2 kyr BP (Figs. 2, 3a,b). 205

206 Drivers of ice-sheet change

The early Holocene intensification of CDW heat supply can be attributed to a southerly 207 position of the SHWW belt between 12.0 and 8.0 kyr BP⁴⁶ (Fig. 3g,h). The northward shift of 208 the westerlies from 8.0 to 6.0 kyr BP⁴⁶ is likely to have reduced CDW inflow as documented 209 in the PIB record (Fig. 3g,h). Thus, our findings expand on the conclusion from modern 210 observational and modelling studies that the SHWW exert a major control on CDW advection 211 onto the ASE shelf^{7,10,11,15}. The role of buoyancy forcing⁹, however, remains unclear from our 212 data. Early Holocene CDW advection may have been intensified along the entire Pacific 213 margin of West Antarctica and thus may also explain the higher sea-surface temperatures 214 observed on the western Antarctic Peninsula shelf during that time^{40,47} (Fig. 3e). 215

Grounded ice in PIB had retreated to within 110 km of its modern grounding line by 216 11.2 kyr BP^{21,22}. Our conclusion of intensified CDW upwelling onto the ASE shelf from 10.4 217 to 7.5 kyr BP and probably since 12.0 kyr BP, when the SHWW already had a southerly 218 position⁴⁶, suggests that, in accordance with results from ice-sheet modeling¹⁶⁻¹⁹, ocean thermal 219 forcing was the main driver of post-LGM WAIS retreat. In contrast, air temperatures in West 220 Antarctica¹⁴ were still relatively low and post-glacial eustatic sea level had risen by not more 221 than ~60 metres above its LGM low stand of 120 metres^{43,50} when PIB was already free of 222 grounded ice, indicating that atmospheric and sea-level forcing played only a subordinate role 223 224 for post-LGM WAIS deglaciation (Fig. 3d,f). Sustained high CDW heat supply and accompanying oceanic melting might have destabilised any extensive ice shelf remaining in 225 PIB until 7.5 kyr BP so that afterwards minor atmospheric warming in West Antarctica 226 combined with continued global sea-level rise were sufficient to trigger its disintegration (Fig. 227 3d-f). 228

The modern benthic and planktic δ^{13} C ratios indicate that CDW inflow onto the ASE 229 shelf must have increased at some time within the last millennium (Fig. 3a,b). The planktic 230 δ^{13} C down-core record available from site PS69/251 on the outer ASE shelf, dating back to AD 231 1888±37 according to its ²¹⁰Pb chronology, reveals that this intensification took place between 232 AD 1947±9 and AD 1963±9 (Fig. 4a). The increased CDW advection can be attributed to a 233 strengthening and/or poleward shift of the SHWW, which is evident from proxy data in West 234 Antarctic ice cores^{48,49} (Fig. 4b,c). Our finding is also in line with the hypothesis that sea-235 surface temperature warming in the central tropical Pacific Ocean during the 1940s, associated 236 with a very large El Ninõ event from AD 1939 to 1942, initiated the current period of CDW 237 inflow onto the ASE shelf and the resulting mass loss in this sector of the WAIS¹². This ice 238 loss was not evident until the 1970s, when the first satellite observational data from the ASE 239 sector became available². 240

241 Importance

Our study extends the observational record of CDW inflow onto the ASE shelf further back in time through a period of significant ice-sheet change and it suggests early Holocene ice-shelf coverage in PIB. Most significantly, it provides the first empirical evidence for the dominant role of ocean forcing in driving past and recent WAIS deglaciation and the suggested control of CDW upwelling onto the ASE shelf by the SHWW. This information is crucial for validating assumptions in numerical models^{6,18,19} and will help to make model-based predictions of future global sea-level rise from Antarctic ice-sheet melting more robust.

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Online content Methods, along with any additional Extended Data display items and Source
Data, are available in the online version of the paper; references unique to these sections appear
only in the online paper.

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Supplementary Information is available in the online version of the paper.

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Author contributions CDH conceived the idea for the study and together with JAS, GK and 405 RDL wrote the manuscript. GK, CDH and KG collected the PS69 sediment core and together 406 with JAS, PEJ and JPK the PS75 cores. CDH, JAS, SJR and RDL developed the ¹⁴C age models 407 for the PS75 cores. TJA conducted the ²¹⁰Pb measurements on the PS69 core and provided its 408 age model. GK, CDH, JAS, JPK and PEJ undertook the sedimentological analyses. DH 409 measured stable isotopes on the foraminifer shells, while MG and HE analysed the trace metals. 410 CJP, SK and MW analysed the foraminifer assemblages. All co-authors commented on the 411 manuscript and provided input to its final version. 412

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414 **Author Information** The data are available from the PANGAEA data base (PANGAEA doi 415 in preparation) and Extended Data Table 1. Reprints and permissions information is available 416 at www.nature.com/reprints. The authors declare no competing financial interests. Readers are 417 welcome to comment on the online version of the paper. Correspondence and requests for 418 materials should be addressed to CDH (hilc@bas.ac.uk).



Fig. 1: Map of modern bottom-water temperatures in the southern Amundsen Sea showing geographical locations and core sites (modified from Dutrieux et al.¹⁰). Green symbols indicate locations of cores analysed in this study, while the black symbol marks the site of a seafloor surface sample investigated for benthic foraminifer δ^{13} C on *Cibicidoides* spp. by Mackensen²⁵. Thick black, gray and white contours indicate seabed elevations of -1000 m, -500 m and -400 m, respectively (for a detailed bathymetric map of inner Pine Island Bay see Fig. 6 in Larter et al.²²).

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431 Fig. 2: Down-core variability of lithological facies, stable carbon isotope (δ^{13} C) 432 composition of foraminifer shells, biological productivity and benthic foraminifer 433 assemblages (core PS75/160 only) alongside mean radiocarbon dates for Holocene marine 434 sediment cores PS75/160 (Fig. 2a) and PS75/167 (Fig. 2b) from Pine Island Bay (PIB).

Lithofacies is from Hillenbrand et al.²¹ (M: bioturbated mud; MC: deformed, sheared pebble-435 sized muddy to sandy soft sediment clasts randomly orientated in muddy matrix; MSa: mud 436 interstratified with sandy layers; S: slightly bioturbated sand; GS: massive sandy gravel). 437 AMS ¹⁴C dates were recalibrated for this study (Extended Data Table 1). Asterisk at lowermost 438 age from core PS75/160 indicates average of replicate dates from the same sample horizon. 439 Facies MC at site PS75/160 probably results from seafloor scouring by iceberg or ice-shelf 440 keels²¹ but AMS ¹⁴C dates indicate reworking only for depth below 405 cm. δ^{13} C composition 441 of planktic (Neogloboquadrina pachyderma sinistral) and benthic (Angulogerina angulosa) 442 foraminifera and log-normalised Ba/Ti peak area ratios are shown for both cores.. The yellow 443 shaded areas highlight the range of $\delta^{13}C_{DIC}$ values typical for Circumpolar Deep Water 444 $(CDW)^{26,27}$, while the blue-green shaded areas indicate the $\delta^{13}C_{DIC}$ range typical for Antarctic 445 Surface Water (AASW)³³. Abundances of the three main benthic foraminifera groups 446 (Angulogerina spp., Globocassisulina spp., Nonionella spp.) among all benthic foraminifera 447 and of morphotype 2 tests among all planktic foraminifera tests of Neogloboquadrina 448 pachyderma sin. (Extended Data Fig. 8) are displayed for core PS75/160. Red asterisks in 449 lithofacies column of core PS75/160 indicate horizons sampled for Mg/Ca analyses on 450 Angulogerina angulosa. 451



Fig. 3: Variability of Circumpolar Deep Water advection onto the Amundsen Sea 454 Embayment shelf since 12 cal. ka BP in comparison to potential forcing mechanisms of 455 ice-sheet change in West Antarctica. Panels (a), (b) and (c): Proxies for CDW advection into 456 PIB. δ^{13} C composition of (a) benthic (A. angulosa) and (b) planktic foraminifera (N. 457 pachyderma sin.) in the spliced PS75/160 and PS75/167 record, with δ^{13} C values of modern 458 benthic (PS75/215) and planktic (PS69/251) foraminifera indicated by the black dashed lines 459 and $\delta^{13}C_{DIC}$ ranges typical for CDW and AASW highlighted by the yellow shaded and blue-460 green shaded areas, respectively; (c) Mg/Ca ratios of benthic foraminifera (A. angulosa) as a 461 462 semi-quantitative proxy for bottom-water temperatures at site PS75/160 (Mg/Ca data corrected for diagenetic coating of shells using Mg/Mn =0.15 mol/mol, see Extended Data Fig. 6 for 463 details; ages for Mg/Ca data points calculated by linear interpolation between maximum and 464 minimum calibrated dates of neighbouring AMS ¹⁴C samples given in Extended Data Table 1). 465 Panel (d): Deuterium (δD) isotope ratio in the WAIS Divide Ice Core from central West 466 Antarctica¹⁴ as proxy for air temperature forcing. Panel (e): Sea water surface temperature 467 (SST) from ODP Site 1098 in Palmer Deep, western Antarctic Peninsula shelf, as proxy for 468 combined atmospheric and/or oceanic forcing⁴⁷. Panel (f): Global sea-level curves^{43,50} as 469 proxies for sea-level forcing. Panels (g) and (h): Atmosphere-ocean temperature gradient in the 470 Sub-Antarctic SW Pacific as a proxy for the position of the Southern Hemisphere westerly 471 winds (SHWW)⁴⁶ (increase of temperature gradient indicates its N-ward shift). (g) surface air 472 temperature (ST) on Campbell Island, (h) SST at core site MD-97120, SW of New Zealand. 473 Light gray shaded area indicates timing of grounded WAIS retreat from the inner ASE 474 shelf^{21,22}, dark gray shaded area highlights that the timing of a major decrease in CDW 475 advection onto the ASE shelf coincided with a N-ward shift of the SHWW. 476

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Fig. 4: Variability of Circumpolar Deep Water advection onto the Amundsen Sea 480 Embayment shelf and Southern Hemisphere westerly winds during the last 150 years. 481 Panel (a): δ^{13} C composition of planktic foraminifer shells in core PS69/251 from the outer ASE 482 shelf versus age (derived from ²¹⁰Pb-dating). Error bars of the data points indicate the 483 uncertainty of the ²¹⁰Pb ages with the light and dark gray shaded areas illustrating the possible 484 timing and the most likely timing, respectively, of an increased admixture of CDW into AASW 485 as indicated by the planktic δ^{13} C values. Panel (b): Annual coarse dust particle percentage 486 (defined as the number of particles mL⁻¹ [4.5–15]/[1–15] μ m diametre ×100) in the WAIS 487 Divide Ice Core (48). Panel (c): Proxy for "northerly air mass incursions" (NAMI) into West 488 Antarctica, based on stacked non-sea-salt Ca^{2+} concentration data from 10 ice cores (49). The 489 orange/blue delineation for the curves in panels (b) and (c) is the mean of that record over the 490

- 491 time interval shown. The ice core proxies indicate that intensification and/or poleward shift of
- the SHWW coincided with increased CDW advection onto the ASE shelf.

Supplementary Online Information

Oceanic forcing of Holocene ice-sheet retreat in the Amundsen Sea Embayment, West Antarctica

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Methods

Core material. Gravity cores PS75/160-1 and PS75/167-1 and giant box core PS75/215-1 were recovered on RV Polarstern expedition ANT-XXVI/3 in 2010 (ref. 51), while giant box core PS69/251-1 was retrieved during RV Polarstern expedition ANT-XXIII/4 in 2006 (ref. 52) (Extended Data Table 1). For metadata, detailed core descriptions and X-radiographs of the al.²¹, al.²². al.⁵³ et Larter et Smith et cores, see Hillenbrand and doi.pangaea.de/10.1594/PANGAEA.751493.

Core processing, sedimentological analyses and chronologies. Methodologies of core processing and sedimentological analyses were reported by Hillenbrand et al.²¹, Larter et al.²² and Smith et al.⁵³. Analytical techniques for AMS ¹⁴C dating and calibrated AMS ¹⁴C ages were also previously published^{21,22,53} (see also doi.pangaea.de/10.1594/ PANGAEA.751493) but for this study we re-calibrated all conventional AMS ¹⁴C dates from cores PS75/160-1 and PS75/167-1 (Extended Data Table 1). Because mixed calcareous (micro-)fossils, which are known to show variable ¹⁴C vital effects⁴¹, had been radiocarbon-dated from cores PS75/160-1 and PS75/167-1, we used a marine reservoir age with a relatively large error (R+ Δ R= 1,100±200 years) for their calibration²¹. In contrast, Larter et al.²² and Smith et al.⁵³ applied a uniform marine reservoir age with a much smaller error (R+ Δ R= 1,300±70 years) to their calibration curve⁵⁵.

All reported and calibrated AMS ¹⁴C dates for core PS75/167-1 increase down-core or overlap within error but no ¹⁴C dates are available above 230 cm depth (Fig. 2b; Extended Data Table 1). The ¹⁴C dates for core PS75/160-1 show an age reversal below ~400 cm core depth (Fig. 2a; Extended Data Table 1). A thick unit of a distinct lithofacies (facies MC: deformed, sheared pebble-sized muddy to sandy soft sediment clasts randomly orientated in muddy matrix) occurs at this depth (Fig. 2a) suggesting that this lower part of core PS75/160-1 was affected by post-depositional turbation of the seafloor sediments by iceberg or ice-shelf keels²¹. Rather than ignoring the three trace metal (i.e. Mg/Ca) data and the benthic foraminifer assemblage data obtained from samples taken below this depth, we assigned for those samples ages by linear interpolation between maximum and minimum calibrated AMS ¹⁴C dates of neighbouring age-depth fix points. This approach leaves some uncertainty in the absolute ages for the corresponding samples but an early Holocene age for the lower part of core PS75/160-1 is documented by the available AMS ¹⁴C dates (Fig. 2a; Extended Data Table 1). We spliced the

stable carbon isotope (δ^{13} C) records of cores PS75/160-1 and PS75/167-1 at 8.2 kyr BP (Extended Data Fig. 1) and used exclusively δ^{13} C data from below 230 cm depth in core PS75/167-1 for the time interval >8.2 kyr BP and from above 380 cm depth in core PS75/160-1 for the time span <8.2 kyr BP (Fig. 3a,b).

Radiocarbon dating of planktic foraminifera (Neogloboquadrina pachyderma sinistral) from undisturbed seafloor surface sediment retrieved in giant box core PS69/251-1 provided an uncorrected AMS ¹⁴C age of 1,144±37 years^{21,53} and thus the youngest ¹⁴C date obtained from modern surface sediments from the continental shelf in the Amundsen and Bellingshausen seas²² (Extended Data Table 1). As calcareous (micro-)fossil material was insufficient for AMS ¹⁴C dating in samples taken from below the surface, lead-210 (²¹⁰Pb) and caesium-137 (¹³⁷Cs) dating were applied to core PS69/251-1 for establishing ages of down-core samples. The ²¹⁰Pb and ¹³⁷Cs activities were measured on 1 cm thick sediment slices by gammaspectrometry using Canberra ultra-low-background Ge-detectors at the Gamma Dating Centre, Department of Geosciences and Natural Resource Management, University of Copenhagen. ²¹⁰Pb was measured from its gamma-peak at 46.5 keV. ²²⁶Ra from the granddaughter ²¹⁴Pb (peaks at 295 and 352keV) and ¹³⁷Cs from its peak at 661 keV. The core showed surface contents of unsupported 210 Pb of ~115 Bq kg⁻¹ with a tendency for exponential decline with depth in the upper 6 cm (Extended Data Fig. 2). Below ~6 cm core depth the activity was at the detection limit or lower. The calculated flux of unsupported ²¹⁰Pb is 100 Bq m^{-2} yr⁻¹ which is in reasonable agreement with the expected flux from atmospheric deposition⁵⁶. The content of ¹³⁷Cs was generally very low (Extended Data Fig. 2). Constant Rate of Supply (CRS) modelling was applied on the profile using a modified method⁵⁶ where the activity below 5.5 cm was calculated on the basis of a regression (Extended Data Fig. 3). The dating results are given in Extended Data Table 2 and Extended Data Figure 4.

X-ray fluorescence scanning. Semi-quantitative uncalibrated major and trace element composition of cores PS75/160-1 and PS75/167-1 were analysed at the Alfred Wegener Institute (Bremerhaven) using an Avaatech XRF core scanner⁵⁷. We applied XRF core scanning predominantly for the determination of the barium/titanium (Ba/Ti) peak-area ratio (hereafter referred to as the Ba/Ti ratio), which is a proxy for biogenic barium (Ba_{bio}). Ba_{bio} is considered to be the most reliable palaeoproductivity proxy in Antarctic continental margin sediments and Southern Ocean sediments south of the Antarctic Polar Front^{30,58-60}. Following Weltje and Tjallingi⁶¹ we plotted the log-normalised [Ln(Ba/Ti)] peak-area ratios in Figures 2 and Extended Data Figure 5. In order to assess down-core changes in the input of barium that is

incorporated in the terrigenous heavy mineral barite we also calculated the log-normalised barium/zirconium (Ba/Zr) peak-area ratios, with Zr being a proxy for the terrigenous heavy mineral zircon (Extended Data Fig. 5).

Stable carbon and oxygen isotope analyses. Stable carbon (δ^{13} C) and oxygen (δ^{18} O) isotope analyses were performed on the planktic foraminifer species Neogloboquadrina pachyderma sinistral (12-24 shells; morphotype 1) and the epibenthic to shallow infaunal foraminifer Angulogerina angulosa (= Trifarina angulosa) (5-20 shells) (Extended Data Figs. 5, 7). The δ^{13} C and δ^{18} O isotopes were measured on 104 planktic and 112 benthic foraminifer samples from cores PS75/160-1 and PS75/167-1, one benthic foraminifer sample from site PS75/215-1 and seven planktic foraminifer samples from core PS69/251-1. The shells were picked from the wet sieved 63-2000 µm sediment size fraction and analysed at the Godwin Laboratory for Palaeoclimate Research at the Department of Earth Sciences, University of Cambridge (UK). Foraminifer shells were first crushed and soaked in a solution of 3% H₂O₂ for 30 minutes, then acetone was added and the sample ultrasonicated for 10 seconds before the liquid was decanted. Samples were then dried at 50°C overnight. Vials were sealed with septa and screw caps and analysed using a Micromass Multicarb sample preparation system attached to a VG SIRA Mass Spectrometer. Each run of 30 samples was accompanied by ten reference carbonates and two control samples. Results are reported with reference to the international standard Vienna Pee Dee Belemnite (VPDB) and analytical precision is better than $\pm 0.06\%$ for δ^{13} C and $\pm 0.08\%$ for δ^{18} O. In this study, we only used the δ^{13} C data of A. angulosa and N. pachyderma sin. The δ^{13} C data of A. angulosa and N. pachyderma sin. were corrected for vital effects by adding +0.85‰ (cf. ref. 24) and +1.0‰ (refs. 28, 29), respectively.

Trace metal analyses. Ten samples from core PS75/160-1 contained sufficient calcareous benthic foraminifer shells of *A. angulosa* (200-300 μ g) and ten samples from this core contained sufficient planktic foraminifer shells of *N. pachyderma* sin. (300-500 μ g) for trace metal analyses. Samples were only taken from horizons with a lithofacies that does not indicate signs of reworking (i.e. no samples were taken from facies MC; Fig. 2a).

Cleaning of the shells followed the procedure described in Barker et al.⁶². Element ratios were determined by ICP–OES⁶³ and ICP-MS⁶⁴. Long-term instrumental precision of element ratio data, determined by replicate analyses of a standard solution containing magnesium/calcium (Mg/Ca)= 1.3 mmol/mol, was $\pm 0.46\%$. Accuracy of Mg/Ca determinations was confirmed by inter-laboratory studies of foraminifer and carbonate reference materials^{65,66}.

Iron/calcium (Fe/Ca) and manganese/calcium (Mn/Ca) were measured to monitor cleaning efficiency and diagenetic effects. The N. pachyderma sin. samples contained high Fe/Ca and Mn/Ca (up to 0.40 and 2.48 mmol/mol respectively) with strong positive correlation between Mg/Ca, Fe/Ca ($r^2 = 0.93$) and Mn/Ca ($r^2 = 0.94$), reflecting diagenetic coating of the shells⁶⁷ not removed by the cleaning procedure. Therefore, the Mg/Ca data on N. pachyderma sin. are not interpreted. The A. angulosa samples contained much lower Fe/Ca and Mn/Ca (maximum 0.14 and 0.64 mmol/mol, respectively) with no correlation observed between Mg/Ca and Fe/Ca (r^2 = 0.02) and a smaller (but still significant) correlation between Mg/Ca and Mn/Ca ($r^2 = 0.56$). In order to evaluate the impact of any diagenetic Mg contribution on the Mg/Ca data, we applied a correction to the measured Mg/Ca of A. angulosa (Fig. 3c, Extended Data Fig. 6a), assuming Mg/Mn of a diagenetic coating is 0.15±0.05 mol/mol, reflecting average composition of Mn nodules and ferromanganese crusts^{68,69}, especially of those from the South Pacific and Pacific sector of the Southern Ocean⁷⁰⁻⁷². The adjusted Mg/Ca ratios show the same (although less pronounced) down-core trend as the non-adjusted Mg/Ca data, with higher Mg/Ca ratios found before 7.5 kyr BP (Extended Data Fig. 6a,b). Also use of other Mg/Mn ratios for diagenetic coatings⁶⁹ still indicates that the early to middle Holocene decrease in our Mg/Ca data is a robust feature (Extended Data Fig. 6b).

Boron/Calcium (B/Ca) was analysed on all. *A. angulosa* and *N. pachyderma* sin. samples aiming to exclude an influence of alkalinity on the Mg/Ca records^{73,74}. Neither the benthic nor the planktic foraminiferal data show a correlation between Mg/Ca and B/Ca. However, B/Ca values measured on the *A. angulosa* shells were very low, in the range 7.1 to 13.0 μ mol/mol, and probably influenced by the laboratory blanks. Consequently, the B/Ca data are not interpreted.

The Mg/Ca composition of calcareous foraminifer shells is frequently used as a quantitative proxy for seawater temperature^{20,75} but can also be controlled by the carbonate ion concentration ($[CO_3^{2-}]$) in the ambient seawater^{73,76}. Consequently, higher Mg/Ca ratios as observed in core PS75/160-1 during the early Holocene (Fig. 3c, Extended Data Fig. 6a) indicate an increase in either the temperatures and/or the $[CO_3^{2-}]$ of the deep-water mass flooding Pine Island Bay (PIB) during that time. The influence of the latter can be evaluated by measuring the B/Ca ratio of the foraminifer shells^{73,76,77}, which, however, provided only inconclusive results for our samples from core PS75/160-1 (see above). The pre-industrial $[CO_3^{2-}]$ in Antarctic Surface Water (AASW) was 120 µmol/kg^{ref.29}, whereas Circumpolar Deep Water (CDW) is characterised by a uniform, relatively low $[CO_3^{2-}]$ of 84±5 µmol/kg throughout the Southern Ocean, including its Pacific sector and the region south of the Antarctic Polar

Front⁷⁸. If the elevated benthic foraminiferal Mg/Ca ratios in PIB recorded during the early Holocene were caused by the presence of an AASW-like water mass that comprised the entire water column, the corresponding benthic foraminiferal δ^{13} C ratios can be expected to resemble δ^{13} C_{DIC} values typical for AASW. In contrast, our early Holocene benthic foraminiferal δ^{13} C data from cores PS75/160-1 and PS75/167-1 are lower and typical for CDW (Fig. 3b, Extended Data Fig. 5) and thus not consistent with a scenario that deep water present on the shelf of the Amundsen Sea Embayment (ASE) during that time was derived from AASW with the benthic Mg/Ca data reflecting its higher [CO₃²⁻].

Deep-water warming on the ASE shelf could have resulted either from intensified CDW upwelling or CDW warming. CDW warming would have resulted from a higher influx of North Atlantic Deep Water (NADW), which is the source water mass for CDW, into the Southern Ocean. NADW is injected into the clockwise flowing Antarctic Circumpolar Current in the Atlantic sector of the Southern Ocean, mixes with recirculated deep waters from the Indian and Pacific oceans, and, most importantly, acts as the "heat source" for CDW^{79} . The $[CO_3^{2-}]$ of NADW is ~125 µmol/kg compared to only 84±5 µmol/kg in CDW⁷⁸. Thus, intensified NADW influx into the Southern Ocean during the early Holocene, which is supported by marine sedimentary records from the Atlantic sector⁸⁰, would have raised both the temperature and the [CO₃²⁻] in deep water flooding the ASE shelf, consistent with our Mg/Ca data from PIB. The $\delta^{13}C_{DIC}$ of NADW is higher than in CDW⁸¹. Consequently, increased NADW supply would have raised the $\delta^{13}C_{DIC}$ of CDW. This, however, is in conflict with the early Holocene benthic for a for a form our PIB cores, which are characterised by low values typical for CDW (Fig. 3b, Extended Data Fig. 5). Consequently, the elevated benthic foraminiferal Mg/Ca ratios recorded in PIB during the early Holocene must reflect deep water warming caused by intensified CDW upwelling.

Analyses of benthic foraminifer assemblages. A total of 38 samples (1 cm thick sediment slices) were taken from core PS75/160-1 for benthic foraminifer analysis⁸². As with the trace metal samples, no samples were taken from the core section with lithofacies MC between ~405 and 485 cm depth. The samples were washed through a 63 μ m mesh-size sieve before the coarse residue was dry-sieved over 2 mm. All samples were screened in order to estimate the abundance of foraminifera. 27 samples with a relatively high content of foraminifera were selected for microscopic analysis and then dry-sieved over 125 μ m and 250 μ m. 23 of the 27 samples contained >300 benthic foraminifer shells in the >125 μ m fraction. Samples with significantly more than 300 specimens and/or a large sample volume were split using a

traditional hand-splitter until a suitable aliquot remained. Picked foraminifera were then identified under an optical microscope and by using Scanning Electron Microscope (SEM) images. Classification and identification were carried out in accordance with literature on recent Antarctic foraminifera (for full reference list, see ref. 82).

Throughout core PS75/160-1, abundances of foraminifera are high, with total numbers ranging from 81 to 1511 specimens per sample, and over 14,000 total specimens were identified from all samples⁸². The 27 samples yielded 84 benthic taxa but only one planktic species (*N. pachyderma* sin.). Between 0 and 341.5 cm core depth the abundance of foraminifera per gram of dry sediment (foraminifera/g) is substantially greater than from 352.5 to 653.5 cm core depth, and there is a large range throughout the core (maximum: 306.19 foraminifera/g at 281.5 cm; minimum: 1.85 foraminifera/g at 631.5 cm) (Extended Data Fig. 7). Between 0 and 341.5 cm core depth, the average concentration of 102.58 foraminifera/g far exceeds the average concentration of only 20.81 foraminifera/g between 352.5 and 653.5 cm core depth.

Benthic calcareous species dominate the assemblages throughout the core while the abundance of arenaceous species accounts for a maximum of 30% (depth: 141 cm) (Extended Data Fig. 7). Over two-thirds of samples comprise more than 85% calcareous foraminifera⁸². However, arenaceous foraminifera are more abundant between 0 and 341.5 cm (mean: 15.2%) than between 352.5 and 653.5 cm (mean: 7.0%). *N. pachyderma* sin. is not particularly abundant, contributing a maximum of 5.5% to the total assemblage at 281.5 cm, though it is present in all samples. Preservation of benthic and planktic foraminifer shells was good to moderate throughout the core.

The upper section of core PS75/160-1 from 0 to 341.5 cm depth typically contains high abundances and percentages of the calcareous benthic species *Angulogerina angulosa*, *A.* sp. 1, *Uvigerina* sp. 1, *Globocassidulina subglobosa*, and *G. biora* (Extended Data Fig. 7). Relatively abundant accessory species include *Astrononion echolsi*, *Alabaminella weddellensis* and the arenaceous species *Miliammina arenacea*, *Portatrochammina* sp. 1 and *Labrospira jeffreysii*. We follow previous researchers on benthic foraminifera from the Antarctic continental margin and summarize under 'A. *angulosa*' the species *A. angulosa*, *A. earlandi*, *Trifarina angulosa* and *T. earlandi*^{37,83-85}, which are morphospecies rather than separate biological species (cf. refs. 86-88). Moreover, molecular genetic data have shown that *Angulogerina/Trifarina* is closely related to *Uvigerina peregrina*^{88,89}, with which it also shares the epifaunal to shallow infaunal habitat^{24,83,90}.

A major faunal change was observed between 341.5 and 352.5 cm core depth. The section below this depth is characterised by low abundances of *A. angulosa* and *A. pauperata*. This

lower core section is dominated by *Nonionella bradii*, *N. iridea*, *G. subglobosa*, and *G. biora* (Extended Data Fig. 7). The subsidiary species include *Ehrenbergina glabra*, *Fursenkoina fusiformis* (calcareous), and *Portatrochammina* sp. 1 (arenaceous). The planktic foraminifera in this lower core section are dominated by *N. pachyderma* sin. type 2, while *N. pachyderma* sin. type 1 dominates the upper section (Extended Data Figs. 7, 8). Principal Component Analysis (PCA), Correspondence Analysis (CA) and Cluster Analysis demonstrate that the pronounced change between the two dominant benthic foraminifera assemblages between 341.5 and 352.5 cm core depth is statistically robust⁸².

Extended Data Figures



Extended Data Figure 1: Age-depth plots for cores PS75/160-1 (above) and PS75/167-1 (below). The red diamonds mark the calibrated median AMS ¹⁴C ages with the error bars

indicating the maximum and minimum ages (Extended Data Table 1). The vertical dashed line shows the age of 8.2 kyr BP where the benthic and planktic δ^{13} C records of the two cores were spliced in Figure 2a and 2b (i.e. only δ^{13} C data from below 230 cm depth were used from core PS75/167-1 and only δ^{13} C data from above 380 cm depth were used from core PS75/160-1).



Extended Data Figure 2: Concentrations of radiogenic lead and caesium isotope concentrations versus depth in giant box core PS69/251-1.

Left: Concentration of unsupported ²¹⁰Pb (²¹⁰Pb_{xs}). Note logarithmic scale. Surface concentration is ~115 Bq kg⁻¹, and an exponential decline of ²¹⁰Pb_{xs} with depth is observed in the upper 6 cm. Activity is at the detection limit or lower below 6 cm depth. The calculated ²¹⁰Pb_{xs} flux is 100 Bq m⁻² yr⁻¹, which is in reasonable agreement with the expected flux from atmospheric deposition⁵⁶. Error bars denote one standard deviation of ²¹⁰Pb_{xs} concentrations.

Right: Concentration of ¹³⁷Cs. Activity is at or below detection limit throughout the core. Error bars denote one standard deviation of ¹³⁷Cs concentrations.



Extended Data Figure 3: Constant Rate of Supply (CRS) modeling of down-core profile of ${}^{210}Pb_{xs}$ concentrations in core PS69/251-1 using a modified method⁵⁶. Black straight line marks regression to calculate ${}^{210}Pb$ concentration below 6 cm core depth. Open circles highlight samples with ${}^{210}Pb_{xs}$ concentrations at or below detection limit. Error bars denote one standard deviation of ${}^{210}Pb_{xs}$ concentrations.



Extended Data Figure 4: Age-depth plot for the upper 5.5 cm of core PS69/251-1 based on CRS modeling of ²¹⁰Pb_{xs} concentrations. Error bars denote one standard deviation of calculated ages.



Extended Data Figure 5: Down-core plots of δ^{18} O and δ^{13} C isotope ratios measured on planktic (*Neogloboquadrina pachyderma* sinistral) and benthic foraminifera (*Angulogerina angulosa*), LN(Ba/Ti), LN(Ba/Zr) and AMS ¹⁴C ages of calcareous microfossils^{21,22,53} of cores PS75/160-1 (above) and PS75/167-1 (below). The ranges of δ^{13} C values typical for AASW and CDW are indicated by the blue-green and yellow shaded areas,

900

10,356±578

•

respectively. While prominent peaks in LN(Ba/Ti) below ~220 cm core depth in core PS75/167-1 are significantly higher than the background LN(Ba/Ti) values, the LN(Ba/Zr) peaks are not significantly higher in this core section. This relation suggests that the LN(Ba/Ti) and Ba/Ti peaks are caused by barium input through increased supply of terrigenous heavy minerals (including barite and zircon), which is confirmed by the down-core increase of terrigenous sand layers below ~220 cm core depth (see Fig. 2 in ref. 21).



Extended Data Figure 6: Mg/Ca ratios of benthic foraminifer shells (*A. angulosa*) in core PS75/160-1 adjusted for a potential diagenetic Mg contribution by assuming the Mg/Mn ratio of a diagenetic coating is 0.15±0.05 mol/mol (Extended Data Figure 6a), and

displayed as sample groups with ages of 10.0-7.5 kyr BP (4 samples), 7.5-4.0 kyr BP (2 samples) and 4.0-0 kyr BP (4 samples) with no adjustment and adjusted for potential diagenetic Mg contributions and Mg/Mn ratios in diagenetic coatings of 0.10, 0.15, 0.20 and 0.25 (Extended Data Figure 6b; references 68-72). Error bars in Extended Data Figure 6a result from the uncertainty of ±0.05 mol/mol for the Mg/Mn composition of the coating and the range of the calibrated AMS ¹⁴C dates of neighbouring sample horizons (Extended Data Table 1), from which the ages for the Mg/Ca samples were calculated.



Extended Data Figure 7: Holocene changes in foraminifer assemblages of core PS75/160-1 from Pine Island Bay. Benthic foraminifer assemblages (only the most abundant and selected benthic taxa are displayed), abundance of *N. pachyderma* sin. type 2 in relation to all foraminifera, abundance of arenaceous benthic foraminifera (in relation to all benthic foraminifera) and total foraminifer concentration (individuals per gram dry sediment). δ^{13} C composition of planktic and benthic foraminifera is also shown. Note major shift in foraminifer abundances and assemblages centred at 7.5 kyr BP. Benthic foraminifer taxa: *Alabaminella weddellensis, Angulogerina angulosa, Angulogerina pauperata, Angulogerina spp.* (= sum of all *Angulogerina spp.* (= sum of all *Globocassidulina biora, Globocassidulina subglobosa, Globocassidulina* spp. (= sum of all *Ronionella spp.* (= sum of all *Nonionella species*).





Extended Data Figure 8: Scanning Electron Microscope (SEM) images of whole shells and detailed shell surfaces of *Neogloboquadrina pachyderma* sin. morphotypes 1 and 2 in core **PS75/160-1.** Morphotype 1 is encrusted with gametogenic calcite and dominates the upper

section of the core (**Extended Data Fig. 8a-d**), while thin-walled and non-encrusted morphotype 2 (**Extended Data Fig. 8e-g**) dominates the lower section of core PS75/160-1. A scale bar (white dots) is shown in the lower right corner of each photo, and the unit of its length (number) is given in µm. Note the indistinct chambers and gametogenic calcite secreted around the whole shell in morphotype 1, while the individual chambers and porous shells of morphotype 2 are clearly visible, thereby showing beginning encrustation (white calcite around pores) (cf. refs. 42, 91-93). Morphotype 1 usually dominates the lower part of the water column in Antarctic and Arctic waters and is preserved in marine sediments, whereas morphotype 2 is abundant in the upper part of the water column and not preserved in the sediments^{42,91-93}. Encrusted morphotype 1 is typical for its neanic (i.e. adolescent) to adult stage^{42,91-94}. In analogy with the SEM-defined progressive dissolution steps distinguished by Henrich⁹⁵, **Extended Data Fig. 8a** and **8b** show shells with intermediate dissolution, **Extended Data Fig. 7c** a shell with initial to intermediate dissolution, **Extended Data Fig. 8d** a shell with initial dissolution and **Extended Data Fig. 8e**, **8f** and **8g** well preserved shells.

Extended Data Tables

Core ID	Lati-	Longi-	Water	Reco-	Sample	Laboratory	Dated	¹⁴ C age	Max.	Min.	Med.	Mean
	tude (°S)	tude (°W)	depth (m)	very (m)	depth (cmbsf)	code	mat.	±error (vr BP)	age (vr BP)	age (vr BP)	age (vr BP)	age (vr BP)
PS2544-1 GBC	71.796	105.191	546	0.21	surface	N/A	N/A	N/A	N/A	N/A	N/A	N/A
PS69/251-1 GBC	72.069	104.483	573	0.38	surface	SUERC-18942	pF	1144±37	N/A	N/A	[0]	[0]
PS75/215-1 GBC	74.592	104.042	556	0.62	surface	Beta-315966	Bz	1190±30	N/A	N/A	[0]	[0]
PS75/160-1 GC	74.564	102.624	337	6.68	0.5	Beta-284601	bF, S, O	2340±40	1615	756	1190	1186
					133.5	Beta-284602	S	2750±40	2118	1212	1634	1665
					230	Beta-284604	bF, S, O	4990±40	4919	3840	4405	4380
					300	Beta-284681	bF, S	7180±40	7400	6514	6967	6957
					380	Beta-284606	bF, S	8410±40	8592	7735	8137	8164
					490+500	Beta-284607	bF, P, S, O	9320±40	9729	8650	9243	9190
					588+600	Beta-284608	bF, S, O, Bp	8560±40	8861	7913	8331	8387
					652	Beta-284610	S	8200±40	8358	7578	7966	7968
					top of CC	Beta-284600	Bp, S, bF	8240±40	8390	7603	8006	7997
PS75/167-1 GC	74.623	105.802	526	9.34	230	Beta-300838	F	8490±40	8734	7820	8254	8277
					250	Beta-300839	F, S	8740±40	9040	8065	8547	8553
					280	Beta-284682	F	9080±50	9432	8451	8949	8942
					401	Beta-284683	F, S, O	9210±50	9541	8550	9113	9046
					478	Beta-300837	F, O, E	9180±40	9516	8547	9077	9032
					598	Beta-300840	F	9490±40	10070	8979	9459	9525
					625.5	Beta-284684	F	9660±50	10206	9193	9690	9700
					649	Beta-300841	F	9280±40	9656	8616	9196	9136
			1		670	Beta-284685	F	9450±50	10029	8899	9405	9464
			1		746	Beta-284686	F, S	9690±50	10228	9231	9726	9730
			1		830	Beta-300842	F	9620±50	10169	9133	9640	9651
					900.5	Beta-284687	F, S	10210±60	10934	9778	10372	10356

Extended Data Table 1: Locations, conventional and calibrated AMS ¹⁴C dates on calcareous microfossils from the investigated sediment cores. Coring devices: GBC: giant box core, GC: gravity core. Sample depths are given in centimetres below seafloor (cmbsf; top of CC: sample was taken from top of core catcher, corresponding to 652 cmbsf). All ¹⁴C-dates were corrected using an offset (ΔR) of 700±200 years from the global marine reservoir effect (R) of 400 years. The corrected ¹⁴C-dates were calibrated with the CALIB Radiocarbon Calibration Program version 7.1html⁵⁴ (http://calib.qub.ac.uk/calib/) using the MARINE13 calibration dataset⁵⁵. Uncorrected dates (=¹⁴C age) are given in ¹⁴C years before present (BP) and maximum, minimum, median and mean dates are given in calibrated years BP. Errors of calibrated dates are given as a 2 σ range. Dated calcareous microfossils: bF: benthic foraminifera, Bp: brachiopods, Bz: bryozoans, E: echinoid spines, F: mixed benthic and planktonic foraminifera, O: ostracods, P: pteropods, pF: planktonic foraminifera, S: unspecified calcareous shell fragments.

Depth (cm)	Age (yr)	Age error (yr)	Date (yr AD)	Mass accumu- lation rate (kg m ⁻² yr ⁻¹)	Mass accumu- lation rate error (kg m ⁻² yr ⁻¹)
0			2006		
0-1	6	3	2000	0.78	0.10
1-2	21	4	1985	0.57	0.10
2-3	40	6	1966	0.41	0.10
3-4	59	9	1947	0.42	0.16
4-5	81	15	1925	0.38	0.19
5-6	118	37	1888	0.24	0.28

3 Extended Data Table 2: Chronology of the upper 6 cm of core PS69/251-1 based

4 on CRS-modeling of ²¹⁰Pb_{xs} concentrations.

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