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F Beny, Viviane Bout-roumazeilles, G R Davies, C Waelbroeck, A Bory, et al.. Radiogenic isotopic and clay mineralogical signatures of terrigenous particles as water-mass tracers: New insights into South Atlantic deep circulation during the last termination. *Quaternary Science Reviews*, 2020, 228, 10.1016/j.quascirev.2019.106089 . hal-02992431

**HAL Id: hal-02992431**

**<https://hal.science/hal-02992431>**

Submitted on 17 Nov 2020

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# **Radiogenic isotopic and clay mineralogical signatures of terrigenous particles as water-mass tracers: new insights into South Atlantic deep circulation during the Last Termination.**

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## **Abstract**

The past evolution of the Southern Ocean, one of the major components of the climatic system, is still a matter of debate. This study provides new insights into the deep Southern Ocean circulation based on the radiogenic isotopes and clay mineralogical signature of the terrigenous fractions transported by the main deep water masses to sediments recovered in core MD07-3076Q from the central South Atlantic. This approach successfully permits: (1) provenance identification of the various grain-size fractions (clay, cohesive silt and sortable silt); (2) assignment of each grain-size fraction to a specific water-mass; (3) reconstruction of past changes in the main deep water-mass pathways. These data document the evolution of deep-water masses in the South Atlantic Ocean during the last deglaciation. The Antarctic Bottom Water (AABW) speed and northward extension were maximum at the

25 end of the Last Glacial Maximum (LGM), associated with strong bottom water production in  
26 the Weddell Sea, together with a vigorous Lower Circumpolar Deep Water (LCDW). In  
27 contrast the North Atlantic Deep Water (NADW) circulation was weaker than today. The  
28 onset of the deglaciation (from 17.5 ka to 15 ka, ~Heinrich Stadial 1, HS 1) was marked by  
29 weakening and southerly retreat of the AABW and by an increase of mixing between AABW  
30 and LCDW. The speed of the AABW remained at its lowest during the Bølling Allerød (B/A)  
31 and the Younger Dryas (YD), and the LCDW slowed and retreated to the south, while the  
32 NADW progressively migrated southward, deepened, and strengthened between the  
33 beginning of the Bølling Allerød and the Holocene (from ~15 to 10 ka).

## 34 **Keywords**

35 Last Deglaciation; Paleoceanography; Southern Ocean; South Atlantic; Radiogenic  
36 isotopes; Clay mineralogy; Grain size distribution.

## 37 **I. Introduction**

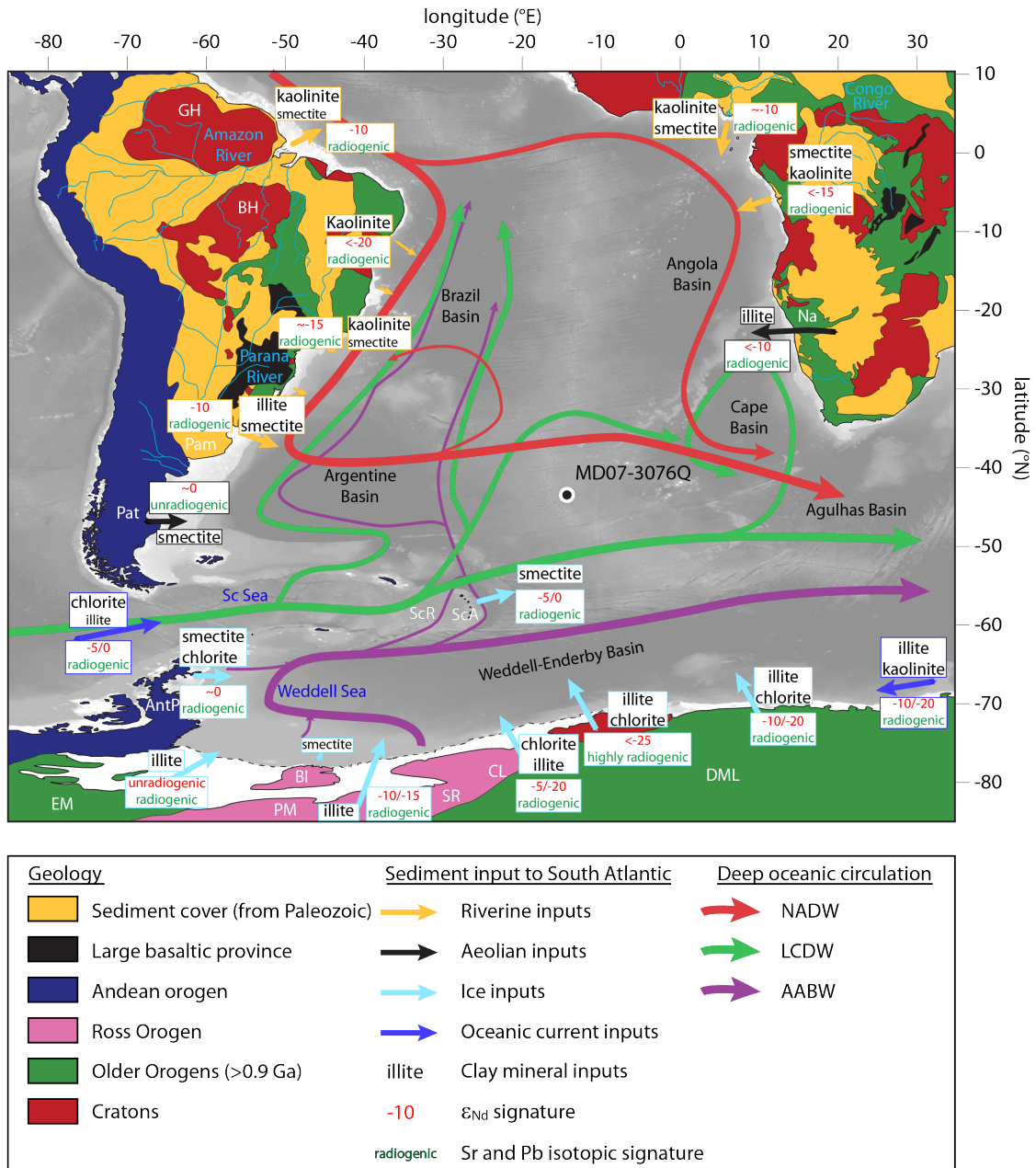
38 The Southern Ocean (SO) is a key area for understanding the present and past carbon  
39 cycle. Direct observations indicate that the SO is a sink of atmospheric CO<sub>2</sub> today ([Le Quéré  
40 et al., 2007](#); [Sallée et al., 2012](#)) while it was likely a source of CO<sub>2</sub> to the atmosphere during  
41 the deglaciation ([Bouttes et al., 2012](#); [Jaccard et al., 2013](#)). Upwelling of CO<sub>2</sub>-rich deep water  
42 likely explains most of the CO<sub>2</sub> rise observed during Heinrich Stadials (HS; [Ahn and Brook,  
43 2008](#); [Anderson et al., 2009](#); [Fischer et al., 2010](#); [Skinner et al., 2010](#); [Menviel et al., 2014,  
44 2017](#)). Several other processes occurring in the Southern Ocean have also been invoked to  
45 explain the deglacial atmospheric CO<sub>2</sub> rise. For example, reduction of biological pump due to  
46 reduced iron fertilization ([Watson et al., 2000](#); [Wolff et al., 2006](#); [Fischer et al., 2010](#);  
47 [Martinez-Garcia et al., 2014](#)); reduced CO<sub>2</sub> solubility due to increased temperature of the  
48 ocean ([Köhler and Fischer, 2006](#); [Fischer et al., 2010](#)); reduction of sea ice extent ([Fischer et](#)

49 [al., 2010](#)). As a consequence, the organization and extension of the Southern Ocean deep  
50 water-masses (i.e., Atlantic Meridional Overturning Circulation, AMOC; [Skinner et al., 2010](#);  
51 [Shakun et al., 2012](#); [Toggweiler et al., 2006](#)) were studied, mainly using paleo-hydrological  
52 proxies ( $\delta^{18}\text{O}$ ,  $\delta^{13}\text{C}$ ,  $\epsilon_{\text{Nd}}$  of sea water,  $\Delta^{13}\text{C}$ ; [Barker et al., 2009](#); [Vasquez-Riveiros et al., 2010](#);  
53 [Skinner et al., 2010, 2013, 2014](#); [Waelbroeck et al., 2011](#); [Gottschalk et al., 2015b](#)). By  
54 contrast, the terrigenous particle signal transported by these deep water-masses in the sub-  
55 Antarctic Atlantic Ocean remain poorly studied, even though locally it was successfully used  
56 to reconstruct the deep ocean circulation dynamical component ([Diekmann et al., 2000](#) –  
57 Scotia Sea; [Walter et al., 2000](#) – Scotia Sea and southwest South Atlantic; [Kuhn and](#)  
58 [Diekmann, 2002](#) – Cape Basin). In this study, we aim to reconstruct the past evolution of  
59 deep-water masses (i.e. contribution to sedimentation, vertical and horizontal extensions, and  
60 speeds) during the last deglaciation using the signal carried in the terrigenous particles  
61 transported by these water masses. Sediment grain size distribution allows identifying the  
62 contribution of distinct oceanic currents ([Weltje and Prins, 2007](#); [Weltje, 2012](#)), and  
63 reconstructing the speeds of the main deep water-masses ([McCave and Hall, 2006](#); [McCave](#)  
64 [and Andrew, 2019](#)). Grain-size distribution displays distinct modes ([Weltje, 2012](#)) that can be  
65 attributed to specific physical transport processes (e.g., wind, water currents, ice) and/or  
66 transport energy. Because grain-size distribution reflects both carrying vectors (e.g., wind,  
67 water currents, ice) and flow speed, several studies focused on the respective isotopic  
68 composition of distinct grain-size fractions in order to track distinct provenance patterns  
69 ([Meyer et al., 2011](#); [Bayon et al., 2015](#); [Beny et al., 2018](#)). Similarly, clay minerals are useful  
70 tracers of provenance, erosion balance, weathering, and transportation pattern of fine-grained  
71 sediments ([Moriarty et al., 1977](#); [Petschick et al., 1996](#); [Kuhn and Diekmann, 2002](#); [Montero-](#)  
72 [Serrano et al., 2010](#)). Finally, radiogenic isotope compositions of sediments can be used to

73 decipher the different provenances of terrigenous particles and determine their respective  
74 contributions through time (Walter et al., 2000; Bayon et al., 2009).

75 This study focuses on the South Atlantic Ocean, where northern water-masses (North  
76 Atlantic Deep Water, NADW) and southern ones (Lower Circumpolar Deep Water, LCDW  
77 and Antarctic Bottom Water, AABW) mix, because this area is key for understanding the  
78 atmosphere-ocean exchanges that control CO<sub>2</sub> sequestration/outgassing. The reorganization of  
79 these deep water-masses during the deglaciation is still a subject of debate and their impact on  
80 deglaciation is not resolved (Barker et al., 2009; Skinner et al., 2010, 2013; Waelbroeck et al.,  
81 2011; Adkins, 2013; Menviel et al., 2014, 2015a,b, 2018). Recent studies of the last  
82 deglaciation ventilation events in the South Atlantic Ocean demonstrated an apparent synergy  
83 between physical and biological processes (Gottschalk et al., 2016), whereas chemical and  
84 physical processes appear to be decoupled (Roberts et al., 2016). These contrasting results  
85 highlight the lack of constraints on the dynamical aspect of the water-masses mixing. In this  
86 paper, we combined grain-size distribution, clay mineralogy, and radiogenic isotope (Sr-Nd-  
87 Pb) analyses (on separated grain-size fractions) in order to investigate the evolution  
88 (horizontal and vertical extensions, pathways, speed) of deep-water-masses in the South  
89 Atlantic Ocean during the last termination.

90 II. General setting



91  
 92 *Figure 1: Simplified geological map of the study area. Light blue text names riverine*  
 93 *systems, dark blue text names seas, black text names oceanic basins, and white abbreviations*  
 94 *names areas. In South America: Guyana Highlands -GH-, Brazilian Highlands -BH-,*  
 95 *Argentinian Pampas -Pam-, Patagonia -Pat-; in Africa: Namibian deserts-Na-; in Antarctica:*  
 96 *Ellsworth Mountains -EM-, Antarctic Peninsula -AntP-, Pensacola Mountains -PM-, Berkner*  
 97 *Islands -BI-, Shackleton Range -SR-, Coats Land -CL-, Dronning Maud Land -DML-. Clay*

98 *mineral inputs to the South Atlantic Ocean derived from the studies of [Petschick et al.](#)*  
99 *(1996); [Diekmann et al. \(2000\)](#); [Walter et al. \(2000\)](#); [Guyot et al. \(2007\)](#); [Bayon et al.](#)*  
100 *(2016); and [Khondoker et al. \(2018\)](#).*

## 101 A. Deep water-masses

102 Modern South Atlantic deep circulation is characterized by the incursion of the  
103 NADW into the Southern Ocean, where it mixes/interacts with both the AABW and the  
104 Circumpolar Deep Water (CDW; [figure 1](#)). The southward NADW mainly flows as Deep  
105 Western Boundary Current along the South American Margin before leaving the continental  
106 shelf and flowing eastward around 45°S ([Reid, 1989](#); [Larqué et al., 1997](#); [Stramma and](#)  
107 [England, 1999](#)). A branch of the NADW crosses the Atlantic Ocean at the equator before to  
108 flow southward along the African margin and to reach the Indian Ocean via the Good Hope  
109 Cape ([Larqué et al., 1997](#)). In the Southern Ocean, NADW flows between two parts of the  
110 CDW: the Upper Circumpolar Deep Water (UCDW) flows above, whereas the Lower  
111 Circumpolar Deep Water flows underneath. The northern extension of the UCDW is very  
112 asymmetric with larger extent in the eastern part of the South Atlantic ([Larqué et al., 1997](#)).  
113 On the west side of the Mid-Atlantic Ridge (MAR), the LCDW penetrates into the Argentine  
114 Basin via the Drake Passage and flows northward into the Brazil Basin ([Speer and Zend,](#)  
115 [1993](#)). On the east of the MAR, the LCDW is a major oceanic feature in the Cape Basin,  
116 where it becomes diluted in the Angola Basin ([Larqué et al., 1997](#)). The lower part of the  
117 LCDW is currently mixed with the upper part of the AABW ([Orsi et al., 1999](#)) that is  
118 produced heterogeneously along the Antarctic Shelf ([Foster and Carmack, 1975](#); [Orsi et al.,](#)  
119 [1999](#)). [Foster and Camarck \(1975\)](#) demonstrated the formation of a very cold type of bottom  
120 water in the vicinity of the shelf break (the Polynyas mode) between 29 and 40°W and along  
121 the Antarctic Peninsula's tip, between 35 and 55°W and north of 65°S. Today this production  
122 occurs mainly in winter when seawater is close to freezing and becomes saltier due to sea ice

123 formation (Foster and Camarck, 1975; Orsi et al., 1999), which increases its density and  
124 promotes convection. This mode is called the Polynyas mode. However, in the Weddell Sea, a  
125 second mode of bottom water formation occurs: the Ice Shelf Water mode, which corresponds  
126 to bottom water formation under the Antarctic ice shelves due to super-cooling (Krueger et al.  
127 and references therein, 2012). The AABW is very thick in the Weddell-Enderby Basin (more  
128 than 4000 m; Orsi et al., 1999) and is partially exported to the Argentine Basin via the eastern  
129 Scotia Sea (east and west of the South Sandwich Islands) toward the Brazil Basin via the  
130 Vema Channel, and to the Agulhas and Crozet basins (Stramma and England, 1999). The  
131 AABW is a dominant oceanic feature south of 50°S although its' northern influence extents to  
132 around 10°S (Reid et al., 1989).

133         The characteristics of deep water masses were not constant through time and the  
134 evolution of the AMOC is controversial (Weber and Drijfhout, 2007; Vasquez Riveiros et al.,  
135 2010; Waelbroeck et al., 2011; Howe et al., 2016). Some studies have suggested that during  
136 the Last Glacial Maximum (LGM) the NADW was as strong as today but shallower (Curry  
137 and Oppo, 2005; Gherardi et al., 2009). Other studies suggest that it was weaker and  
138 shallower (Menviel et al., 2012; Howe et al., 2016). The NADW southern extension seems  
139 limited during glacial times, whereas southern-born water-masses extended further north  
140 (Kuhn and Diekmann, 2002; Skinner et al., 2013). Similarly, modifications of the Antarctic  
141 Circumpolar Current (ACC, including the CDW) during the last climatic cycle are still  
142 debated. Some studies reported severe modifications of the ACC intensity in the Atlantic  
143 (Noble et al., 2012) and Indian (Dezileau et al., 2000; Mazaud et al., 2010) sectors of the  
144 Southern Ocean: the activity of the ACC increased during the LGM, probably linked to  
145 enhanced westerly winds (Dezileau et al., 2000; Walter et al., 2000; Mazaud et al., 2010). In  
146 contrast, other studies support insignificant modification of the ACC in the Atlantic sector of  
147 the Southern Ocean (Gersonde et al., 2003) and Drake Passage (McCave et al., 2014). Finally,



148 the evolution of the AABW is also uncertain. [Krueger et al. \(2012\)](#) suggests that the  
149 production of AABW was similar during glacial and interglacial, but dropped during  
150 terminations (including the last deglaciation). In addition, [Menviel et al. \(2017\)](#) suggest that  
151 the “AABW” (corresponding to CDW and AABW in our study) was weaker during the LGM  
152 than today. In contrast, other studies suggest enhanced northward flow of AABW via deep  
153 western boundary currents in all three major ocean (Pacific: [Hall et al., 2001](#); [Govin et al.,](#)  
154 [2009](#); Indian: [Govin et al., 2009](#); Atlantic: [Govin et al., 2009](#); [Spooner et al., 2018](#)).

155 Any modifications of the AMOC have a potential effect on carbon storage/outgassing  
156 that is tightly controlled by ocean stratification/ventilation, itself resulting from contrasting  
157 physical and chemical properties of the water-masses. High ocean stratification – as during  
158 the last glacial – inhibits vertical mixing and isolates the deep ocean from the atmosphere. In  
159 contrast, ventilation events – breakdown of the stratification that occurred during the last  
160 deglaciation – allow CO<sub>2</sub> to upwell from the deep ocean and to be rapidly transferred to the  
161 atmosphere ([Toggweiler et al., 2006](#); [Barker et al., 2010](#); [Skinner et al., 2010, 2013, 2014](#);  
162 [Burke and Robinson, 2012](#)).

## 163 B. Geological setting

164 The Atlantic sector of the Southern Ocean may receive sediments from Antarctica, the  
165 Scotia Sea area (i.e., Scotia Arc, southern Patagonia, and northern Antarctic Peninsula) South  
166 America, and Africa. Here, we present the general geological structure of these areas, and  
167 their respective isotopic signatures ([figure 1](#)).

168 Antarctica corresponds to two distinct geological domains: East and West Antarctica.  
169 West Antarctica includes the Ellsworth Mountains (EM), the Antarctic Peninsula (Ant P), the  
170 Scotia Arc region (SCA). The geology of the Ellsworth Mountains ([figure 1](#)) is poorly known  
171 due to the lack of outcrop. These mountains are composed of Precambrian rocks,  
172 corresponding mainly to granodiorite orthogneiss intruded by granitic and pegmatitic sheets

173 (Tingey et al., 1991). The isotopic composition of the area is not known, but the age of the  
174 rock suggests radiogenic Sr and Pb isotope ratios, and unradiogenic  $\epsilon_{Nd}$  (figure 1). This area  
175 mostly delivers illite to the sea (Petschick et al., 1996; figure 1). The Antarctic Peninsula,  
176 composed of alkaline and calcalkaline magmatic rocks associated with deformed sediments of  
177 similar composition (Anderson, 1965; Tingey et al., 1991), is a potential source of terrigenous  
178 sediment for CDW and AABW. The Antarctic Peninsula is characterized by radiogenic  $\epsilon_{Nd}$ ,  
179 and low Sr and high Pb isotope ratios (figure 1). Clay minerals from this area mostly  
180 correspond to well-crystallized smectite (crystallinity around 1.5, Petschick et al., 1996;  
181 figure 1), and chlorite. The Scotia Arc islands rocks are similar to those of the Antarctic  
182 Peninsula but with slightly more radiogenic Nd, and slightly less radiogenic Sr and Pb (figure  
183 1). This area exports exclusively well-crystallized smectite to the ocean (crystallinity lower  
184 than one; Petschick et al., 1996).

185         The East Antarctic domain is an old shield with rocks  $> 0.9$  Ga. Rocks from the  
186 Pensacola Mountains (PM) to the Shackleton Range (SR) represent the Ross Orogen (i.e.,  
187 Transantarctic Mountains) and are mainly composed of varied metamorphic rocks aged from  
188 the Late Proterozoic to Late Paleozoic (Barrett, 1991; Laird, 1991; Tingey, 1991; Hauptvogel  
189 and Passchier and references therein, 2012). The isotopic composition of the area is  
190 characterized by unradiogenic Nd, with intermediate to high Sr isotope ratios (figure 1). The  
191 Pb isotope composition unknown, as is the case for most of East Antarctica rocks. The few  
192 studies reporting Pb isotope ratios have been carried out on feldspars only. The area releases  
193 mostly illite and chlorite (Petschick et al., 1996). Smectite is also exported from the Berkner  
194 Island, but is found only locally (Petschick et al., 1996). To the east, the Grenville Belts (920 to  
195 1330 Ma) outcrop on the Coats Land (CL) and western Dronning Maud Land (WDML) with  
196 a wide range of gneissic and clastic sediments (Capurro, 1955; Anderson, 1965; Tingey et al.,  
197 1991) with unradiogenic Nd, and radiogenic Sr isotope ratios (figure 1). Clay minerals

198 produced in this area are mainly chlorite and illite (Petschick et al., 1996). In the northernmost  
199 part of the Dronning Maud Land, the old Grunehogna Craton is composed mainly of Archean  
200 granitoids (Tingey et al., 1991; Mieth and Jokat, 2014; Pierce et al., 2014) with very  
201 unradiogenic Nd ( $\epsilon_{Nd} \ll -20$ ) and radiogenic Sr ratios (figure 1). The clay minerals generated  
202 in that area are mainly chlorite and illite (Petschick et al., 1996).

203 South America can be divided into two blocks: the recent Andean Orogen, and the old  
204 South American Shield. In the Atlantic Sector of the Southern Ocean, the Andean Orogen  
205 mostly corresponds to Patagonia (Pat). The geology and the isotopic composition of this  
206 region is similar with the Antarctic Peninsula: radiogenic Nd, unradiogenic Sr and Pb isotope  
207 ratios, but with higher  $^{207}\text{Pb}/^{204}\text{Pb}$  than the Antarctic Peninsula (figure 1; Olivero and  
208 Martinioni, 2001; Sayago et al., 2001 and references therein; Smith et al., 2003; Khondoker et  
209 al., 2018). Even if the climate is slightly more humid than in the Antarctic Peninsula,  
210 Patagonia also produces mainly well-crystallized smectite (Petschick et al., 1996; Desiage et  
211 al., 2018; Khondoker et al., 2018).

212 The South American Shield is composed of several cratonic areas separated by vast  
213 ancient orogenic belts. (Cordani and Sato, 1999; Engler, 2009; Oyhantçabal and Siegesmund,  
214 2011; Siegesmund et al., 2011; de Alkmim, 2015). Cratonic areas are mainly composed of  
215 granitoids and a wide range of metamorphic rocks, while orogenic belts are composed of a  
216 large variety of metamorphic rocks with high original composition (from the granite to the  
217 ophiolite and the juvenile magmatic arc; Almeida et al., 1981; Barros et al., 1982; Moura and  
218 Gaudette, 1993; Neves, 2003; de Alkmim, 2015; Nogueira et al., 2007; Pimentel et al., 2011;  
219 Fuck et al., 2014; McGee et al., 2015; D'el-Rey Silva et al., 2016). The overall isotopic  
220 composition of the South American Shield is radiogenic Sr and Pb, and very unradiogenic Nd  
221 isotope ratios (figure 1). South of the South American Shield, a large tholeiitic-basaltic  
222 province is known as the Parana basalt (Waichel et al., 2012; Milani and Ramos, 2017). The

223 composition of the Parana basalts differs from the South American Shield with radiogenic Nd,  
224 and very low  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio ( $<0.5$ ; [Tapani Ramo et al., 2016](#)).

225 This South American Shield is drained by important riverine systems such as the  
226 Amazon and Parana rivers. The Amazon drains sediments from the Andes, as well as the  
227 South American Shield. The South American Shield is characterized by rocks with  
228 unradiogenic Nd and radiogenic Sr and Pb ([figure 1](#)). The clay mineral exported to the ocean  
229 is mainly kaolinite ([Petschick et al., 1996](#); [Guyot et al., 2007](#); [Bayon et al., 2015](#)). However,  
230 sediments are mainly exported northwards toward the Caribbean Sea. The Parana River drains  
231 the southern part of the South American Shield, the Parana basalts, and the north of  
232 Patagonia. Its isotopic composition has intermediate Nd, Sr, and Pb isotope ratios ([figure 1](#)).  
233 Clay minerals transported via this river are illite ( $\sim 60\%$ ; [Petschick et al., 1996](#); [Campodonico](#)  
234 [et al., 2016](#)), kaolinite, and smectite. Other smaller rivers drain the south of Brazil, and mainly  
235 deliver sediments characterized by unradiogenic Nd, and radiogenic Sr and Pb isotope ratios,  
236 containing mainly smectite derived from the Parana basalts and kaolinite ([figure 1](#); [Petschick](#)  
237 [et al., 1996](#)). On the other hand, Patagonia and Pampas are dry and windy areas ([Clapperton,](#)  
238 [1993](#)). Consequently, part of the detrital particles is exported to the ocean via aeolian  
239 processes as loess ([Mathias et al., 2014](#)). The overall isotopic composition of this smectite-  
240 rich loess is radiogenic Nd, unradiogenic Pb and Sr ratios. According to [Mathias et al. \(2014\)](#),  
241 the sediment delivery to the southwest Atlantic is currently dominated by loess, and by the  
242 Parana River sediments.

243 A large proportion of the African continent comprises Archean Cratons (western  
244 Africa, Congo), and rocks related to the Pan African Orogen. The cratons are composed of  
245 gneisses while the Pan African Orogen is mainly composed of flysch deposits associated with  
246 volcanic formations, and intruded by granitoids. The overall geochemical composition of  
247 Africa is intermediate to very unradiogenic Nd, and intermediate to very radiogenic Sr and Pb

248 isotope ratios (figure 1). Clay minerals formed in the Congo depression are mainly poorly  
249 crystallized smectite and kaolinite, whereas the Namibian deserts release illite-rich loess  
250 (Bremner and Willis, 1993; Petschick et al., 1996; Bayon et al., 2015; figure 1).

### 251 III. Material and methods

#### 252 A. Material

253 This study is based on sediment samples from core MD07-3076Q (44°9.2'S,  
254 14°13.7'W) recovered by the French Vessel *Marion Dufresne* in 2007 on the eastern flank of  
255 the Mid-Atlantic Ridge at 3770 m depth (figure 1). The core chronology is based on  
256 planktonic foraminifer radiocarbon dates (30 measurements for the studied interval) and  
257 stratigraphic control points (Skinner et al., 2010, 2013, 2014; Gottschalk et al, 2015b). Today,  
258 the site is bathed in the LCDW. From the seafloor to the surface, the water masses observed  
259 in the water column above the study site are the LCDW, NADW, UCDW, and intermediate  
260 and surface waters. In nearby deeper oceanic basins, the AABW is also found, and it is  
261 thought that particles transported by this bottom water can reach our site by advection and  
262 mixing with the LCDW.

#### 263 B. Sample preparation and analytical procedure

264 All samples were first decarbonated using 0.1N HCl and then deflocculated by  
265 repeated washing using pure water.

##### 266 1. Grain-size distribution

267 Grain-size distribution was measured on carbonate-free sediment using a Malvern  
268 Mastersizer 2000 laser (0.02-2000  $\mu\text{m}$ ) following standard protocols (Montero-Serrano et al.,  
269 2009). The clay-size fraction corresponds to the proportion (in volume) of particles with a  
270 diameter smaller than 2  $\mu\text{m}$ , the cohesive silts correspond to the 2 to 10  $\mu\text{m}$  diameter

271 particles, the sortable silts correspond to the 10 to 63  $\mu\text{m}$  diameter particles, and the sand  
272 fraction to the particles coarser than 63  $\mu\text{m}$ .

## 273 2. Clay mineralogy

274 The clay-size fraction ( $< 2\mu\text{m}$ ) was isolated by differential settling according to  
275 Stoke's Law. This fraction was then concentrated by centrifugation (40 minutes, 3500 rotation  
276 per minute) and an aliquot was placed on glass slides. X-ray diagrams were obtained using a  
277 Bruker D4Endeavor (standard 30 kV and 30 mA). Each sample is run three times between  
278  $2.49$  and  $32.5^\circ 2\theta$ : (1) dried sample, (2) glycolated sample (12 hours in ethylene glycol), and  
279 (3) heated sample ( $490^\circ\text{C}$  for 2 hours). The proportion of each clay mineral (smectite,  
280 chlorite, illite, and kaolinite) is determined using the main X-ray diffraction peaks (layer and  
281 interlayer) on X-ray spectra according to their crystallographic characteristics (Brown, 1980).  
282 Smectite is characterized by a main peak at  $14\text{-}15 \text{ \AA}$  on the air-dried samples that expands at  
283  $17 \text{ \AA}$  after ethylene glycol saturation and collapses at  $10 \text{ \AA}$  after heating. Illite is characterized  
284 by a main peak at  $10 \text{ \AA}$  on the three runs. Kaolinite and chlorite are measured based on their  
285 main peak at  $7\text{-}7.2 \text{ \AA}$  and further distinguished using pseudo-void deconvolution of the  
286 kaolinite/chlorite doublet at  $3.54/3.58 \text{ \AA}$ . All measurements and semi-quantitative estimations  
287 are processed using the McDiff software (Petschick, 2000). The index of Esquevin (Esquevin,  
288 1969) corresponds to the ratio between the intensity of the peak of illite at  $5 \text{ \AA}$  and at  $10 \text{ \AA}$   
289 (illite  $5\text{\AA}/10\text{\AA}$ ). In this study, the crystallinity of illite and smectite has been measured using  
290 the method of integral breath to enable comparison with the work of Petschik et al. (1996) in  
291 the South Atlantic. High crystallinity (high values of the index) corresponds to badly  
292 crystallized mineral.

## 293 3. Isotope Analysis

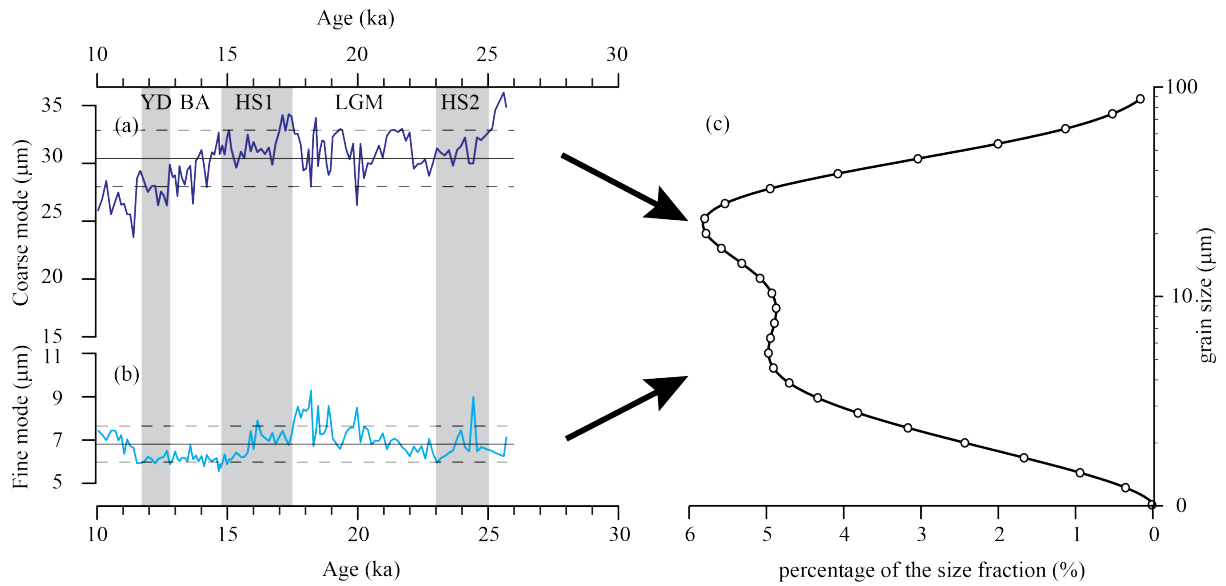
294 The grain-size distribution recognized three distribution modes. These three grain-size  
295 fractions were then isolated in order to examine their provenance. The clay-size fraction (0-

296 2  $\mu\text{m}$ ) was isolated by repeated extractions by settling. The coarse fraction (20-40  $\mu\text{m}$  in the  
297 old section of the core before 11.5 ka cal BP, noted hereafter as ka, and 20-32  $\mu\text{m}$  after 11.5  
298 ka) was isolated by sieving the clay-free residue. The 2-20  $\mu\text{m}$  grain-size fraction is thus the  
299 residue of clay separation and coarse fraction sieving. An aliquot of 10 to 150 mg (depending  
300 of the sediment quantities available) was then used for chemical procedures.

301 Chemical and isotopic measurements were performed at the Faculty of Science of the  
302 Vrije Universiteit, Amsterdam. Sediment was digested in closed Savillex<sup>TM</sup> Teflon beakers  
303 using a mixture of concentrated HF and HNO<sub>3</sub> (~2/3 and 1/3, respectively) for a few days at  
304 120°C. After digestion, samples were dried out and re-dissolved in ~6.5N HCl to remove  
305 fluorides. Nd, Sr, and Pb were purified using conventional ion chromatography. Pb isotopes  
306 were processed first and purified using AG-X8 200-400 mesh resin. Light rare earth elements  
307 (LREE) were isolated using true-spec resin medium (Eichrom<sup>TM</sup>, 100-150  $\mu\text{m}$  mesh size) and  
308 Nd was eluted from the LREE fraction using LN-resin (Eichrom<sup>TM</sup>, 50-100  $\mu\text{m}$  mesh size). Sr  
309 was recovered using Sr resin column. Nd and Pb measurements were performed by bracketing  
310 using respectively CIGO and NBS981 standards on Thermo Scientific Neptune multi-  
311 collector inductive coupled plasma-mass spectrometer (MC-ICPMS), whereas Sr  
312 measurements were performed on a thermal ionization multi-collector-MS (TIMS). Due to  
313 relatively low content of Nd in some of the coarse grain-size fraction (20-32/40 $\mu\text{m}$ ), Nd  
314 measurements of these fractions were also performed by TIMS.

## 315 IV. Results

### 316 A. Grain-size distribution



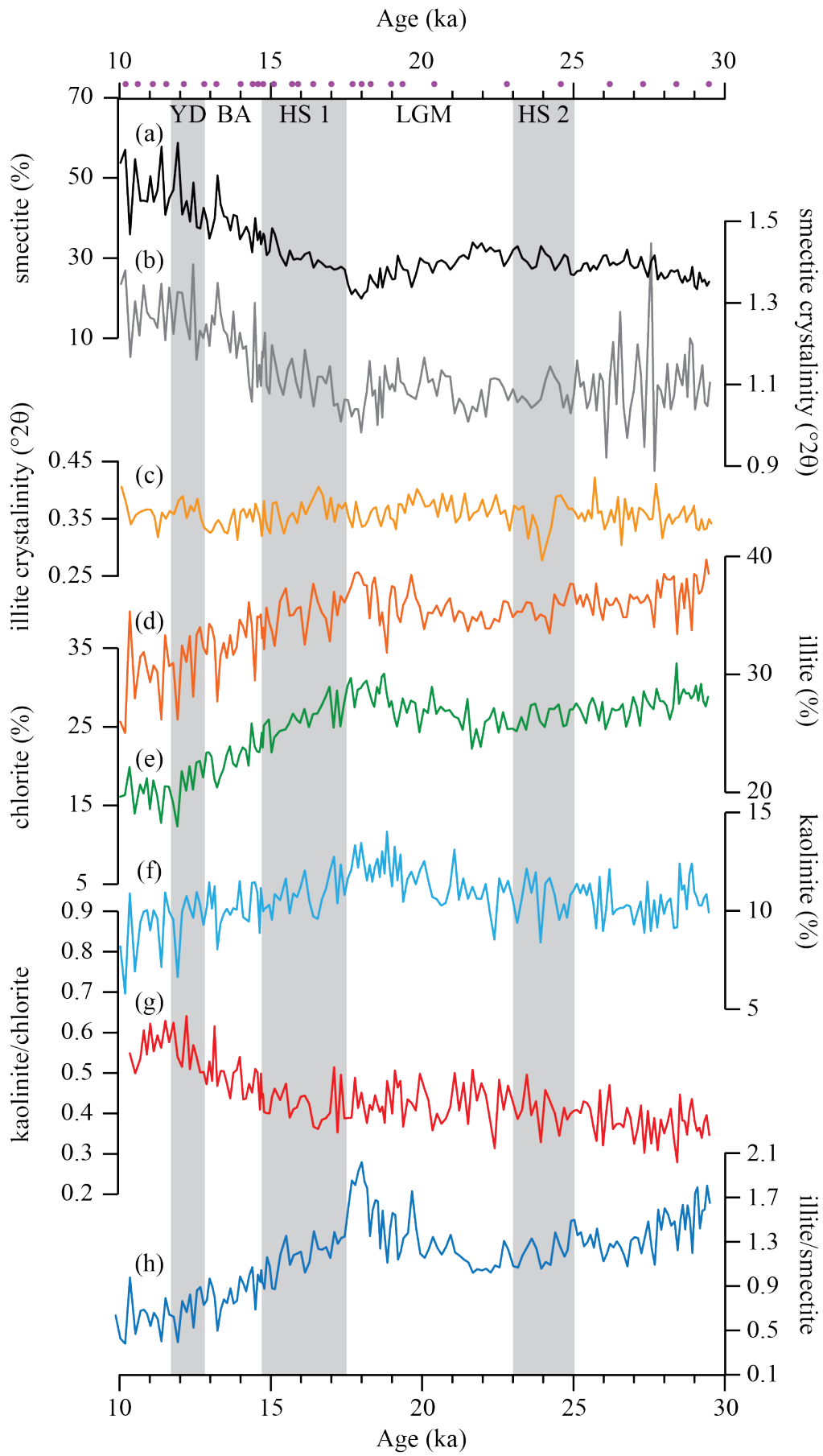
317  
318 *Figure 2: Grain-size results from core MD07-3076Q. (a) and (b) Variation of the coarse*  
319 *and fine grain size mode values, respectively, through time. Black lines represent the average*  
320 *value for both grain-size modes while dashed lines represent their standard deviation; (b)*  
321 *typical grain-size distribution curve from the core MD07-3076Q. Two grain-size modes can*  
322 *be identified from the grain-size distribution curve. YD: Younger Dryas, BA: Bølling Allerød,*  
323 *HS: Heinrich Stadial; LGM: Last Glacial Maxima.*

324 The grain-size distribution is bimodal throughout the core (figure 2) with a coarse  
325 mode ranging in the sortable silts fraction (from 23.6 and 36.2 µm, average: 30.4, standard  
326 deviation: 2.48; table 1) and a fine mode ranging in the cohesive silts fraction (from 5.6 and  
327 9.3 µm, average: 6.8, standard deviation: 0.8; table 1). The sortable silt (10-63 µm) fraction is  
328 dominant (41 to 52% in volume of the total terrigenous fraction, average: 47%, standard  
329 deviation: 2.4; table 1) and the cohesive silts (2-10 µm) represent 35 to 45% (average: 39%,  
330 standard deviation: 2.4; table 1). The clay (<2 µm) and sand fractions (>63 µm) represent



331 respectively 2 to 5% (average: 3.4%, standard deviation: 0.7; [table 1](#)) and 1 to 6% (average:  
332 3.0, standard deviation: 1.0; [table 1](#)) of the total terrigenous fraction ([table 1](#)).

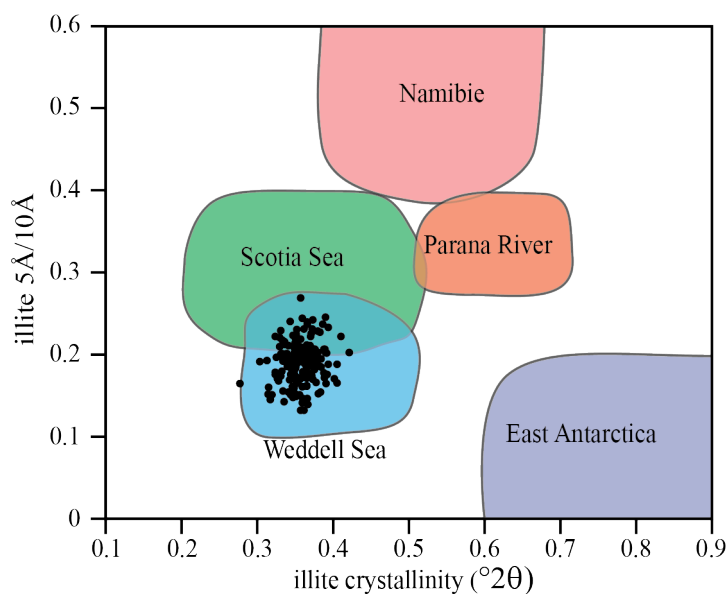
333 In details, the mean of the coarse mode (hereafter mentioned as the sortable silt mode)  
334 is slightly coarser during the LGM ([figure 2.a](#)), and decreases during the deglaciation and  
335 reaches its minimum values after the Younger Dryas (YD). The average size of the fine mode  
336 (hereafter mentioned as the cohesive silt mode) is maximum at the end of the LGM and  
337 displays its lowest values during the deglaciation ([figure 2.b](#)).



339 *Figure 3: Clay mineral data from core MD07-3076Q. Purple dots represent the*  
340 *radiocarbon dates used for the age model. (a) relative proportion of smectite (%); (b)*  
341 *smectite crystallinity ( $^{\circ}2\theta$ ); (c) illite crystallinity ( $^{\circ}2\theta$ ); (b) and (c) note that the lower is the*  
342 *crystallinity index, the better the mineral is crystallized; (d) relative proportion of illite (%);*  
343 *(e) relative proportion of chlorite (%); (f) relative proportion of kaolinite; (g)*  
344 *kaolinite/chlorite ratio (K/C); (h) illite/smectite ratio (I/S).*

## 345 **B. Clay Mineralogy**

346 The clay mineral assemblage is composed of smectite ((S) 18-59%; [figure 3.a](#)), illite  
347 ((I) 21-40%; [figure 3.d](#)), chlorite ((C) 12-33%; [figure 3.e](#)), and kaolinite ((K) 5-14%; [figure](#)  
348 [3.f](#)) with higher smectite content at the end of the deglaciation, and higher illite, chlorite and  
349 kaolinite contents during the LGM. The smectite crystallinity ranges from 0.89 and 1.45  
350 ([figure 3.b](#)) and varies in phase with the smectite content with slightly higher variability  
351 between 25 and 30 ka. The illite crystallinity varies between 0.27 and 0.42 ([figure 3.c and 4](#))  
352 without major variation. Clay mineral proportions are relatively stable during the LGM but  
353 records a shift during the deglaciation with an increase of smectite compared to the other clay  
354 minerals. The smectite crystallinity varies simultaneously with smectite proportion changes.  
355 The K/C ratio is stable during the LGM, and increases during the deglaciation. The I/S ratio  
356 presents two maxima at the beginning of the record and at the end of the LGM, sharply  
357 decreases at the beginning of the HS 1, and steadily decreases during the deglaciation.



358

359 *Figure 4: Crystallographic properties of illite from site MD07-3076Q (black dots) and*  
 360 *comparison with properties of illite from potential sources to the South Atlantic basin. The*  
 361 *data used for comparison are from [Petschick et al., \(1996\)](#). Illite 5Å/10Å ratio gives*  
 362 *information on the chemical properties of illite (Fe-Mg- vs Al-rich illite) while the*  
 363 *crystallinity gives clues on the quality of the mineral crystal (high crystallinity: badly*  
 364 *crystallized mineral).*

### 365 C. Radiogenic Isotopes

366 Isotopic results are presented for the sortable silt mode fraction (20-32/40 μm) and the  
 367 cohesive silt mode fraction (2–20 μm) and for the clay-size fraction (<2 μm) ([tables 2 to 4](#)).  
 368 Lead isotope ratios were only been measured on the clay-size fraction ([table 2](#)).

#### 369 1. Clay-size fraction (<2 μm)

370 The  $\epsilon_{Nd}$  of the clay-size fraction ranges from -7.2 to -4.8. It is relatively stable during  
 371 the last glacial ( $<\pm 0.5 \epsilon_{Nd}$ ), decreases by 2  $\epsilon$  units during the deglaciation and shows  
 372 minimum values at the end of the deglaciation ([figure 5.c](#)). The  $^{87}Sr/^{86}Sr$  ratio ranges from  
 373 0.711 to 0.712: it is rather high and stable ( $<\pm 0.0003$ ), during the last glaciation and displays  
 374 a rapid decrease at ~15 ka. The ratio records a short rebound at 12.7 ka before reaching its

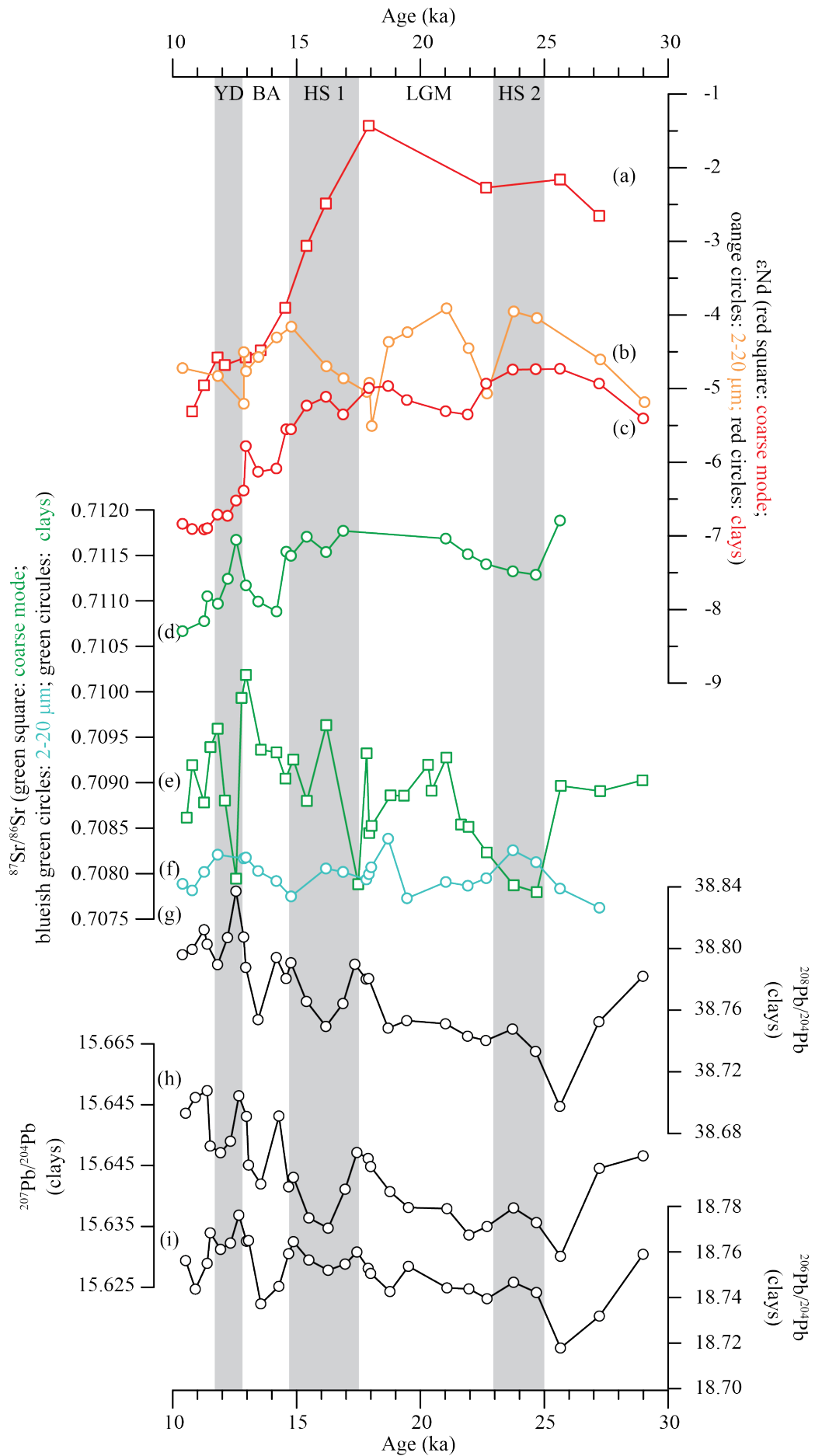
375 minimum values at the end of the deglaciation (0.7107; [figure 5.d](#)). The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio and the  
376  $\epsilon_{\text{Nd}}$  show a similar temporal evolution (they both decrease during the deglaciation), which is  
377 surprising as in nature they usually record opposite relationship. The three Pb isotope ratios  
378 record their lowest values around 25 ka, and increase thereafter. This trend is more  
379 pronounced for the  $^{207}\text{Pb}/^{204}\text{Pb}$  ratio during the deglaciation. Some fluctuations of the three  
380 ratios are observed during the deglaciation with higher ratios at climatic transitions  
381 (LGM/Heinrich Stadial 1 –HS1–, HS1/Bølling-Allerød – B-A –, B-A/Younger Dryas –YD–,  
382 Holocene onset), and low values during the middle of the HS1, the middle of the B-A, and the  
383 second half of the YD (except for the  $^{206}\text{Pb}/^{204}\text{Pb}$ ). We note that the Sr isotope rebound  
384 observed at 12.7 ka is also associated with variations Pb isotope ratios.

## 385 2. Cohesive silt mode fraction (2-20 $\mu\text{m}$ )

386 The  $\epsilon_{\text{Nd}}$  of the 2-20  $\mu\text{m}$  grain-size fraction (cohesive silt mode fraction) ranges from -  
387 5.5 to -3.9 with the lowest values at 13, 18, 23 and 29 ka ([figure 5.b](#)). The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios are  
388 generally constant with higher values around 13, 19, and 24 ka ([figure 5.f](#)).

## 389 3. Sortable silt mode fraction: (20-32/40 $\mu\text{m}$ )

390 The  $\epsilon_{\text{Nd}}$  of the sortable silt mode fraction ranges from -4.9 to -1.4. Values are  
391 relatively unchanging during the last glacial period ( $-2 \pm 0.5 \epsilon_{\text{Nd}}$ ) but decreases after 18 ka and  
392 reaches a minimum (-5) at the end of the deglaciation ([figure 5.a](#)). The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio ranges  
393 from 0.708 and 0.710.  $^{87}\text{Sr}/^{86}\text{Sr}$  values are lower during the HS 2, at the beginning of the HS  
394 1, and at the beginning of the YD, while the highest value is recorded at the B-A offset ([figure](#)  
395 [5.e](#)).



397 *Figure 5: Isotopic results from core MD07-3076Q. (a)  $\epsilon_{Nd}$  from the sortable silt mode*  
398 *fraction (20-32/40  $\mu\text{m}$ ); (b)  $\epsilon_{Nd}$  from the cohesive silt mode fraction (2-20  $\mu\text{m}$ ); (c)  $\epsilon_{Nd}$  from*  
399 *the clay-size fraction (<2  $\mu\text{m}$ ); (d)  $^{87}\text{Sr}/^{86}\text{Sr}$  from the clay-size fraction (<2  $\mu\text{m}$ ); (f)  $^{87}\text{Sr}/^{86}\text{Sr}$*   
400 *from the sortable silt mode fraction (20-32/40  $\mu\text{m}$ ); (g)  $^{87}\text{Sr}/^{86}\text{Sr}$  from the cohesive silt mode*  
401 *fraction (2-20  $\mu\text{m}$ ); (h)  $^{208}\text{Pb}/^{204}\text{Pb}$ , (i)  $^{207}\text{Pb}/^{204}\text{Pb}$ , and (j)  $^{206}\text{Pb}/^{204}\text{Pb}$  from the clay-size*  
402 *fraction (<2  $\mu\text{m}$ ).*

## 403 V. Discussion

### 404 A. Sediment provenance

#### 405 1. Clay size fraction: an overall provenance signal

406 To determine the provenance of this grain-size fraction, both clay mineral proportions  
407 and radiogenic isotopes have been used. Kaolinite and chlorite have distinct provenances: at  
408 present, kaolinite is mainly found at low latitudes where chemical weathering is strong.  
409 Brazilian rivers (Amazon and Parana), as well as equatorial and tropical African rivers are  
410 characterized by high contents of kaolinite (figure 1). Inherited kaolinite is also found in East  
411 Antarctica marine sediments (figure 1), but is transported by the westward Antarctic Coastal  
412 Current and remains trapped in the Weddell-Enderby Basin due to the Weddell Gyre  
413 circulation (Petschick et al., 1996) and this kaolinite is therefore unlikely to contribute to  
414 sedimentation at the studied site. Consequently, kaolinite can be used to trace the low latitude  
415 provenance. Conversely, presently, chlorite mainly characterizes high latitude regions, where  
416 weak chemical weathering allows its preservation. It is abundant in Antarctica and the Scotia  
417 Sea (figure 1). Consequently, chlorite can be considered a tracer of high latitude provenance.  
418 Thus, the kaolinite/chlorite ratio (K/C) discriminates southern-born versus northern-born  
419 material in the southern Atlantic (Petschick et al., 1996; Kuhn and Diekmann., 2002; Kruger  
420 et al., 2008).

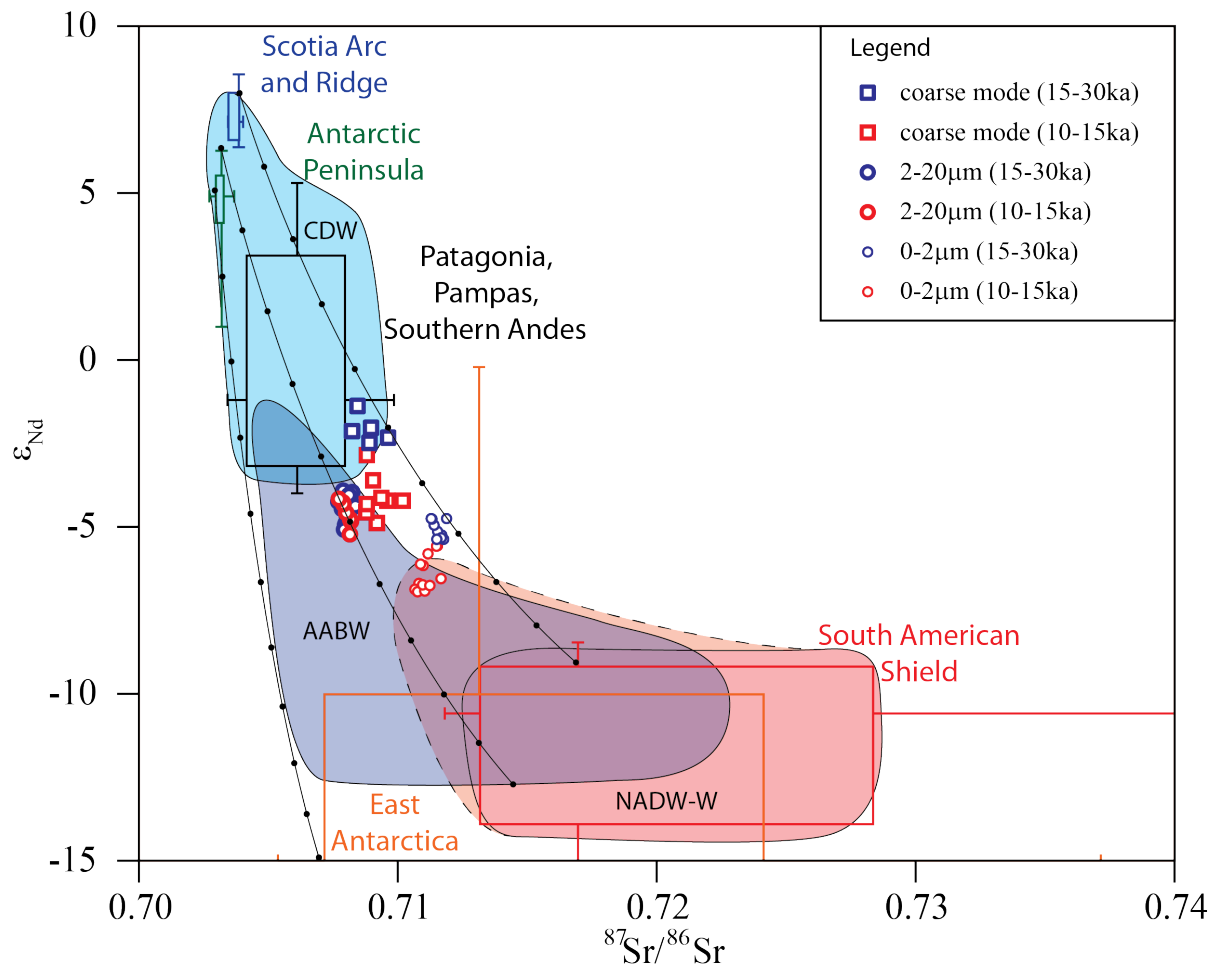
421 Illite provenance is more complex. It is very abundant in the Weddell Sea, Namibian  
422 deserts, the Parana River (60% of the delivered sediment), and it is also present, although at  
423 lower concentrations (<30%), in the Scotia Sea (figure 1). To decipher the provenance of  
424 illite, we used the chemical (5/10Å ratio, i.e., Fe-Mg vs. Al rich illite) and crystallographic  
425 properties (crystallinity) of this mineral. Data are compared with the 5/10Å ratio and the  
426 crystallinity of illite of potential illite sources from the database of Petschick et al. (1996)  
427 (figure 4). This comparison demonstrates that the illite from core MD07-3076Q is similar to  
428 illite from the Weddell and Scotia seas while the chemical and crystallographic properties of  
429 the illite recovered at site MD07-3076Q is not compatible with a Namibian source. In the  
430 Scotia Sea, illite represents less than 30% of the total clay mineral fraction (from 10-30%;  
431 Petschick et al., 1996) and represents a mix of illite coming from the Weddell Sea embayment  
432 via the AABW, and illite from the Pacific sector of Antarctica delivered by the CDW  
433 (Diekmann et al., 2000; Walter et al., 2000; Petschick et al., 1996). Importantly, in the  
434 Weddell Sea, illite represents at least 50% of the total clay mineral fraction (more than 80% in  
435 some areas; Petschick et al., 1996). Consequently, the Weddell Sea embayment is considered  
436 as the major, source of illite for our studied site.

437 Smectite has several potential provenances: the Angola Basin, southern South  
438 America, Scotia Sea Islands, and Antarctic Peninsula. The crystallinity of this mineral was  
439 measured in order to decipher its provenance. The well-crystallized smectite (low  
440 crystallinity; figure 3) is compatible with a provenance from southern South America  
441 (Pampas and Patagonia), Scotia Arc, or Antarctic Peninsula, which are all characterized by  
442 crystallinity lower than 1.5 (Petschick et al., 1996; figure 3). It rules out a contribution of  
443 poorly crystallized smectite (crystallinity > 2; Petschick et al., 1996) from Angola. In detail,  
444 the crystallinity of the smectite from our studied site likely corresponds to a mix between two  
445 types of sources: (1) Patagonia and Antarctic Peninsula with a crystallinity of 1.5; and (2)



446 Scotia Arc with a crystallinity lower than 1.0. Interestingly, the proportion and crystallinity of  
447 smectite record very similar variations that suggest a dominant and rather continuous flux of  
448 smectite from the Scotia Arc (low crystallinity - ~1.1) during the last glacial period, with a  
449 supplementary supply of less well crystallized smectite (slightly higher crystallinity – 1.5;  
450 ~20%) from Patagonia or Antarctic Peninsula during the deglaciation ([supplementary figures](#)  
451 [S1 and S2](#))

452 In summary, both the proportions of the clay minerals and their crystallographic  
453 properties suggest that kaolinite mainly derives from low latitude areas, chlorite originates  
454 from Antarctica and the Scotia Sea, while illite mainly derives from the Weddell Sea  
455 embayment and smectite is derived from several sources (Scotia Arc, Patagonia, and  
456 Antarctic Peninsula). Moreover, mineralogical and crystallographic properties exclude Africa  
457 as a source of clay minerals consistent with the dominant westward oceanic circulation in the  
458 ACC ([figure 1](#)).

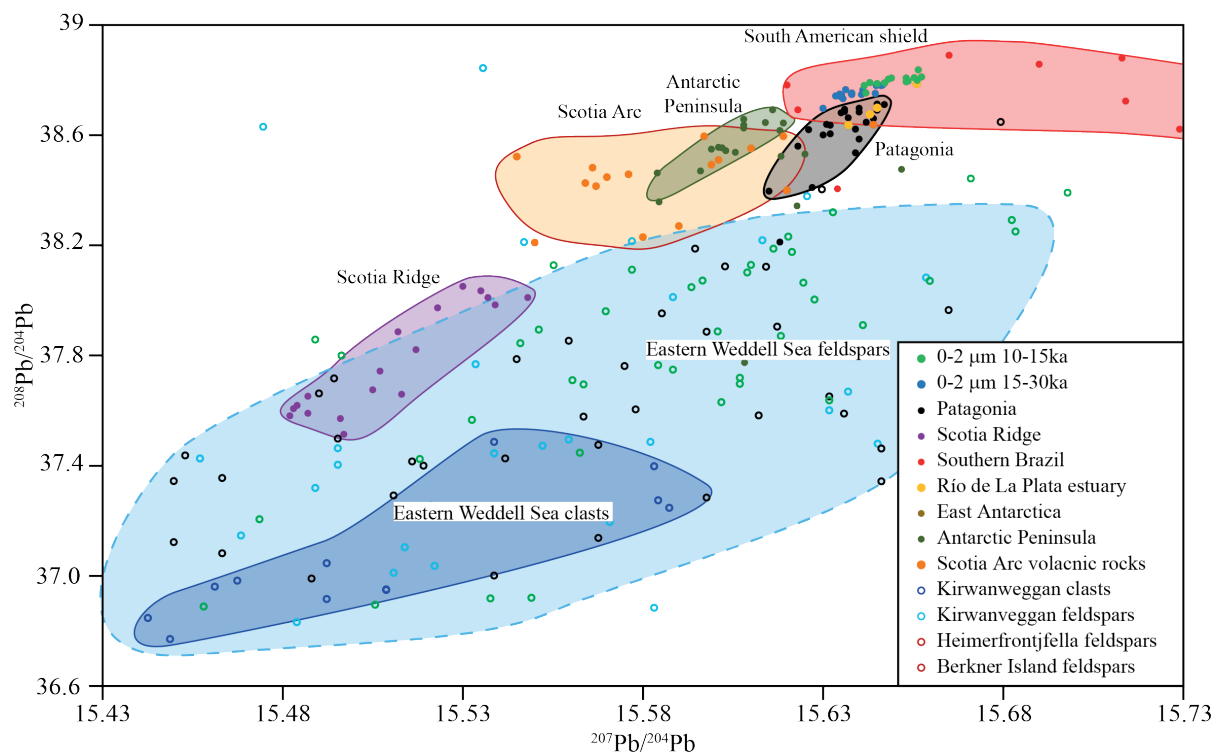


459

460 *Figure 6:  $^{87}\text{Sr}/^{86}\text{Sr}$ - $\epsilon_{\text{Nd}}$  diagram of MD07-3076Q sediments and of potential sources of*  
 461 *sediments for the South Atlantic Basin. Boxes representing the reference data delimit the first*  
 462 *and the third quartile. The median values are represented by horizontal and vertical lines*  
 463 *attached to the boxes for the  $\epsilon_{\text{Nd}}$  and the  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios, respectively. The short strokes at the*  
 464 *end of these lines represent the 10<sup>th</sup> and 90<sup>th</sup> percentile values. Accordingly, the box represent*  
 465 *50% of values found for each potential provenance, and 80% is represented with the lines.*  
 466 *Data sources: -South American Shield- Allegre et al. (1996); Basile et al. (1997); Walter et*  
 467 *al. (2000); de Mahiques (2008); Jeandel et al. (2007), and references therein; Mallmann et*  
 468 *al. (2007); Ganade de Araujo et al. (2014); Fuck et al. (2014); -Patagonia, Pampas, Southern*  
 469 *Andes- Stern et al. (1990); Grousset et al. (1992); Basile et al. (1997); Walter et al. (2000);*  
 470 *Smith et al. (2003); Jeandel et al. (2007), and references therein; Khondoker et al. (2018); -*  
 471 *Scotia Arc and Ridge- Jeandel et al. (2007); Harrison et al. (2003);-Antarctic Peninsula-*

472 *Hole et al. (1993); Kosler et al. (2009); Roy et al. (2007); Lee et al. (2005); -East Antarctica-*  
 473 *DePaolo et al. (1982); Luttinen et al. (1998, 2010); Roy et al. (2007); Will et al. (2010).*

474 The coupled  $\epsilon_{Nd} - ^{87}Sr / ^{86}Sr$  isotope variation of the clay-size fraction (figure 6) reveal  
 475 a mixed provenance between young sources around the Scotia Sea (Scotia Arc, Antarctic  
 476 Peninsula, Patagonia) and southern South America (Patagonia and Pampas), and an old source  
 477 from East Antarctica or the South American Shield (i.e., Brazil and Parana River).



478  
 479 *Figure 7: Pb isotope composition of core MD07-3076Q and of potential sources of*  
 480 *sediments. The reference data used for comparison are from Fretzdorff et al. (2002), -Scotia*  
 481 *Ridge-; Khondoker et al. (2018), -Patagonia-; Cohen and O’Nions (1982); Barreiro (1983), -*  
 482 *Scotia Arc-; Lee et al. (2005), Kosler et al. (2009), Flowerdew et al. (2012) -Antarctic*  
 483 *Peninsula-, de Mahiques et al. (2007), -South American Shield-; Flowerdew et al., (2012), -*  
 484 *Eastern Weddell Sea-*.

485 The coherent Pb isotope variations in figure 7 support this interpretation recording a  
 486 mix between the Antarctic Peninsula, Scotia Arc, and South American Shield. However, since

487 very few data are available for the Pb isotope compositions of East Antarctica, a contribution  
488 from this area cannot be excluded. The three isotopic systems suggest an increased  
489 contribution of older material during the deglaciation compared with the LGM. Based solely  
490 on Nd and Pb isotope data, it is rather difficult to distinguish South America and East  
491 Antarctica as the old end-members during the deglaciation. However, the increase of the K/C  
492 ratio during the deglaciation supports the hypothesis of enhanced relative contribution from  
493 the South American Shield.

#### 494 2. Cohesive silt mode fraction (2-20 $\mu\text{m}$ )

495 Both  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\epsilon_{\text{Nd}}$  of the cohesive silt mode fraction record limited variation  
496 throughout the deglaciation (figure 5). The isotopic results indicate that the provenance is  
497 located in Patagonia and in the Antarctic Peninsula, with an additional contribution of old  
498 source (South American Shield and/or East Antarctica). These results suggest that no major  
499 modification of provenance occurred during the deglaciation. Strontium and Nd isotope ratios  
500 do not allow discrimination of the source of old material for the cohesive silt mode fraction.  
501 The Nd isotope ratios suggest on average a higher radiogenic contribution for cohesive silts  
502 compared with the clay-size fraction (figure 6).

#### 503 3. Sortable silt mode fraction (20-32/40 $\mu\text{m}$ )

504 Both Sr and Nd isotope ratios of the sortable silt during the LGM record one dominant  
505 provenance: the Scotia Sea area (Patagonia, Scotia Arc, Antarctic Peninsula; figure 5). During  
506 the deglaciation, an important shift to less radiogenic  $\epsilon_{\text{Nd}}$  values (4 units) is observed implying  
507 an increased contribution of an older end-member such as East Antarctica and/or the South  
508 American Shield. However, because East Antarctica and the South American shield have  
509 similar Sr and Nd isotope composition, it is not possible to distinguish them with these  
510 isotopes (figure 6). Consequently, investigating the main transportation patterns is necessary

511 to identify the source of old material for both the cohesive (2-20  $\mu\text{m}$ ) and the sortable silt  
512 mode fractions (20-32/40  $\mu\text{m}$ ).

513 In summary, (figure 6) all three grain-size fractions compositions are located between  
514 two end-members: the Scotia Sea-southern South America area (Antarctic Peninsula, Scotia  
515 Arc and Ridge, Patagonia, Pampas, Southern Andes), and crustal shields (East Antarctica,  
516 South American Shield). The clay-size fraction appears the most crustal-like, whereas the  
517 sortable silt mode presents the most mantle-like composition. The 2-20  $\mu\text{m}$  grain-size fraction  
518 presents the lowest  $^{87}\text{Sr}/^{86}\text{Sr}$ , meaning that the difference with the other grain size fraction is  
519 related to distinct provenances rather than grain size effect. Both the clay size and the sortable  
520 silt mode fractions display an increase of the crustal end-member. In the clay size fraction,  
521 clay mineralogy enabled the identification of this old crustal source (i.e., the South American  
522 Shield).

## 523 B. Transportation pattern and palaeoclimatic implications

524 The provenances of the sediments for the three distinct grain-size fractions have been  
525 identified (or partially identified) in the previous section using their mineralogical and  
526 geochemical characteristics. Now, we will focus on identifying their transportation patterns  
527 and vectors from the source to the sedimentation site.

528 Several studies propose that aeolian processes may lead to significant particle transfers  
529 from Patagonia into the central Atlantic sector of the Southern Ocean (Hegner et al., 2007;  
530 Desiage et al., 2018). The aeolian contribution of loess from Patagonia was enhanced during  
531 the LGM as far as Antarctica (Petit et al., 1990; Grousset et al., 1992; Diekmann, 2007;  
532 Lambert et al., 2008). Although this hypothesis could be supported by the radiogenic Nd  
533 signal observed in both the clay-size and the sortable silt mode fraction during the LGM, the  
534 mineralogical of MD07-3076Q sediments is not in agreement with an enhanced contribution

535 from Patagonia. The dominant clay mineral in Patagonia is smectite (up to 95%), and the  
536 contribution of Patagonian smectite in our clay-size fraction is lowest during the LGM ([figure](#)  
537 [3](#)). Consequently, direct aeolian inputs cannot be significant at our site, and oceanic currents  
538 appear to be the main contributors to sedimentation ([Moriarty, 1977](#); [Petschick et al., 1996](#);  
539 [Kuhn and Diekmann, 2002](#)). This means that any aeolian particles from Patagonia delivered  
540 to the southwest Atlantic ([Mathias et al., 2014](#)) likely reach our site via oceanic transport  
541 processes.

542 The size of particles transported by water flow is directly related to the flow energy,  
543 with size increasing with the flow speed. This property was used in previous studies to track  
544 the temporal evolution of paleo-current flow (e.g. [McCave and Hall, 2006](#); [McCave and](#)  
545 [Andrews, 2019](#)). Below we show that the different grain-size modes not only correspond to  
546 distinct flow speeds, but specific water mass at site MD07-3076Q. Consequently, grain-size  
547 modes (2-20  $\mu\text{m}$  and 20-32/40  $\mu\text{m}$ ) can be associated to specific deep-water-masses.

548 The clay-size fraction (i.e., the  $< 2 \mu\text{m}$  particles) with very high specific surface  
549 readily floats and hence can undergo long-range transportation ( $>500 \text{ km}$ ). Consequently,  
550 particles within this grain-size fraction can be used to record the relative contribution of the  
551 main deep-water masses.

### 552 1. Clay-size fraction: the distant terrigenous signal

553 The NADW carries fine suspended particles from as far as the Labrador Sea.  
554 However, this signal is probably overprinted by the supply of sediments from more proximal  
555 sources such as Africa or South America. The western branch of the NADW (NADW-W; i.e.,  
556 Deep Western Boundary Current) can transport sediments from Brazil, Parana River, Pampas,  
557 and Patagonia ([figure 1](#)). Both mineralogical and geochemical compositions of the particles  
558 transported by the western branch of the NADW are likely controlled by the southern  
559 extension of the water-mass. Currently the western branch drifts away from the South

560 American coast around 45°S. This southern extension likely changed during the last glacial  
561 period and deglaciation (Skinner et al., 2010, 2014; Vasquez-Riveiros et al., 2010;  
562 Waelbroeck et al., 2011; Gottschalk et al., 2015a,b; Howe et al., 2016). Depending on its  
563 southern extent, the NADW is thus able to carry either mostly kaolinite (equatorial Brazil), or  
564 kaolinite and smectite (southern Brazil), or illite (Parana River basin), or well-crystallized  
565 smectite (crystallinity ~1.5; Patagonia). As shown above, the  $\epsilon_{Nd}$  signature of these sediments  
566 is more radiogenic when the NADW extends further south (-20 in equatorial Brazil, to ~0 in  
567 Patagonia) while the Pb and Sr yield a slightly less radiogenic signature (figures 6 and 7).

568 The CDW transports sediments from the Scotia Sea area, including Scotia Arc,  
569 Patagonia, and the Antarctic Peninsula, as well as from the Weddell Sea when mixing occurs  
570 between the lower part of the LCDW and AABW. Then particles transported by the CDW  
571 likely record a younger isotopic signature (i.e., more radiogenic  $\epsilon_{Nd}$  and unradiogenic Sr and  
572 Pb isotopes), and a mineralogical composition dominated by smectite and chlorite, while  
573 additional contribution of old material (unradiogenic  $\epsilon_{Nd}$  and radiogenic Sr and Pb isotopes)  
574 and illite is possible when the LCDW and AABW mix together. In summary, illite is  
575 considered as injected into the sedimentary system via the AABW through mixing processes  
576 between LCDW and AABW while smectite and chlorite are likely transported by the CDW.

577 The AABW in the Atlantic sector mainly forms in the Weddell Sea embayment (Orsi  
578 et al., 1999), and thus, must carries sediments from its drainage basin (East Antarctica,  
579 Antarctic Peninsula), but may also transport sediment winnowed on its pathway such as the  
580 Scotia Arc or southern Patagonia in the Argentine Basin. The main clay minerals in the  
581 Weddell Sea embayment are illite associated with chlorite. Smectite may be injected in the  
582 AABW at the vicinity of the Scotia Arc. Kaolinite is almost absent in both the Weddell Sea  
583 embayment and the Scotia Sea area (Patagonia, Antarctic Peninsula and Scotia Arc) and is  
584 unlikely to be transported by the AABW. The isotopic composition of fine particles

585 transported by the AABW ranges between two end-members: the very old sediments from  
586 East Antarctica ( $^{87}\text{Sr}/^{86}\text{Sr} > 0.72$ ;  $\epsilon_{\text{Nd}} < -20$ ), and younger material from the Antarctic  
587 Peninsula and the Scotia Arc ( $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.70$ ;  $\epsilon_{\text{Nd}} > -5$ ; intermediate Pb isotope ratios).  
588 Consequently, the composition of particles transported by the AABW flowing out of the  
589 Weddell Sea is intermediate, with  $\epsilon_{\text{Nd}}$  around -9 (Walter et al., 2000), and is potentially  
590 enriched in young material when entering the Scotia Sea.

591 The above observations are summarized in table 5, in which specific isotopic and  
592 mineralogical signatures are associated to each deep water mass. It is clear that the K/C ratio  
593 as well as the  $\epsilon_{\text{Nd}}$  composition reflect northern- versus southern-born water mass contributions  
594 (i.e., NADW versus CDW and AABW). Similarly, smectite and illite are transported by  
595 distinct water-masses and the illite/smectite ratio (I/S) can efficiently distinguish the AABW  
596 from the other water-masses. Thus, this ratio will further be used as a tool for estimating the  
597 AABW contribution through time.

598 The illite/smectite and kaolinite/smectite ratios have been plotted in figure 8. The three  
599 end-members for AABW, CDW and NADW are plotted according to the clay composition of  
600 their suspended loads. Figure 8 highlights how the relative contribution of the three water  
601 masses (i.e., AABW, CDW, and NADW) to clay deposition has changed with time since the  
602 LGM. To interpret this figure it is important to keep in mind that changes observed in the  
603 contribution one deep water mass are relative compared to the other water masses. To  
604 construct figure 8.b, the proportion of clay particles deposited by each deep water mass was  
605 calculated as described below.

606 End-member deep water mass compositions were calculated from the average  
607 composition of deep water-masses shown in figure 8.a. Weights are computed as the inverse  
608 of the distance between the position of water-mass end-member ( $x_w; y_w$ ) with  $x_w = \text{K}/\text{C}_w$  and

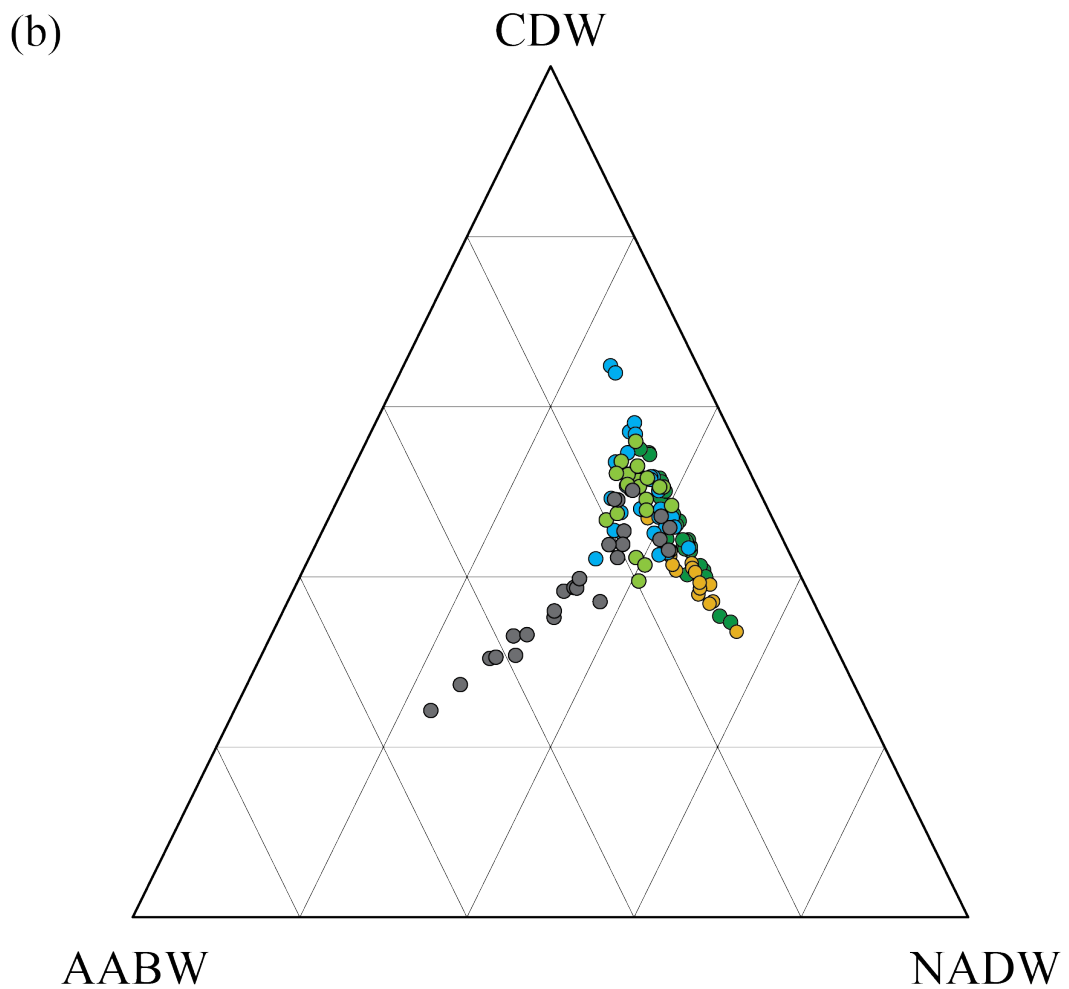
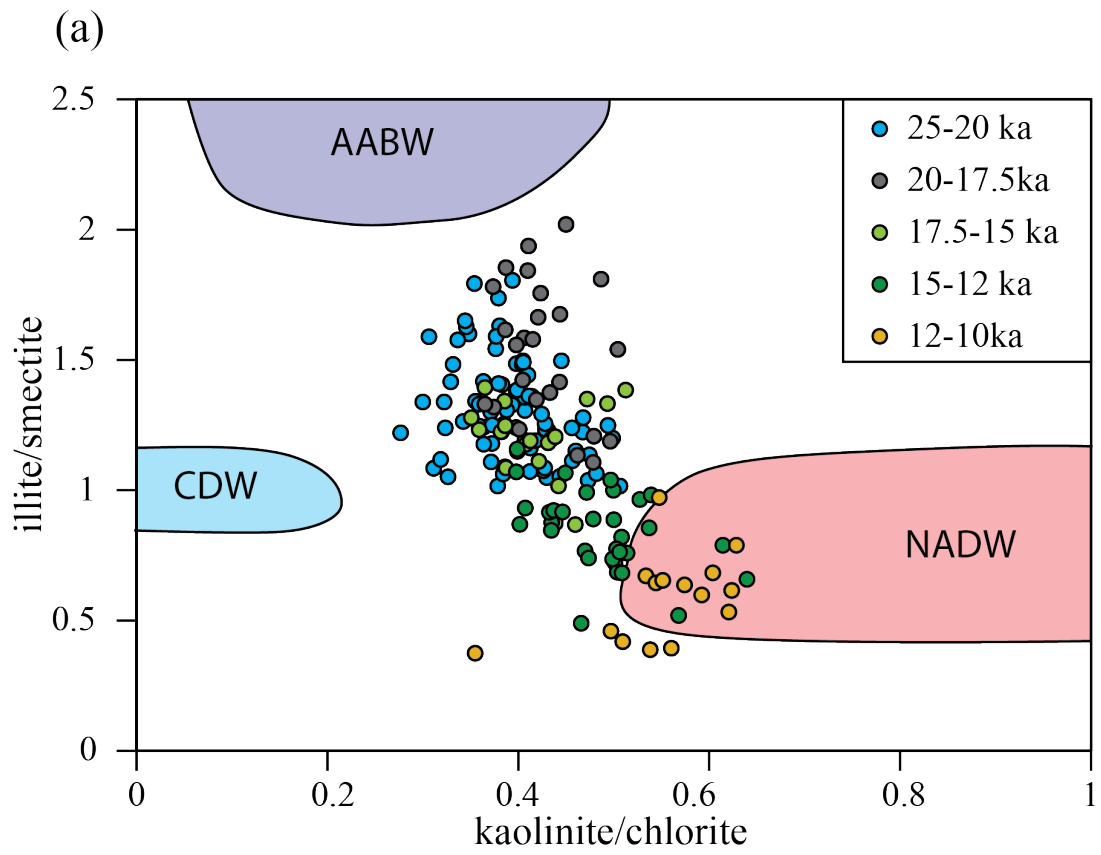


609  $y_w=I/S_w$  and the position of the data point  $(x_d;y_d)$ , with  $x_d=K/C_d$  and  $y_d=I/S_d$  taken from the  
610 graph ([figure 8.a](#)) for each water-mass:  $W_w=1/[ (x_d-x_w)^2+(y_d-y_w)^2]^{1/2}$ . The position used for  
611 the deep water masses are:  $x_{AABW} = K/C_{AABW} = 0.3$  ;  $y_{AABW} = I/S_{AABW} = 2.5$  ;  $x_{CDW} = K/C_{CDW}$   
612  $= 0.1$  ;  $y_{CDW} = I/S_{CDW} = 1$  ;  $x_{NADW} = K/C_{NADW} = 0.9$  ;  $y_{NADW} = I/S_{NADW} = 1$ . The values used  
613 for the NADW correspond to the mineralogical composition of the NADW during the LGM.  
614 As a result, the contribution of NADW to the deposition of clay particles is underestimated  
615 (~10%) at the end of the deglaciation while the contribution of the CDW is overestimated  
616 (~10%). This choice does not have a strong impact on the contribution of the AABW (less  
617 than 4.5%). Second, the percentage of the contribution of each water mass is calculated by  
618 bringing the sum of the weights computed to 100: [NADW's contribution] =  $W_{NADW}/(W_{NADW}$   
619  $+ W_{CDW} + W_{AABW}) \times 100$

620 To test the validity of this method to reconstruct the evolution of the past deep  
621 circulation, we compared our results with those of other water-masses proxies (e.g. seawater  
622  $\epsilon_{Nd}$  and  $\delta^{13}C$ ). However, these latter methods proved incapable of distinguishing AABW from  
623 CDW. Consequently, we based the comparison on the proportion of the NADW compared to  
624 the southern sourced deep waters. Based on data from the same core, our estimates give  
625 comparable results to these two proxies (see [table 6](#)). This suggests that our approach  
626 successfully reconstructs the evolution of these three deep water masses. However, our  
627 estimate of the proportion of the NADW is higher than the estimate of [Howe et al. \(2016\)](#)  
628 based on  $\epsilon_{Nd}$ . This discrepancy can be explained by two factors. First, we were unable to  
629 make corrections for concentration differences of particles in the three deep water masses  
630 (i.e., quantity of clay particles per volume of water) due to the large uncertainties involved. In  
631 addition, it is very likely that the NADW is more concentrated in clay particles than the  
632 AABW and CDW due to sediment source effect (e.g., dry East Antarctica below thick ice  
633 sheets vs. humid South America highly drained by equatorial rivers). Consequently, even

634 though our tool successfully reconstruct the variations of the NADW, the contribution of  
635 NADW is probably overestimated. In other words, our estimate of NADW contribution to the  
636 clay deposition is likely higher than the proportion of NADW (in volume of water). Second,  
637 the choice of the end member for the seawater  $\epsilon_{Nd}$  of AABW+CDW made by [Howe et al.](#)  
638 [\(2016\)](#) is debatable (-5.5 during the LGM, -8.5 during the Holocene). [Skinner et al. \(2013\)](#)  
639 used a compositional end-member between -5 and -2, which yields a better match with the  
640 isotopic composition of seawater Nd at our site (-5.5, [Skinner et al., 2013](#)). Therefore, the  
641 estimate of [Howe et al. \(2016\)](#) appears to underestimate the contribution of NADW at our  
642 site.

643 From 25 to 20 ka, the CDW is clearly dominant, with a contribution of 50% on  
644 average, while the NADW and AABW deliver 35% and 15% of the clay fraction,  
645 respectively. The highest contribution of the AABW is observed at the end of the LGM  
646 between 20 and 17.5 ka (27% in average, but up to 52% at 18 ka). During this interval, the  
647 NADW and CDW deliver respectively 34% and 39% on average. At the end of this period,  
648 the mineralogical composition of the fine sediments deposited at site MD07-3076Q suggests  
649 that the AABW was dominant. At the beginning of the deglaciation, between 17.5 and 15 ka  
650 (i.e., during HS 1), the clay mineral composition of core MD07-3076Q indicates a decreased  
651 relative contribution of the AABW end-member (15% on average) associated to an increase  
652 of the relative contribution of the CDW (49%) end-member. This observation of an increase  
653 of the CDW contribution is supported by the radiogenic  $\epsilon_{Nd}$  of the clay-size fraction and a  
654  $^{87}Sr/^{86}Sr$  close to the Scotia Arc composition ([figures 5 and 6](#)).



656 *Figure 8: (a) Plot of kaolinite/chlorite vs illite/smectite ratios from core MD07-3076Q*  
657 *from 25 to 10 ka. The composition of deep-water masses has been determined using the data*  
658 *from [Petschick et al. \(1996\)](#). (b) Ternary diagram comparing the contribution of the three*  
659 *deep water masses to the clay deposition at site MD07-3076Q.*

660 From 15 ka on, the clay mineral composition at the study site records an increasing  
661 contribution of the NADW until the end of the deglaciation (at least 47% for the interval 10-  
662 12 ka, [figure 8](#)). Compared with the LGM this suggests enhanced contribution of the NADW  
663 with reduced contributions from the CDW (40% for the interval 10-12 ka) and AABW (13%  
664 for the interval 10-12 ka). These results are strongly supported by the sharp decrease in  $\epsilon_{Nd}$   
665 and by the concomitant increase of  $^{208}Pb/^{204}Pb$  and  $^{207}Pb/^{204}Pb$  ratios from the clay size  
666 fraction ([figures 5 and 7](#)). This corroborates the conclusions based on the *cibicides*  $\delta^{13}C$   
667 measured in the same core ([figure 9.e, Waelbroeck et al., 2011](#)). In addition, increasing inputs  
668 of less crystallized smectite (slightly higher crystallinity:  $\sim 1.5$ , [supplementary figure S1](#)) from  
669 Patagonia support the southward penetration of the NADW in the southern Atlantic Ocean,  
670 consistent with an increasing contribution of the NADW-W to sedimentation at site MD07-  
671 3076Q. Interestingly, the transition between B/A (that starts simultaneously with the southern  
672 Hemisphere cooling, the Antarctic Cold Reversal) and the YD appears to have no effect on  
673 the relative contribution of deep water masses and shows a continuous increase of the relative  
674 contribution of the NADW during these two periods.

675 Overall, the clay size fraction demonstrates that the contribution of the three deep  
676 water-masses changed through time. An increase of contribution of the AABW is recorded at  
677 the end of the LGM and is followed by a two-step deglaciation. First, an important reduction  
678 of the AABW is observed at the HS 1 onset. Second, an increase in the NADW contribution  
679 compared to the CDW is observed during the B/A and YD.

680                   2.       Cohesive silt mode fraction (2-20  $\mu\text{m}$ ): an AABW signal

681           According to our geochemical data, the provenance of the cohesive silt fraction is  
682 dominated by sources from the Scotia Sea area, especially Patagonia and Antarctic Peninsula,  
683 and by an old, unidentified, crustal source. According to the modern deep oceanic circulation,  
684 only the CDW and AABW are able to transport sediment from the Scotia Sea to the studied  
685 site (figure 1). It was also likely the case during the LGM and deglaciation. The AABW and  
686 CDW were likely spreading further north while the NADW was likely shallower, and  
687 retreated to the north (Kuhn and Diekmann, 2002; Barker et al., 2010; Waelbroeck et al.,  
688 2011). These modifications do not affect the continental sources of terrigenous sediment for  
689 the CDW and AABW compared the present day, and the NADW is still unable to transport  
690 sediments from the Scotia Sea. Consequently, it excludes the NADW as a transport medium  
691 for the cohesive silt mode fraction to site MD07-3076Q. Therefore, because the NADW is  
692 also the only water-mass able to transport sediment from the South American Shield to our  
693 study site, it excludes the South American Shield as a source of sediment for the cohesive silt  
694 mode fraction, and implies East Antarctica as the source of old material identified in the  
695 cohesive silt mode fraction. The terrigenous supplies deriving from East Antarctica/Scotia Sea  
696 strongly suggest the AABW has been the dominant carrier of the 2-20  $\mu\text{m}$  grain-size fraction  
697 since the LGM. In addition, because the mean isotopic composition of this grain-size fraction  
698 transported by the AABW is relatively constant through time (figure 6), the relative  
699 contributions of East Antarctica and Scotia Sea area seem to have remained largely  
700 unchanged since the LGM.

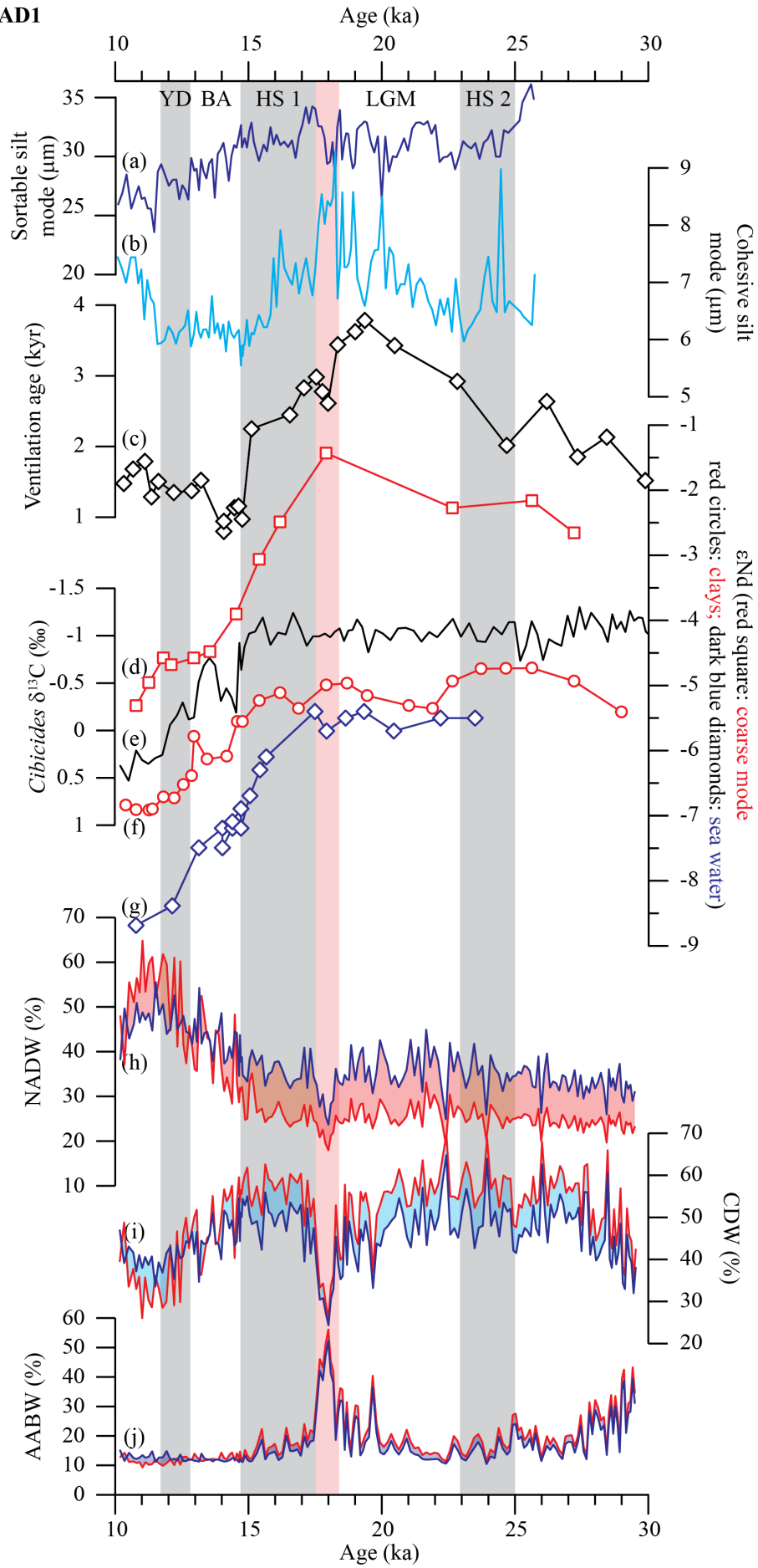
701                   3.       Sortable silt mode fraction (20-32/40  $\mu\text{m}$ ): A LCDW signal

702           A part of the particles transported by the AABW and the NADW are derived from old  
703 crustal sources (i.e., East Antarctica and South American Shield, respectively). Thence, these  
704 two deep-water masses cannot be the source of the sortable silt mode fraction during the

705 LGM. In contrast, the CDW is a good candidate to transport particles solely from the Scotia  
706 Sea, as identified for the sortable silt mode fraction during the LGM. Consequently, the only  
707 possible source of this grain size fraction is the CDW. During the deglaciation, the Nd isotope  
708 composition of the sortable silt mode fraction became gradually less radiogenic while the  
709  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio fluctuated, suggesting a modification of provenances and/or an additional  
710 contribution of less radiogenic Nd source. Such isotopic changes could be explained in two  
711 ways; (1) a change in the CDW pathway; (2) the incorporation of terrigenous material by  
712 mixing with other water-masses (such as AABW or NADW). The CDW slowed down and  
713 migrated southward during the deglaciation in concert with the southward migration of the  
714 subpolar and polar fronts (Howard and Prell, 1992; Anderson et al., 2009; Denton et al.,  
715 2010). However, such a change in the CDW pathway would have no effect on the terrigenous  
716 sources of particles. In contrast, several studies proposed increased ventilation in the Southern  
717 Ocean, enabling the degassing of oceanic  $\text{CO}_2$  to the atmosphere (Ahn and Brook, 2008;  
718 Fischer et al., 2010; Skinner et al., 2010, 2013, 2014; Siani et al., 2013). Currently the base of  
719 the LCDW is mixed with the AABW, which was apparently not the case during the LGM  
720 (Skinner et al., 2010). The onset of AABW/LCDW mixing could explain the incorporation of  
721 unradiogenic terrigenous material from the Weddell Sea embayment (i.e., East Antarctica)  
722 into the LCDW. This mechanism likely explains the shift in  $\epsilon_{\text{Nd}}$  composition during the  
723 deglaciation. This hypothesis is supported by the results of Walter et al. (2000) who reported  
724 lower  $\epsilon_{\text{Nd}}$  in terrigenous sediments in the southern Scotia Sea during the interglacial period  
725 and interpreted this as higher contribution of sediment from the Weddell Sea embayment due  
726 to the mixing of the AABW into the ACC. From 15 ka, the Nd isotope composition of the  
727 sortable silt mode fraction is very close to the one from the cohesive silt mode fraction  
728 transported by the AABW. It suggests that the sortable silt mode fraction is specifically  
729 transported by the LCDW rather than by the entire CDW. This highlights the fact that the  $\epsilon_{\text{Nd}}$

730 variations from the sortable silt mode fraction provide a useful tool to study the mixing  
731 between the AABW and LCDW.

Figure 9 AD1





733 *Figure 9: Comparison between paleoceanographic proxies from core MD07-3076Q and*  
734 *data produced in this study. Grey areas delimit the main climatic events and the red band*  
735 *highlights the particularly strong AABW event. (a) size of the sortable silt mode ( $\mu\text{m}$ ), a proxy*  
736 *of the speed of the LCDW; (b) size of the cohesive silt mode ( $\mu\text{m}$ ), a proxy of the speed of the*  
737 *AABW; (c) bottom water ventilation age from site MD07-3076Q (data from [Skinner et al.,](#)*  
738 *[2010](#)); (d)  $\epsilon_{\text{Nd}}$  from the sortable silt size fraction; (e) *Cibicides*  $\delta^{13}\text{C}$  from site MD07-3076Q  
739 *([Waelbroeck et al., 2011](#)); (f)  $\epsilon_{\text{Nd}}$  from the cohesive silt size fraction; (g) seawater  $\epsilon_{\text{Nd}}$  from*  
740 *site MD07-3076Q ([Skinner et al., 2013](#)); (h) proportion of clay particles delivered by the*  
741 *NADW (see [figure 8.b](#)); (i) proportion of clay particles delivered by the CDW (see [figure 8.b](#));*  
742 *(j) proportion of clay particles delivered by the AABW (see [figure 8.b](#)). On curves (h), (i) and*  
743 *(j), the red line represent values calculated using the Holocene end-member of the NADW ( $I/S$*   
744 *= 0.5), while the blue line correspond to the values calculated using the LGM end-member of*  
745 *the NADW ( $I/S = 1$ ).**

#### 746 4. Paleoceanographic implication

747 Our results establish that different grain-size modes can be attributed to distinct water-  
748 masses and that variation of the particle size of the different grain-size modes ([figure 2](#)) can  
749 be used to record the variations of flow speed of these water-masses as well as changes in  
750 pathways/mixing. This method has the advantage of distinguishing the speed of each water  
751 mass instead of giving the average speed of all the water masses, as it is the case for the  
752 sortable silt method ([McCave and Hall, 2006](#)). During the last glacial period, the large mean  
753 particle diameter ( $>30 \mu\text{m}$ ) of the sortable silt mode fraction reflects strong LCDW activity.  
754 At the beginning of the deglaciation, the variation of the mean particle size suggests that the  
755 LCDW speed slightly decreased, but remained high. Since  $\sim 15$  ka, the LCDW speed has been  
756 showing a gradual decrease. [McCave et al. \(2014\)](#) observed minor change in the Antarctic  
757 Circumpolar Current (which includes the CDW) flow speed at the Drake Passage during the

758 deglaciation. Our observation, however, is consistent with other studies in the south Indian  
759 sector of the Southern Ocean indicating that the Antarctic Circumpolar Current (which  
760 includes the CDW) flow speed was maximum during the LGM and decreased during the  
761 deglaciation (Mazaud et al., 2010), with a notable increase from ~15 ka on (Mazaud et al., in  
762 prep.). Altogether, these studies and our results suggest that the weakening of the ACC started  
763 at ~15 ka.

764 From ~26 to 21 ka, the low particle sizes recorded in the cohesive silt mode (figure  
765 2.b, transported by the AABW) indicate slow AABW, which is consistent with the observed  
766 moderate relative contribution of AABW in clay particles (figure 8 and 9). In this time  
767 interval, one isolated high value is recorded at 24 ka and suggests that, temporarily, AABW  
768 activity was stronger. However, no change is registered in the clay minerals (figure 3 and 9).  
769 The grain-size data indicate that the AABW became more vigorous from 21 ka to reach its  
770 strongest speed at the end of the LGM. This maximum speed is concomitant with the  
771 strongest relative contribution of the AABW of the clay record (figures 3.h and 9), as shown  
772 by the peak of the illite/smectite ratio (figure 3) illustrated on figure 8. Interestingly, this  
773 AABW maximum preceded the HS 1, and both its speed and relative contribution fell at the  
774 LGM/HS 1 transition as defined by Missiaen et al. (2019). At the beginning of the  
775 deglaciation, the abrupt decrease of the fine grain-size mode indicates a slowdown of the  
776 AABW (figure 2). This is supported by a sharp decrease in the illite/smectite ratio in the clay-  
777 size fraction (figure 3), which also suggests an abrupt decrease of the relative contribution of  
778 the AABW. At ~16 ka, a slowdown of the AABW is indicated by a second sharp decrease of  
779 the cohesive silt mode (figure 2), which reaches its lowest values of the entire record,  
780 indicating low AABW speed from 16 to 11.5 ka, when the AABW activity increased  
781 suddenly and recovered the speed observed before 16 ka.

782           These variations of the AABW speed could indicate the transition from the Ice Shelf  
783 Mode dominated bottom water formation at the end of the LGM to the Polynyas Mode  
784 dominated bottom water formation during interglacial periods. The variations of the cohesive  
785 silt mode support a strong AABW during the LGM and the early Holocene, and a minimum  
786 flow during the deglaciation. This minimum flow would result from the reduction of bottom  
787 water production via the Ice Shelf Mode and still low production via the Polynyas Mode. Our  
788 results confirm the results of [Krueger et al. \(2012\)](#) who demonstrated the relation between the  
789 flow speed changes in the Cape Basin and the higher AABW production rates and indicate  
790 that the size of the cohesive silt mode is a reliable tracer of the AABW flow speed and hence  
791 of AABW production rates in the Weddell Sea.

792           Overall, the variations of the AABW flow speed and ventilation age are similar, with  
793 higher average speeds corresponding to the old ventilation ages. Taken together with the  
794 reconstructed vigorous LCDW during the LGM ([figure 9.a.](#)), these observations suggest that  
795 the deep Southern Ocean was poorly ventilated but with a vigorous circulation. It means that  
796 both the AABW and LCDW remained isolated from the surface during the LGM. This  
797 conclusion is consistent with our hypothesis that AABW was formed under ice shelves during  
798 this period and that the major expansion of sea ice ([Bianchi and Gersonde, 2004; Bostock et](#)  
799 [al., 2012; Adkins 2013](#)) kept the AABW isolated from the atmosphere during the LGM. In  
800 addition, poorly ventilated deep Southern Ocean associated with a vigorous deep circulation  
801 (i.e., high AABW and LCDW flow) suggests that a barrier existed between the deep ocean  
802 and the surface, which is consistent with the study of [Howe et al. \(2016\)](#). In contrast, direct  
803 comparison highlights the fact that the peaks of ventilation age and of the AABW flow are not  
804 in phase. Indeed, the peaks of the AABW flow ([figure 9.b](#)) and of the AABW contribution to  
805 clay deposition ([figure 9.j](#)) are synchronous with the first decrease in deep water ventilation  
806 ages after the LGM and prior to HS1. Moreover, this well-defined change in the AABW

807 behavior is recorded by two distinct proxies (i.e., clay minerals and grain size distribution)  
808 and in two distinct grain size fractions (clays and silts). These synchronous changes are  
809 associated with a drop in ventilation age prior to the HS 1. The variations in two grain size  
810 fractions that are transported and deposited by distinct dynamic processes indicates a major  
811 perturbation in Southern Ocean circulation. Interestingly, the increase in the AABW strength  
812 associated to the early decrease in deep water ventilation age took place significantly earlier  
813 than the changes recorded by the other proxies measured in core MD07-3076Q (i.e., sea water  
814  $\epsilon_{Nd}$  between 17.5 and 16 ka,  $\delta^{13}C$  at 14.5 ka,  $\delta^{18}O$  at 15 ka, LCDW flow speed at 15 ka,  
815 NADW proportion at 15 ka) and also precedes the atmospheric CO<sub>2</sub> rise (Bereiter et al., 2015)  
816 and the change in the opal flux (Anderson et al., 2009). For the first time, these data indicate a  
817 change in the deep Southern Ocean circulation that is associated with ventilation change that  
818 precedes the HS 1. This suggest that the identified large temporary (~1 kyr long) increase in  
819 the AABW contribution and flow speed may have had a substantial impact on the deep  
820 Southern Ocean ventilation, and by extension, on the deglacial CO<sub>2</sub> rise. Indeed, the  
821 magnitude of the change in the AABW behavior has the potential to “break” the hypothetical  
822 physical barrier between the deep Southern Ocean and the surface that was responsible of  
823 poorly ventilated deep water during the LGM. This combination of events also explains the  
824 unexpected early fall of the deep water ventilation age observed at ~18.2 ka.

## 825 Conclusion

826 The mineralogical characteristics of the clay-size fraction provide new insights into  
827 the general thermohaline circulation in the southern South Atlantic. The mineralogical signal  
828 demonstrates strong influence of both the CDW and AABW during the last glacial period.  
829 Clay assemblages reveal a particularly high contribution of the AABW at the end of the  
830 LGM, which is interpreted as reflecting strong Ice Shelf Mode bottom water production in the  
831 Weddell Sea. The contribution of the CDW and AABW then progressively decreases during

832 the deglaciation together with the growing influence of the NADW (figure 8). Indeed, clay  
833 size fraction data monitor the progressive incursion of NADW into the Southern Ocean  
834 during the deglaciation from ~15 ka (figure 8), as well as the synchronous southward retreat  
835 of the southern-born deep water-masses (i.e., CDW and AABW).

836 The isotopic compositions of the distinct grain size modes provides invaluable  
837 information. The sortable silt mode fraction helps to trace the LCDW activity whereas the  
838 cohesive silt mode fraction is associated with the AABW signal. Consequently, the temporal  
839 evolution of flow speeds, pathways and interactions of these deep water-masses can be  
840 determined in a very robust way. This study suggests: (1) vigorous LCDW during the LGM  
841 and a marked weakening from ~15 ka; (2) strong AABW at the end of the LGM and during  
842 the early Holocene, and weak AABW during the deglaciation between 16 and 11.5 ka. These  
843 results suggest a decrease of AABW production rates during the deglaciation that could result  
844 from the transition from strong Ice Shelf Mode bottom water production during the LGM to  
845 stronger Polynyas Mode bottom water during the Holocene.

846 The strong LCDW speed observed during the LGM is associated with increased clay  
847 deposition by the CDW, and the reduction of LCDW speed since the B/A is accompanied by  
848 an increased clay contribution from the NADW. Similarly, the highest AABW speed recorded  
849 at the end of the LGM is concomitant with the highest relative contribution of the AABW to  
850 clay mineral deposition. Interestingly, this AABW maximum preceded the HS 1 and both its  
851 speed and contribution fell markedly at the LGM/HS 1 transition.

852 Finally, this study demonstrates that based on (1) the combination of isotopic and  
853 mineral investigation of the clay size fraction, (2) the combination of isotopic measurement of  
854 isolated grain size modes and of grain size distribution results, and (3) a complete synthesis of  
855 the potential sources of sediments, it is possible to obtain reliable information on the  
856 lithogenic material transported by each deep water mass. This approach allows efficient

857 tracking of the evolution of distinct water-masses through time, since it helps to isolate the  
858 specific signal associated to each water-mass. Based on these observations, we conclude that  
859 the increase in the AABW strength at the end of the LGM may have had a substantial impact  
860 on the ventilation recovery of the deep Southern Ocean, and by extension, to the deglacial  
861 CO<sub>2</sub> rise.

## 862 Acknowledgments

863 We wish to thank Richard Smeets for his invaluable help in the clean lab, Emmeke  
864 Bos, Janne Koorneef, and Mathijs van de Ven for the help with TIMS Triton measurements,  
865 Bas van der Wagt and Kirsten van Zuilen for Neptune MC-ICPMS measurements. Tristan  
866 Lippens is thanked for computer support. We also wish to thank all organizations that  
867 provided financial support to the project: the “region Haut de France” and the Vrije  
868 Universiteit of Amsterdam. The research was supported by Europlanet 2020 RI, which  
869 received funding from the European Union's Horizon 2020 research and innovation  
870 programme under grant agreement No 654208. We also want to express our special thanks to  
871 the “flotte océanographique française” for retrieving the sediment core used in this study. FB  
872 and VBR thanks Monique Gentric for administrative support. CW acknowledges support  
873 from the European Research Council ERC grant ACCLIMATE/n° 339108.

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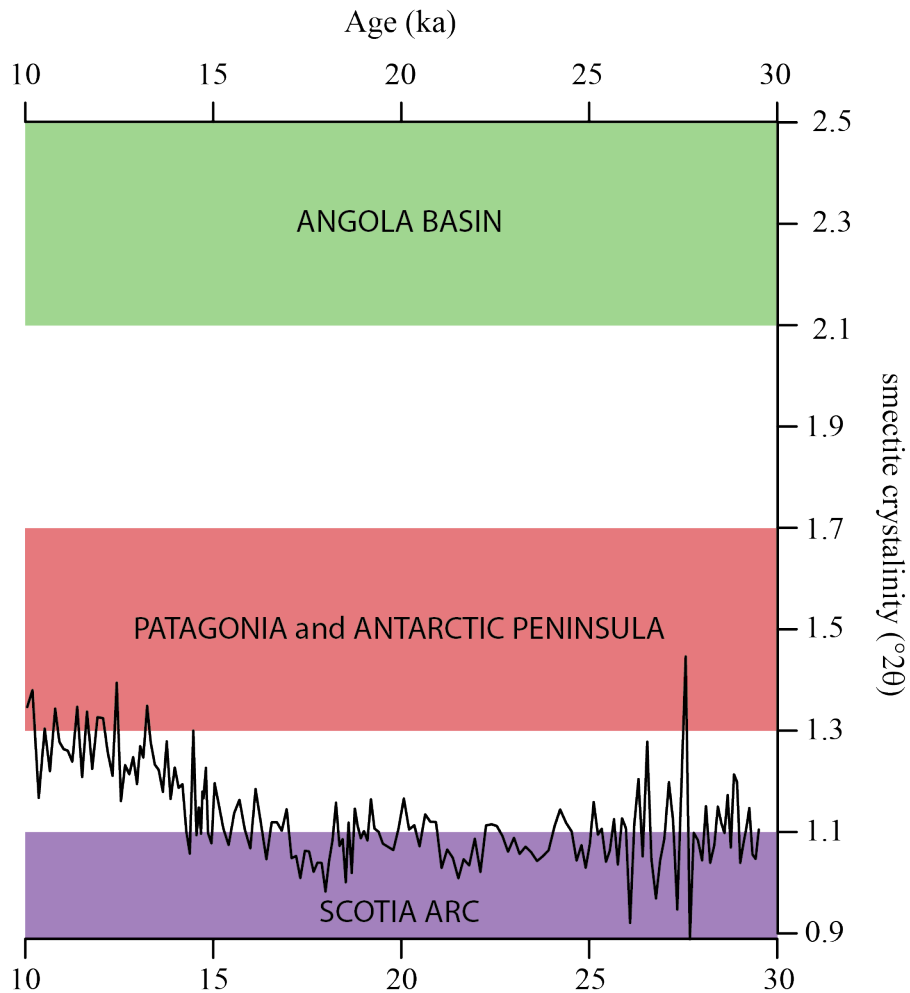
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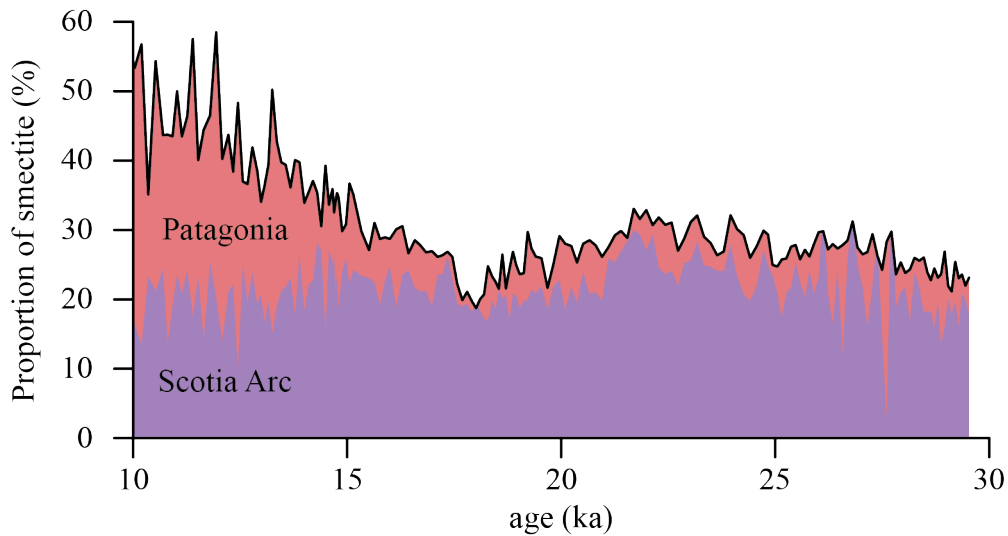
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1369 **Supplementary figures**



1370  
1371 SUPPLEMENTARY FIGURE S1: Crystallinity of smectite from site MD07-3076Q  
1372 (black curve) and comparison with crystallinity of smectite from potential sources to the  
1373 South Atlantic basin. The data used for comparison are from [Petschick et al., 1996](#).



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SUPPLEMENTARY FIGURE S2: Contribution of the two types of smectite from site

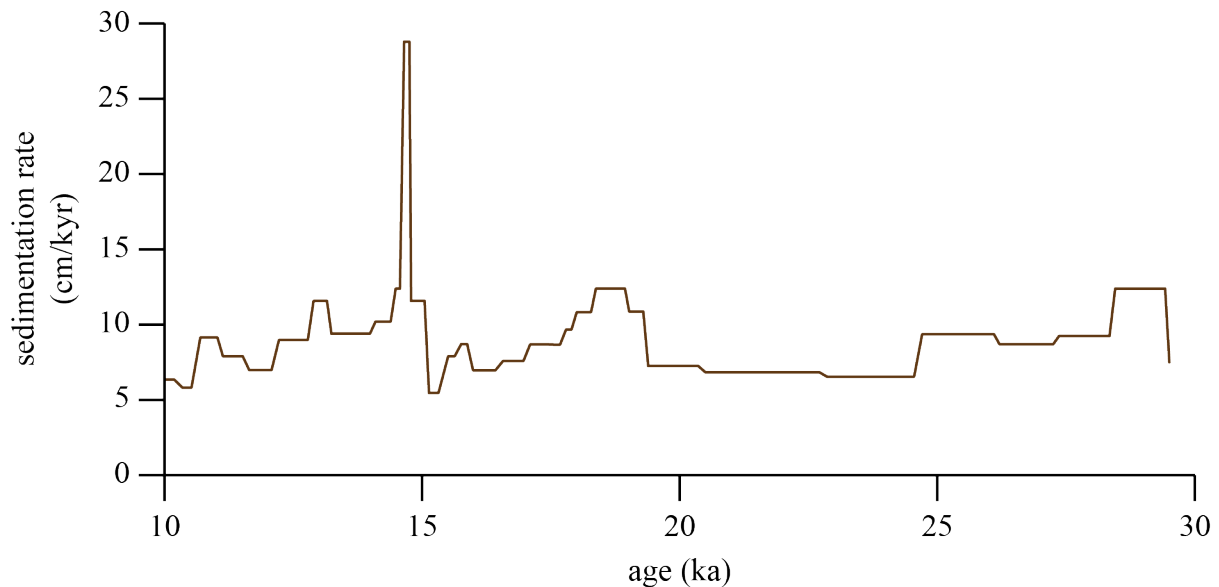
MD07-3076Q. The black line represent the total smectite content, the purple area represent

the smectite content from the Scotia Arc, and the red area represent the smectite content from

Patagonia. The proportion of the two types of smectite are calculated as followed:

$$[\% \text{Patagonian smectite}] = [\% \text{total smectite}] \times \frac{(\text{sample crystallinity}) - (\text{Scotia Sea crystallinity})}{(\text{Patagonia crystallinity}) - (\text{Scotia Sea crystallinity})} \times 100$$

$$[\% \text{Scotia Sea smectite}] = [\% \text{total smectite}] - (\% \text{Patagonian smectite})$$



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SUPPLEMENTARY FIGURE S3: Sedimentation rate from site MD07-3076Q

calculated from the age model published in [Skinner et al. \(2010, 2013, 2014\)](#) and [Gottschalk](#)

[et al. \(2015b\)](#).

	average (10-30 ka)	LGM (25-17 ka)	Deglaciation (17-11.7 ka)	standard deviation (10-30 ka)
fine mode ( $\mu\text{m}$ )	6.8	7.2	6.3	0.8
coarse mode ( $\mu\text{m}$ )	30.4	31.3	30.4	2.5
clay fraction (%)	3.4	3.0	3.8	0.7
sortable silts (10-63 $\mu\text{m}$ ; %)	47.0	47.2	46.2	2.4
cohesive silts (2-10 $\mu\text{m}$ ; %)	39.4	38.8	40.2	2.4
sand (%)	3.0	3.5	2.6	1.0
smectite (%)	30.7	26.7	35.7	7.9
illite (%)	33.5	34.7	31.7	3.3
chlorite (%)	25.1	27.2	22.3	4.2
kaolinite (%)	10.6	11.4	10.3	1.3
I/S	1.2	1.3	0.9	0.3
K/C	0.4	0.4	0.5	0.1
$\epsilon_{\text{Nd}}$ (<2 $\mu\text{m}$ )	-5.8	-5.0	-6.0	0.8
$\epsilon_{\text{Nd}}$ (fine mode)	-4.6	-4.6	-4.7	0.4
$\epsilon_{\text{Nd}}$ (coarse mode)	-3.3	-1.8	-3.7	1.2
$^{87}\text{Sr}/^{86}\text{Sr}$ (<2 $\mu\text{m}$ )	0.71130	0.71144	0.71136	0.00036
$^{87}\text{Sr}/^{86}\text{Sr}$ (fine mode)	0.70795	0.70800	0.70798	0.00019
$^{87}\text{Sr}/^{86}\text{Sr}$ (coarse mode)	0.70889	0.70859	0.70926	0.00058
$^{208}\text{Pb}/^{204}\text{Pb}$ (<2 $\mu\text{m}$ )	38.774	38.775	38.785	0.012
$^{207}\text{Pb}/^{204}\text{Pb}$ (<2 $\mu\text{m}$ )	15.644	15.639	15.645	0.008
$^{206}\text{Pb}/^{204}\text{Pb}$ (<2 $\mu\text{m}$ )	18.752	18.749	18.758	0.030

1384

1385 TABLE 1: Average values of the different proxies for the entire record, the LGM, and

1386 the deglaciation.