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Drake Passage Gateway opening and Antarctic Circumpolar Current onset 31 Ma ago: the message of foraminifera and reconsideration of the Neodymium isotope record

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Abstract

The establishment of the Antarctic Circumpolar Current (ACC) is one of the most important events for both global oceanic circulation and climate of the Cenozoic. The onset of this major current hinges on the opening of two major oceanic passages, the Drake Passage and the Tasmanian gateways that connect Pacific, Atlantic and Indian oceans allowing a modern-like thermohaline circulation. For decades, the ACC onset has been considered as the trigger of the Oligocene glaciation at 33.7 Ma, which marks the beginning of the modern icehouse climate. Today, this scenario is debated. The main obstacle to evaluate the ACC influence on the Oligocene glaciation remains the ill-constrained timing of the Drake Passage gateway opening. Here, we analyse the geochemical composition and Sr isotope ratio of dated planktonic and benthic foraminifera from two IODP and ODP legs in the Southern Atlantic and Pacific oceans (SAO and PO, respectively) to assess the variability of seawater masses' chemical composition through time and to better constrain the timing of the Drake Passage gateway opening along the Eocene-Oligocene interval. These results, based on seawater paleo temperature (Mg/Ca molar ratios), redox (Ce/Ce* anomaly) and provenance (⁸⁷Sr/⁸⁶Sr) proxies, highlight a gradual seawater mass mixing between the SAO and PO from 31 Ma to 26 Ma. Combined with a reconsideration of the fossil fish teeth Neodymium isotope records, these geochemical tracers evidencing the SAO-PO interconnection depicts the Drake Passage gateway opening and deepening during this 31-26 Ma interval and thus, the timing of the ACC onset. Hence, antecedence of the Oligocene glaciation onset (at 33.7 Ma) relative to the ACC onset (31-26 Ma) implies that the ACC did not trigger the Oligocene glaciation and that the role of atmospheric *p*CO₂ should be further considered.

Key words: Drake Passage gateway, Oligocene glaciation, Antarctic Circumpolar Current, Climate, Paleoceanography, Foraminifera

1. Introduction

The inception of the modern icehouse climate goes back to a long-term Eocene global cooling, which ultimately led to the onset of the Oligocene glaciation at 33.7 Ma (Zachos et al., 2001). This pivotal climatic event of the Eocene-Oligocene transition (EOT) is recorded worldwide in the foraminifera geochemistry by a major shift toward higher $\delta^{18}\text{O}$ values ($\sim +1.5$ ‰), attesting to an abrupt seawater temperature decrease and ice-sheets formation (Coxall et al., 2005; Zachos et al., 2001). It is

43 nowadays consensual that the Eocene cooling resulted from a progressive atmospheric $p\text{CO}_2$
44 decrease, which in its turn lowered the global average oceanic temperature (DeConto and Pollard,
45 2003; Pagani et al., 2011; Ladant et al., 2014). Yet, the causes of the Oligocene glaciation itself are
46 still debated. Two dominant theories have been suggested: the $p\text{CO}_2$ hypothesis and the ACC
47 hypothesis. The CO_2 hypothesis states that the glaciation was triggered by an atmospheric $p\text{CO}_2$
48 decrease below a critical threshold (~ 750 ppm, DeConto et al., 2008; Ladant et al., 2014; Galeotti, et
49 al., 2016), potentially combined with an orbital configuration that reduced polar seasonality (e.g.
50 DeConto and Pollard, 2003; Coxall et al., 2005; Pagani et al., 2011; Ladant et al., 2014). The ACC
51 hypothesis argues that the tectonic separation of Australia and South America from Antarctica led to
52 the onset of the ACC. The ACC reduced heat transport to southern high latitudes by isolating
53 Antarctica from the global oceanic circulation, thereby inducing the glaciation (Kennett, 1977;
54 Kennett and Exon, 2004). Indeed, recent circulation model experiments have been able to generate
55 strong southern hemisphere cooling in response to ACC onset (e.g. Sauermilch et al., 2019).

56 After decades of debate, the $p\text{CO}_2$ hypothesis is nowadays more and more admitted, notably since
57 numerical climate and ocean circulation models have been able to simulate glaciation without
58 involving the ACC onset (e.g. Ladant et al., 2014; Kennedy-Asser et al., 2020). However, the role of
59 the ACC in the Eocene-Oligocene climate evolution remains largely misunderstood, especially
60 because the timing of its onset is not well constrained. The onset of the ACC is driven by the opening
61 of two major oceanic apertures around Antarctica, i.e. the Tasmanian and the Drake Passage gateways
62 (Fig. 1). These gateways allowed for the ACC to form and to connect the Pacific, Atlantic and Indian
63 oceans, leading to a modern-like thermohaline circulation (Stow, 2001; Stickley et al., 2004; Scher
64 and Martin, 2006; Lagabrielle et al., 2009; Katz et al., 2011; Bijl et al., 2013; Scher et al., 2015;
65 Sarkar et al., 2019). Stickley et al. (2004) showed that the Tasmanian gateway opened to deep oceanic
66 circulation between 35.5 and 33.5 Ma. However, the timing of the Drake Passage gateway opening
67 remains poorly constrained (Fig. 1), hindering any evaluation of the ACC influence on the Eocene-
68 Oligocene climate (Scher, 2017).

69 Although numerous studies have focused on the Drake Passage gateway aperture, its opening is still
70 a matter of debate. While some authors attest of a shallow corridor from 50 Ma (Livermore et al.,
71 2005, 2007), most tectonic and sedimentology studies argue for its opening to a deep eastward
72 circulation during the Oligocene (e.g. Latimer and Fillipelli, 2002; Lawver and Gahagan, 2003;
73 Livermore et al., 2005, 2007; Brown et al., 2006; Eagles et al., 2006; Lagabrielle et al., 2009; Eagles
74 and Jokat, 2014) (Fig. 1). Only a few studies using seawater geochemical tracers proposed an age for
75 the Drake Passage gateway opening. Based on the Nd isotopes record of fossil fish teeth at Maud
76 Rise (ODP Leg 113, site 689) and Agulhas Ridge (ODP Leg 177, site 1090), (see Fig. 1 for sites'
77 location), Scher and Martin (2004, 2006) proposed a connection between the Pacific and Southern
78 Atlantic oceans at *ca.* 41 Ma, evidenced by a gradual increase toward more radiogenic ϵ_{Nd} values at
79 that time. Recently, Wright et al. (2018) showed a rather stable ϵ_{Nd} signal in fossil fish teeth from the
80 Kerguelen Plateau (ODP Leg 119, sites 738 and 744, Fig. 1) between 33 and 23 Ma. This stability
81 may reflect the absence of major paleoceanographic changes in the Southern Ocean during this
82 interval, comforting the idea that the Drake Passage gateway was already open. However, the ϵ_{Nd}
83 signature of bottom seawater recorded by fossil fish teeth can be strongly influenced by continental
84 weathering fluxes (e.g. Lacan and Jeandel, 2005; Jeandel et al., 2007; Scher et al., 2011, 2014). This
85 is particularly true for the Southern Ocean drilling sites, which are located near the Antarctica margin
86 and thus potentially exposed to highly non-radiogenic Nd fluxes from the Proterozoic to Precambrian
87 Antarctica's basement (e.g. Scher et al., 2011, 2014). Alternatively, Egan et al. (2013) showed a

88 decline in sponge $\delta^{30}\text{Si}$ values at 31.5 Ma from the Agulhas Ridge (site 1090, Fig. 1) that they
89 interpreted as signing the final establishment of a deep-water circulation through the Tasmanian and
90 Drake Passage gateways. Hence, if the Drake Passage gateway opened to deep circulation after the
91 Tasmanian gateway, which opened *ca.* 35.5-33.5 Ma (Stickley et al., 2004), then its opening should
92 have conditioned the ACC onset, potentially combined with westerlies alignment with the already
93 opened Tasmanian gateway (Scher et al., 2015). Better constraining the timing of the Drake Passage
94 gateway opening is thus fundamental to apprehend the ACC onset and to test its influence on the
95 global climate (e.g. Scher, 2017).

96 Here, we establish robust temporal constraints on the Drake Passage gateway opening and the ACC
97 onset by tracking Pacific/Atlantic seawater masses mixing over the Eocene-Oligocene interval. We
98 selected benthic and planktonic foraminifera tests from the IODP Exp 320 (Hole 1334C) in the Pacific
99 Ocean (PO) and the ODP Leg 113 (Hole 689B) in the Southern Atlantic Ocean (SAO) on which we
100 performed major and trace element chemistry and Sr isotope ratio analysis (Fig. 1). The Hole 1334C
101 has been chosen as a high quality oligotrophic Pacific Ocean endmember while the site 689B is ideally
102 located in vicinity to the Drake Passage gateway outlet, thus allowing us to track PO inputs into the
103 SAO following the gateway opening. Foraminifera use bioavailable elements of their living
104 environment to build their calcitic test. For this reason, they are one of the most used pristine
105 biomineral archive to assess past ocean chemistry since the pioneering works of Shackleton and
106 Kennett (1975) and Palmer (1985). In this present work, we first verify that the geochemical features
107 of the selected samples (e.g. major and trace element concentration and PAAS-normalised REE
108 patterns) are in line with well-preserved seawater characteristics. Then, we use the Mg/Ca paleo-
109 thermometer (Barker et al., 2005), the Ce anomaly [$\text{Ce}/\text{Ce}^* = \text{Ce}_N / (\text{Pr}_N \times (\text{Pr}_N/\text{Nd}_N))$] paleo-redox
110 proxy (e.g. Bolhar et al., 2014; Kamber et al., 2014) and the $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratio (source tracer) to
111 track temporal variations of seawater chemistry in Pacific and Southern Atlantic oceans during the
112 EOT. By investigating these tracers in both planktonic and benthic foraminifera species, we aim to
113 reconstruct both surface and bottom seawater chemical signatures in both PO and SAO, to decipher
114 when water masses mixing between the PO and SAO occurred. Finally, we confront our results with
115 the fossil fish teeth Nd isotope records available in the literature for the Pacific and the Southern
116 oceans, which are commonly used in paleoceanographic studies. Altogether these investigations
117 allow us to propose a more accurate timing for the Drake Passage gateway opening.

118

119 2. Methods

120 2.1. Sediments and foraminifera treatment

121 Based on the detailed core descriptions (Barker et al., 1988; Pälike et al., 2010), we selected 12
122 sediment fractions of the IODP Exp. 320 (Hole 1334C) in the PO and 10 from the ODP Leg 113
123 (Hole 689 B) in the SAO. A fraction of each sample was resuspended into Milli-Q water and filtrated
124 back to back at 200 μm and 100 μm . Largest fractions ($>200 \mu\text{m}$ and 200-100 μm if needed) were
125 dried in a stove at 50 $^\circ\text{C}$. Foraminifera were hand-picked under a binocular and separated according
126 to planktonic and benthic species described in detailed shore-based studies (Barker et al., 1988; Pälike
127 et al., 2010). We paid a particular attention to select solely entire, clean and well-preserved specimens
128 devoid of any clay aggregate and oxyhydroxide patch or coating (Figs. 3 a, c, d). Back-scattered
129 electrons (BSE) images of foraminifera tests and residues after chemical treatment were obtained

130 with using a JEOL JSM-6360LV electron microscope at the Géosciences Environnement Toulouse
131 laboratory (Toulouse, France) operating at 20 kV.

132 In a clean laboratory, for each sample foraminifera fraction (between ~0.5 mg and ~20 mg depending
133 of the foraminifera abundance) was rinsed with Milli-Q water, ultrasonized, and centrifuged 5 to 8
134 times. The supernatant was discarded between each cleaning step. This sample treatment allowed to
135 remove any potential source of contamination such as clay particles or coccolithophores. Then, we
136 added 500 μ L of Milli-Q water to each cleaned foraminifera fraction, prior to stepwise dissolution by
137 successively adding five times 100 μ L aliquots of 1N acetic acid, resulting in 1 mL of a 0.5N acetic
138 acid solution (e.g. Osborne et al., 2017). After ~10-20 minutes of reaction at room temperature, we
139 centrifuged each tube during 8 minutes at 4000 rpm. The pipetted supernatant was then transferred in
140 acid-cleaned Savillex beaker. Supernatants were dried on a hotplate at 70 °C. Residues were dissolved
141 in 30 μ L of double-distilled concentrated HCl and dried on a hotplate at 70 °C to remove any trace
142 of acetates. Stock solutions were prepared with 2N double-distilled HNO₃. All acids used for sample
143 preparation are trace element grade high purity acids. Quality controls of the described protocol are
144 further presented in the results section 3.2. *Test preservation and contamination evaluation.*

145 2.2. Elemental chemistry

146 Aliquot of the stock solution corresponding to ~1 mg of foraminifera was pipetted and dried on a
147 hotplate at 70 °C. 5 ml of 0.37N double-distilled HNO₃ solution doped with 1ppb In-Re solution were
148 added to the residue. Major and trace element concentrations were measured using a Thermo
149 Scientific Element XR HR-ICP-MS at the Observatoire Midi-Pyrénées (Toulouse, France). In and Re
150 were used as internal standards during ICP-MS measurements to monitor instrumental drift during
151 the course of the analyses. The instrument was calibrated using a synthetic multi-element solution
152 routinely used by the Observatoire Midi-Pyrénées ICP-MS service. This calibration solution has been
153 analysed with three different dilution factors, at the beginning and at the end of the analyses, in order
154 to calibrate the instrument response and correct the drift along the analytical sequence. Oxide
155 interferences (BaO⁺ on Nd⁺ and Eu⁺, LREE on MREE and MREE on HREE, see table in
156 Supplementary data 1) were then corrected. Oxide production rate was determined based on Cal-S
157 standard material, the latter having similar matrix than our foraminifera. The precision and accuracy
158 of ICP-MS analyses were assessed by measuring the reference materials GSJ CRM Jct-1 Giant Clam
159 (*Tridacna gigas*, i.e. biocarbonate, National Institute of Advanced Industrial Science and Technology,
160 Geological Survey of Japan, Inoue et al., 2004) and Cal-S (Calcite, Yeghicheyan et al., 2003).
161 Because of the very low REE content of Jct-1, accurate REE concentrations cannot be measured
162 through our "all-in-one" protocol for this standard (see the Tanaka et al., 2018 procedure combining
163 isotopic dilution, REE co-precipitation and chromatographic separation). Therefore, we used Cal-S,
164 which shows REE contents more comparable to our samples, for REE, Ca, Mg and Sr concentrations,
165 and Jct-1 for Ca and Sr contents as well as for Sr isotope ratio analysis.

166 2.3. Sr isotope ratio analysis

167 Aliquot of each stock solution corresponding to ~0.5 mg of foraminifera was pipetted, transferred to
168 a savillex beaker and dried on a hotplate at 70 °C. Sr was isolated from the matrix using liquid
169 chromatographic columns loaded with Sr-Spec® ion selective resin (Pin et al., 2014). Sr isotope ratios
170 were measured using a Thermo Scientific Triton Plus Thermal Ionization Mass Spectrometer (TIMS)
171 at the Observatoire Midi-Pyrénées (Toulouse, France). Measured ⁸⁷Sr/⁸⁶Sr ratios are an average of
172 150 measurements of ion intensities following the static multi-collection mode normalized to ⁸⁶Sr/⁸⁸Sr

173 = 0.1194. During the whole measurement period, NBS 987 standard was measured repetitively to
174 precisely quantify the external reproducibility. The NBS 987 measurements gave a mean $^{87}\text{Sr}/^{86}\text{Sr}$
175 isotope ratio of 0.710286 ± 0.000009 ($2\sigma\text{D}$, $n=18$). Thus we applied a correction of -0.000038 to all
176 our measurements to fit with the reference NBS987 value of 0.710248 . International and certified
177 carbonate standards were processed alongside our sample series. We measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of
178 0.709175 ± 14 ($2\sigma\text{D}$, $n=9$) for Jct-1 and 0.706953 ± 12 ($2\sigma\text{D}$, $n=5$) for Cal-S, in agreement with
179 certified values of $0.709150 \pm 0,000050$ (Ohno and Hirata, 2007) and $0,709169 \pm 0,000009$ (Weber
180 et al., 2017) and $0.706924 \pm 0,000018$ (Yeghicheyan et al., 2003), respectively.

181 3. Results

182 3.1. Age models

183 Both magnetostratigraphic and biostratigraphic data available for each site (Barker et al., 1988 for
184 Hole 689B and Pälake et al., 2010 for Hole 1334C) were considered to build age models. Data for the
185 Hole 689B have been updated to the GTS2012 (Gradstein et al., 2012). We used the R software (R
186 Development Core Team., 2019, packages *changepoint* of Killick and Eckley, 2014 and *segmented*
187 of Vito and Muggeo, 2003) in order to identify statistically robust regressions (of sedimentation rate)
188 and to calculate the absolute age of each selected sample (Fig. 2). The 95 % confident interval of
189 these regressions defines uncertainties on obtained ages (Fig. 2). The ages of the investigated samples
190 range from 24.6 Ma to 38.2 Ma for Hole 1334C and from 25.5 Ma to 46 Ma for Hole 689B (Fig. 2).
191

192 3.2. Test preservation and contamination evaluation

193 After having carefully hand-picked the cleanest complete foraminifera tests, we performed Scanning
194 Electron Microscopy (SEM) observations of both hand-picked foraminifera tests and residues after
195 chemical treatment. Observations revealed very well-preserved tests devoid of significant calcite
196 recrystallization (Fig. 3 a, b, c and d). The residues after foraminifera dissolution are clays, detrital
197 (e.g. quartz grains, mica) and early-diagenetic phases (e.g. framboidal pyrite), with intact morphology
198 and without evidence of partial dissolution. Thus, only carbonates were dissolved during the chemical
199 treatment applied on the samples (Fig. 3).

200 In addition, for all fractions we modelled clay/calcium carbonate geochemical mixing using the major
201 element compositions of planktonic and benthic foraminifera, in order to detect any potential clay
202 contamination during chemical treatment (Supplementary data 2). Considering the major elements of
203 clay minerals, e.g. notably Si and Al, the calculations always attest of less than 0.005 wt.% of potential
204 clay contamination (less than 0.002% for more than 95 % of the samples, Fig. 4). In other words, for
205 ~ 20 mg of foraminifera fraction, the maximum potential contribution of clay would be up to 1×10^{-3}
206 mg, thus extremely limited. This confirms that the cleaning and chemical protocol used here allowed
207 for proper extraction of the geochemical signature of foraminiferal biogenic calcite and thus to
208 reconstruct the seawater chemistry at the time of the tests formation.

209 3.3. Foraminifera as past seawater chemistry recorder

210 3.3.1. Rare earth element contents and Ce anomaly temporal evolution

211 All analyzed foraminifera fractions from the two investigated oceans display smooth PAAS (Post
212 Archean Australian Shales) normalized REE patterns typical of seawater (Fig. 5, Table 1 and
213 Supplementary data 1). They are characterized by light rare earth elements (LREE) depletion

214 compared to heavy rare earth element (HREE), (Pr/Yb)_N ratio ranging from 0.23 to 0.66 and 0.39 to
215 0.59 in PO and SAO, respectively. They also exhibit pronounced negative Ce/Ce* anomalies
216 [Ce/Ce* = C_{EN} / (Pr_N × (Pr_N/Nd_N))], Ce/Ce* ranging from 0.06 to 0.16 in PO and 0.21 to 0.58 into SAO
217 (Fig. 5), as well as positive La/La* anomalies [La/La* = La_N / (Pr_N × (Pr_N/Nd_N)²)], ranging from 1.30
218 to 2.26 in PO and 1.04 to 1.74 into SAO, both typical of modern seawater REE chemistry. This
219 confirms the reliability of our chemical treatment to extract clean seawater signals. Moreover, the
220 preservation of the both LREE depletion compared to MREE and HREE and the Ce/Ce* anomaly
221 discards any diagenetic overprint (e.g. Deng et al., 2017).

222 Interestingly, the intensity of the Ce/Ce* anomaly in foraminifera from PO and SAO is not random
223 and displays clear temporal evolutions though the investigated time period (Fig 6a). Both planktonic
224 and benthic foraminifera from PO exhibit very negative and rather constant Ce/Ce* anomalies
225 ranging from 0.06 to 0.16 between 35 and 26 Ma (Figs. 5a and 6a). In contrast, SAO foraminifera
226 show much lower Ce/Ce* anomalies between 48 and 31 Ma, which slightly increase over that time
227 period (e.g. Ce/Ce* = 0.51 at 46 Ma and 0.43 at 31 Ma in benthic foraminifera, Fig. 6a). Then, these
228 negative Ce anomalies strengthen, particularly in benthic foraminifera, until reaching similar values
229 to PO foraminifera *ca.* 26 Ma (Ce/Ce* down to 0.21, Fig. 6a). In SAO benthic foraminifera, the
230 La/La* positive anomaly strengthen after 29 Ma (La/La* < 1.30 before 29 Ma and La/La* > 1.65
231 afterwards, see Supplementary data 1).

232 3.3.2. Mg/Ca ratios temporal evolution

233 The foraminifera from the investigated sites display Mg/Ca molar ratios between 4.4 and 1.8
234 mmol/mol (Fig. 6b). From 34 Ma onwards, the Mg/Ca molar ratios of SAO and PO planktonic
235 foraminifera decrease (Fig. 6b). Then, as for the Ce/Ce* anomaly, a major change occurs from 31 Ma
236 onwards. After 31 Ma, Mg/Ca molar ratios of SAO planktonic foraminifera increase up to values
237 similar to PO planktonic foraminifera (3.7 mmol/mol for SAO at 27.3 Ma and 3.9 mmol/mol for PO
238 at 28.1 Ma, Fig. 6b in red and yellow, respectively). Contrary to planktonic foraminifera, Mg/Ca
239 molar ratios of SAO benthic species do not increase but remain stable from 31 Ma (Mg/Ca molar
240 ratios *ca.* 2.7 mmol/mol, Fig. 6b). Hence, SAO bottom and surface Mg/Ca signals, which were similar
241 at 46, 44 and 31 Ma, start to dissociate from one another between 31 and 29 Ma.

242 243 3.3.3. Strontium isotope ratios

244 All analyzed foraminifera display ⁸⁷Sr/⁸⁶Sr isotope ratios similar to those reported in the literature for
245 the Southern Ocean (Fig. 6c, Table 1 and Supplementary data 3). They follow a first order increasing
246 trend towards more radiogenic values from *ca.* 37 Ma onwards, which is commonly interpreted as a
247 consequence of increasing Sr continental weathering inputs associated with the Himalayan and
248 Andean orogenies (e.g. Raymo and Ruddiman, 1992). At the 31 Ma tipping point previously
249 identified in the Ce/Ce* and Mg/Ca records (Fig. 6a, b), a slight inflexion in the long-term rise of
250 ⁸⁷Sr/⁸⁶Sr isotope ratios is noticeable for both SAO and PO foraminifera (Fig. 6c). This inflexion,
251 highlighted by normalized derivatives of the ⁸⁷Sr/⁸⁶Sr curves (see Hodel et al., 2017, 2020 for a similar
252 approach), is noticeable for both planktonic and benthic foraminifera which have been analyzed
253 separately (Fig. 6c). It is also worth noticing that PO and SAO foraminifera ⁸⁷Sr/⁸⁶Sr isotope ratios
254 slightly, but significantly, differ from one another prior to 31 Ma, before displaying statistically
255 similar ⁸⁷Sr/⁸⁶Sr isotope ratios after 31 Ma (Fig. 6c). In details, PO foraminifera show more radiogenic
256 ⁸⁷Sr/⁸⁶Sr isotope composition than SAO ones by up to 0.000043 *ca.* 38 Ma (i.e. 9 times higher than

257 the analytical error of isotopic measurements and 4.8 times higher than the internal reproducibility),
258 before displaying statistically identical $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios from 31 Ma onwards (Fig. 6c).

259 4. Discussion

260 4.1. PO and SAO water masses mixing between 31 Ma and 26 Ma

261 The Mg/Ca molar ratios, Ce/Ce* anomalies and $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios of both planktonic and benthic
262 foraminifera from both studied sites show clear temporal variations, all characterized by significant
263 changes from *ca.* 31 Ma until *ca.* 26 Ma (Fig. 6). Understanding the reasons for these concomitant
264 variations relies on identifying the processes that could have influenced past seawater chemical
265 composition.

266 The integration of Mg ions in calcite during the foraminifera test precipitation being temperature
267 dependant (favoured by high temperature), the Mg/Ca molar ratio in marine biocarbonates is
268 extensively used to assess past seawater temperature evolutions: higher ratios signing higher
269 temperature and *vice-versa* (e.g. Barker et al., 2005). In our samples, Mg/Ca molar ratios of SAO and
270 PO planktonic foraminifera decrease from 34 Ma onwards, indicating a major seawater cooling at
271 that time (Fig. 6b). This cooling is synchronous with the onset of the Oligocene glaciation at the EOT,
272 evidenced by a marked increase of the $\delta^{18}\text{O}_{\text{bf}}$ values of benthic foraminifera worldwide (Zachos et
273 al., 2001) ($\sim +1.5$ ‰, Fig. 8a), including those from the Hole 689 in the SAO (Kennett and Stott, 1990,
274 in grey on Fig. 6b). Here, it is worth noting that both $\delta^{18}\text{O}_{\text{bf}}$ values of Kennett and Stott (1990) and
275 our Mg/Ca molar ratios of benthic foraminifera denote similar bottom seawater temperature
276 variations, attesting of the reliability of our Mg/Ca signals. It is nonetheless important to mention that
277 our multi-species approach and our low sampling resolution can possibly mask additional
278 complexities. Then, from 31 Ma onwards, Mg/Ca molar ratios of SAO planktonic foraminifera
279 increase until reaching values similar to PO planktonic foraminifera *ca.* 27 Ma (in red and yellow
280 respectively, Fig. 6b). This increase in Mg/Ca molar ratios of SAO planktonic foraminifera attests of
281 a progressive warming of SAO surface waters between *ca.* 31 and *ca.* 27 Ma. In addition, Mg/Ca
282 molar ratios of SAO planktonic and benthic foraminifera progressively become distinct from one
283 another during this time interval, suggesting a SAO stratification with warmer sea-surface water
284 masses than deep-seawater masses (Fig. 6b). This could sign the Drake Passage gateway opening.
285 Indeed, a such decoupling between bottom and surface water masses is in line with modelling studies
286 (e.g. Ladant et al., 2018; Zhang et al., 2020), which attests of a drastic decrease of the mixed layer
287 depth (MLD) in the SAO following the Drake Passage gateway opening. These studies model an
288 entirely mixed water column in the SAO before the Drake Passage gateway opening and a MLD at
289 less than 200 m after the Drake Passage gateway opening (e.g. Ladant et al., 2018; Zhang et al., 2020).
290 Therefore, the progressive surface waters warming that we observe between 31 Ma and 27 Ma
291 possibly results from a more and more restricted MLD along the Drake Passage gateway opening and
292 deepening, leading to the isolation of surface waters from cold bottom waters, hence promoting their
293 warming (Fig. 7).

294 The hypothesis of water masses stratification in the SAO due to PO input following the Drake Passage
295 gateway opening from 31 Ma and its deepening until at least 26 Ma is also supported by the Ce/Ce*
296 anomaly temporal evolutions. The Ce/Ce* anomaly is one of the most used seawater redox and
297 paleoredox proxy (e.g. Deng et al., 2017; Kamber et al., 2014; Osborne et al., 2017). Contrary to other
298 REEs that are only tetravalent (with the exception of Eu), Ce is characterized by two redox states, III
299 and IV. Under reducing conditions Ce^{3+} is soluble in water, while under oxidizing conditions CeO_2

300 precipitates. This Ce-oxide precipitation from seawater, i.e. Ce removal from the dissolved chemical
301 load, induce negative Ce/Ce* anomalies ($Ce/Ce^* < 1$, i.e. Ce depletion compared to its neighbouring
302 REE, e.g. Zhang and Nozaki, 1996; Osborne et al., 2017). Hence, variations of the Ce/Ce* anomaly
303 in seawater and so in foraminifera tests, can reflect dissolved pO_2 fluctuations (e.g. Zhang and Nozaki,
304 1996; Deng et al., 2017; Osborne et al., 2017). In parallel to the changes observed in the Mg/Ca molar
305 ratios, the temporal evolution of Ce/Ce* anomalies shows a major decrease in SAO benthic
306 foraminifera after the 31 Ma tipping point (Fig. 6a). Between 31 and 26 Ma, this pattern attests of a
307 progressive oxygenation of SAO deep water up to a redox state similar to the PO seawater at *ca.* 26
308 Ma (Fig. 6a). This suggests a clear redox state homogenization between PO and SAO deep waters
309 starting from 31 Ma and intensifying to 26 Ma, which argues for deep-water masses connexion
310 between these two oceans during this time interval.

311 Finally, the $^{87}Sr/^{86}Sr$ ratios of analyzed foraminifera tests also testify important paleoceanographic
312 changes through the studied interval (Fig. 6c). Despite similar rising trends, the $^{87}Sr/^{86}Sr$ ratios from
313 the PO and SAO appear decoupled around 38-36 Ma (up to 0.000043 of difference, Fig. 6), before to
314 merge together at 32 Ma and mark a rapid and global rise of $^{87}Sr/^{86}Sr$ ratios culminating at 31 Ma
315 (Fig. 6c). The initial offset of $^{87}Sr/^{86}Sr$ ratios recorded between the two basins is quite enigmatic
316 because the seawater $^{87}Sr/^{86}Sr$ ratio is generally considered as homogeneous worldwide. This is due
317 to the long residence time of Sr in the ocean (~ 2.5 Myr) with respect to the global ocean mixing time
318 (< 1500 years) (see McArthur et al. 2020 for a review). Recent studies with high analytical precisions
319 have well reported slight differences (± 0.000010), both spatially and bathymetrically, in the $^{87}Sr/^{86}Sr$
320 isotope ratio of modern oceanic waters far from any continental sources (Huang et al., 2011;
321 Mokadem et al., 2015). However, this isotopic heterogeneity is generally more typical of coastal
322 domains where seawater and carbonate shells are more directly influenced by river and submarine
323 groundwater Sr influxes (Jones et al., 2014; El Meknassi et al., 2018, 2020; Briard et al., 2020).
324 Hence, this primarily calls to question the preservation of initial oceanic signature. Here, the mixing
325 models based on the Si and Al concentrations of foraminifera tend to exclude any significant clay
326 contamination and to validate our analytical protocol (Fig. 4). However, Sr exchanges with pore water
327 and subsequent carbonate recrystallization during early-diagenesis cannot be totally excluded to
328 explain the more radiogenic signature of the PO foraminifera before 32 Ma. Yet, this hypothesis does
329 not fully explain neither why such processes only affected this Late Eocene interval, nor why
330 planktonic and benthic foraminifera records from same core samples were differently affected.
331 Moreover (1) we did not observe any evidence of such recrystallization in SEM observations (Fig. 3)
332 and (2) foraminifera from site 1334C older than 32 Ma display typical seawater-like REE pattern
333 characterized by very marked Ce anomalies (lowest Ce/Ce* ratio of all our sampling, Figs. 5 and 6),
334 incompatible with pore waters influenced recrystallization, pore waters being characterized by the
335 loss of the Ce anomaly of their originating seawater (e.g. Deng et al., 2017). In an alternative
336 hypothesis validating the genuine oceanic origin of our $^{87}Sr/^{86}Sr$ records, the observed offset between
337 SAO and PO foraminifera before 32 Ma would imply a partial decoupling of Sr reservoirs between
338 the Pacific and Southern Atlantic oceans at that time. Although impossible to consider in a modern
339 oceanic context dominated by global thermohaline circulation, such a mechanism was supposed at
340 the geological time scale for periods with a net degree of water mass restriction and stratification
341 between the oceanic basins (Banner 2004). This remains difficult to prove in our case but this
342 potential restriction is not incoherent with recent numerical models suggesting a very active Weddell
343 gyre isolating the Southern Atlantic domain in the Eocene, then vanishing when Pacific currents
344 intruded the Drake Passage gateway (Toumoulin et al. 2020). If true, the spatial decoupling of Sr

345 isotopic patterns could be due to more radiogenic Sr supplies from proto-Andean and Himalayan
346 reliefs in the Pacific and Indian oceans and low transfers towards Atlantic domains. Then, the merging
347 of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that we observe could sign PO and SAO water masses homogenization from *ca.* 31
348 Ma, consistently with the temporal evolutions of Mg/Ca molar ratios and Ce/Ce* anomalies
349 previously described (Fig. 6). This time was also marked by a global positive inflexion of the $^{87}\text{Sr}/^{86}\text{Sr}$
350 patterns possibly signing a global rise of continental weathering rates and radiogenic Sr flux to the
351 ocean. But given (1) the slighthness of this inflexion and (2) the 1 Ma decay between ours and compiled
352 data, this hypothesis needs to be further scrutinized, notably by extensive high-resolution
353 measurements of foraminiferal $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratio from Pacific, Atlantic (at low, equatorial and
354 high latitudes) and Indian oceans during the Eocene-Oligocene interval.

355 Even if additional, notably higher resolution data are required, altogether Mg/Ca molar ratios, Ce/Ce*
356 anomalies and Sr isotope ratios of PO and SAO foraminifera attest to a PO/SAO connexion from *ca.*
357 31 Ma and to an established deep-seawater flow *ca.* 26 Ma. This provides a better timing of the Drake
358 Passage gateway opening, in line with the tectonic synthesis of Eagle and Jokat (2014) who conclude
359 that the Drake Passage gateway likely opened to deep seawater flow between 30 and 26 Ma.

360

361 4.2. Reconsideration of the Neodymium isotope record

362 If our results and interpretations fit with the last tectonic synthesis of Eagle and Jokat (2014) and most
363 of the other tectonic studies previously cited, they are contradictory with the most cited "chemical
364 age" attributed to the Drake Passage gateway opening: 41 Ma inferred from the fossil fish teeth
365 Neodymium isotope records (Scher and Martin 2006).

366 Besides tectonic and sedimentary studies, previous dating of the Drake Passage gateway opening
367 through geochemical tracers is essentially based on the record of fossil fish teeth Neodymium isotopic
368 compositions ($^{143}\text{Nd}/^{144}\text{Nd}$, expressed as $\epsilon_{\text{Nd}} = ((^{143}\text{Nd}/^{144}\text{Nd}_{\text{sample}} / ^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}}) - 1) * 10\,000$; with
369 CHUR standing for Chondrite Uniform Reservoir, $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$; Bouvier et al., 2008). Nd
370 has a short residence time of *ca.* 500 years in the ocean compared with to the global oceanic overturn
371 of *ca.* 1500 years, making it sensitive to modification of seawater mass circulation (e.g. Scher and
372 Martin, 2006; Scher et al., 2015). Because fossil fish teeth are considered to record bottom seawater
373 REE content, their Nd isotope compositions are used to track seawater masses through geological
374 time (e.g. Martin and Haley, 2000; Scher and Martin, 2004). However, seawater Nd isotope
375 composition has to be carefully reconsidered due to its high sensitivity to continental-weathering
376 fluxes to the ocean, particularly in proximal oceanic domains (e.g. Frank, 2002; Lacan and Jeandel,
377 2005; Jeandel et al., 2007; Jeandel, 2016).

378 At present, the most consensual "isotopic age" for the Drake Passage gateway opening is around 41
379 Ma (Scher and Martin 2006). This hypothesis relies on an ϵ_{Nd} increase toward more radiogenic values
380 in fish teeth from the Agulhas Ridge (from $\sim -8 \epsilon_{\text{Nd}}$ at *ca.* 41 Ma to $-6 \epsilon_{\text{Nd}}$ at *ca.* 39 Ma, site 1090,
381 Fig. 8b in blue) and to a much slighter extent at Maud Rise (site 689, Scher and Martin, 2004, Fig. 8b
382 in red). This Nd isotope excursion is interpreted as signing Pacific radiogenic seawater inflow (in
383 yellow) to Southern Ocean from 41 Ma (Fig. 8b). Nonetheless, compiled data for the Kerguelen
384 Plateau, which is supposed to have also been strongly affected by Pacific seawater inputs through the
385 Drake Passage gateway do not show such an increasing ϵ_{Nd} trend starting at 41 Ma. Then, from *ca.*
386 37 Ma to *ca.* 34 Ma, a second ϵ_{Nd} increase is noticeable in fossil fish teeth from for Agulhas Ridge
387 and Maud Rise, as well as in those of the Kerguelen Plateau (Fig. 8b). The latter is also interpreted
388 as enhanced Pacific radiogenic seawater inputs related to the Drake Passage gateway opening (Scher

389 and Martin, 2004, 2006). On the other hand, recent fish teeth Nd isotope data from Huck et al. (2017),
390 which enlarge the Kerguelen plateau ϵ_{Nd} record to the early Eocene, attest that ϵ_{Nd} values gradually
391 increased from $\sim -10 \epsilon_{Nd}$ at 50 Ma to $\sim -7 \epsilon_{Nd}$ at 35 Ma (Fig. 8b, in green). This suggests that either
392 (1) the Drake Passage gateway opened to significant seawater flow as soon as the early Eocene (*ca.*
393 50 Ma), or (2) this trend reflects the contribution of an additional Nd source, other than radiogenic
394 Pacific seawater input into Southern Ocean.

395 If some tectono-sedimentary studies attest of a possible shallow opening of the Drake Passage
396 gateway at the early-Eocene (e.g. Livermore et al., 2005, 2007), most of them argues for its opening
397 to eastward deep circulation during the Oligocene (Latimer and Fillipelli, 2002; Lawver and Gahagan,
398 2003; Eagles et al., 200; Livermore et al., 2005, 2007; Brown et al., 2006; Lagabriele et al., 2009;
399 Eagles and Jokat, 2014) (Fig. 1). Hence, geological evidences of these studies do not support a
400 significant connection between Pacific and Southern oceans through the Drake Passage gateway as
401 early as 50 Ma to explain the ϵ_{Nd} increasing trend recorded in fish teeth from the Kerguelen Plateau
402 (Fig. 8b). This supports the hypothesis of an additional Nd source, potentially continental weathering
403 fluxes from Antarctica related the Eocene-Oligocene cooling and associated ice-sheet formation (Fig
404 8a). Indeed, potential Antarctica's weathering regime evolutions have to be carefully considered in
405 the interpretation of Southern Ocean fossil fish teeth ϵ_{Nd} record because of (1) the highly non-
406 radiogenic Nd isotope composition of Antarctica's lithologies bordering the Southern Atlantic and
407 Indian oceans (e.g. core top sediments reach $-21.3 \epsilon_{Nd}$ near the Prydz Bay, Roy et al., 2007; Van de
408 Fliedert et al., 2007; Scher et al., 2014), (2) the vicinity of Antarctica to the Southern Ocean sites
409 investigated for Eocene-Oligocene paleoceanographic studies (Kerguelen Plateau, Maud Rise and
410 Agulhas Ridge, Fig. 1), and (3) the isolated setting of the Southern Ocean before the Drake and
411 Tasmanian gateways deep opening.

412 Punctually, brief episodes of intense continental export associated with glacial pulses, e.g. the PrOM-
413 event (*ca.* 37 Ma) and EOT (*ca.* 34 Ma) glaciations, are marked by negative ϵ_{Nd} excursions of 1.5 to
414 3 ϵ_{Nd} units of amplitude in fossil fish teeth from the Kerguelen Plateau (Scher et al., 2011, 2014) (Fig.
415 8). In these studies, as well as in Huck et al (2017), authors attributed these excursions to brutal
416 changes in weathering regime (glacial dynamic onset and debacle episodes) of Archean-Proterozoic-
417 Early Phanerozoic Antarctica's lithologies (Roy et al., 2007; Musumeci et al., 2000), delivering
418 highly non-radiogenic Nd flux to the ocean, hence drawing down the ϵ_{Nd} record toward less
419 radiogenic values in fossil fish teeth at that time. This clearly demonstrates the high sensitivity of the
420 fish teeth Nd record to variations of continental weathering fluxes, which can produce ϵ_{Nd} shift of a
421 similar amplitude compared to those imputed to oceanic circulation changes (e.g. potential Pacific
422 input from *ca.* 41 Ma and 37 Ma, Scher and Martin, 2006) (Fig. 8b). This questions the effect of
423 longer climate phenomenon and the associated weathering regime changes on the ϵ_{Nd} fish teeth
424 record, notably along the Eocene cooling that led to the Oligocene glaciation.

425 Indeed, the Eocene cooling drastically impacted Antarctica's continental weathering regimes, as
426 evidenced by the emergence of illite to the detriment of smectite - signing hydrolysis and so chemical
427 weathering weakening - and by the concomitant occurrence of detrital chlorite in the Maud Rise and
428 Agulhas Ridge sedimentary piles (notably from 41-40 Ma, Ehrmann and Mackensen, 1992;
429 Diekmann et al., 2004). Thus, the progressive reduction of chemical inputs from Antarctica along the
430 Eocene cooling likely resulted in a decrease of highly non-radiogenic Nd fluxes from Antarctica
431 during this time interval. This is consistent with, and could explain, the progressive increasing trend
432 toward more radiogenic ϵ_{Nd} values of fish teeth from the Kerguelen plateau from the early-Eocene (\sim

433 -10 ϵ_{Nd} at 50 Ma) to the EOT (e.g. $\sim -7 \epsilon_{Nd}$ at 35 Ma) (Fig. 8b, in green). Similarly, increasing ϵ_{Nd}
434 trends observed from 41 Ma at the Agulhas Ridge (and Maud Rise) and from 37 Ma at Maud Rise,
435 the Agulhas Ridge and the Kerguelen Plateau (Fig. 8b), could also result from a diminution of
436 continental weathering fluxes from Antarctica, coupled with, or rather than, Drake Passage gateway
437 opening and Pacific seawater inputs (Scher and Martin, 2004, 2006). Indeed, this two-stepwise
438 increase of the Southern Ocean ϵ_{Nd} record is synchronous with the two main glacial events preceding
439 the EOT glaciation, namely the first icecaps evidences at Antarctica *ca.* 42-41.5 Ma (Tripathi et al.,
440 2005) accompanied by the clay mineralogical evolutions previously mentioned *ca.* 41 Ma, and the
441 PrOM-event *ca.* 37 Ma (Fig. 8).

442 This high sensitivity of the Southern Ocean ϵ_{Nd} fish teeth record to both punctual and long-term
443 Antarctica's weathering fluxes evolutions, opens up the door to the idea of a buffering of the Southern
444 Ocean (here Kerguelen Plateau, Maud Rise and Agulhas Ridge) fish teeth Nd isotope compositions
445 by highly non-radiogenic continental weathering fluxes from Antarctica. This "buffering hypothesis"
446 is also consistent with the slight ϵ_{Nd} offset ($\sim 0.5 \epsilon_{Nd}$) noticed by Wright et al. (2018) between the sites
447 744 (less radiogenic and closer to Antarctica) and 748 (more radiogenic and further away from
448 Antarctica) which could testify of a latitudinal gradient of this buffering. In addition, the vicinity of
449 the Maud Rise and the Kerguelen Plateau sites with Antarctica could explain why no major changes
450 have been identified in the ϵ_{Nd} fish teeth records from these localities after 33 Ma (Wright et al.,
451 2018). Climate and so continental weathering having stabilized after the EOT glaciation onset, this
452 stable ϵ_{Nd} record could be the expression of the ϵ_{Nd} buffering by Antarctica's weathering fluxes, which
453 could have hindered the detection of the Drake Passage gateway opening Nd isotope signal (i.e.
454 Pacific radiogenic water input). Therefore, the Nd isotope fish teeth records from high latitude
455 Southern Ocean sites might represent an Antarctica's continental weathering history rather than a
456 Southern Ocean's seawater masses history.

457 If the Southern Ocean Nd isotope record was indeed partially buffered by Antarctica continental
458 weathering fluxes, identifying Pacific and Southern oceans mixing would rely on the Pacific record.
459 The latter had a much more radiogenic ϵ_{Nd} signature than any other oceans during the Eocene (Fig.
460 8b). Thus, an abrupt collapse in the ϵ_{Nd} temporal evolution of PO fossil fish teeth should occur in case
461 of mixing with Southern Ocean non-radiogenic seawater. An extensive fossil fish teeth records have
462 been made available by Scher et al. (2015) for the PO (Fig. 8b, in yellow). A first drop in ϵ_{Nd} values
463 of fossil fish teeth from the Western Tasmanian Margin (site 1168) occurs between 36 Ma and 33 Ma
464 (down to $-7 \epsilon_{Nd}$ while the Pacific endmember shows much more radiogenic values from -4 to $-5 \epsilon_{Nd}$)
465 possibly signing the Tasmanian gateway opening (Stickley et al., 2004). Then, Scher et al. (2015)
466 evidenced a major ϵ_{Nd} collapse in the Southern Pacific Ocean record (sites 1124, 1168 and 1172) from
467 *ca.* 30 Ma (Fig. 8b). Authors interpreted this collapse as signing the ACC onset and Pacific-Southern
468 Ocean connection following westerlies alignment with a previously opened Tasmanian gateway
469 (deep seawater circulation through the Tasmanian gateway starting from 35.5 to 33.5 Ma, Stickley et
470 al., 2004). In this study, authors involved the migration of the northern margin of the Tasmanian
471 gateway into the zone of the westerlies as the trigger for the ACC because they consider the Drake
472 Passage gateway as being already open by the late Eocene (*ca.* 41-37 Ma, Scher and Martin, 2006).
473 However, as previously developed, this Eocene opening is based on the Southern Ocean fish teeth
474 Nd isotopes record (at Agulhas Ridge and Maud Rise), and can also be related to decreases of
475 weathering rates on Antarctica in response to the Eocene cooling (particularly the first icecap
476 formation *ca.* 42-41 Ma and the PrOM-event *ca.* 37 Ma, e.g. Tripathi et al., 2005; Scher et al., 2014,
477 Fig. 8) rather than to paleoceanographic changes. Alternatively, Scher et al. (2015) did not discard

478 the influence of far field tectonic in that ϵ_{Nd} collapse, notably the influence of the Drake Passage
479 gateway opening. Interestingly, this major collapse of ϵ_{Nd} values at Pacific sites is synchronous with
480 the SAO-PO connection evidenced from *ca.* 31 Ma (with a probable deepening to 26 Ma) by our
481 Mg/Ca, Ce/Ce* and Sr isotopes data from Maud Rise (site 689) and the Pacific Ocean (site 1334)
482 (Figs. 6 and 8). We thus propose that this ϵ_{Nd} collapse more likely signs the ACC onset following the
483 Drake Passage gateway opening rather than, or combined with, westerlies alignment with the
484 Tasmanian gateway (Scher et al., 2015). This statement is in adequacy with other geochemical tracers
485 such as Si isotopes on sponge tests that attests of a PO-SAO connection from *ca.* 31.5 Ma (Egan et
486 al., 2013), as well as most tectonic and sedimentary studies, which argue for an Oligocene Drake
487 Passage gateway opening to deep seawater circulation (see above).

488

489 4.3. Implication for the global climate and ocean circulation

490 Here, we evidence a progressive Drake Passage gateway opening to significant seawater flow starting
491 from 31 Ma and its deepening to 26 Ma. In this case, the Drake Passage gateway opening postdates
492 the Tasmanian gateway aperture (35.5-33.5 Ma, Stickley et al., 2004). Hence, the ACC onset has
493 necessarily relied on the Drake Passage gateway opening and so has likely been initiated during the
494 31-26 Ma interval. Within this temporal framework, the antecedence of the Oligocene glaciation onset
495 (33.7 Ma) relative to the ACC onset, possibly from 31 Ma and strengthening from 31 Ma to 26 Ma,
496 confirms that the ACC alone did not trigger the Oligocene glaciation. This reinforces the pCO_2
497 hypothesis, which states that the Oligocene glaciation resulted from atmospheric pCO_2 drawdowns
498 below threshold value of ~ 750 ppm (DeConto and Pollard, 2003; Pagani et al., 2011; Ladant et al.,
499 2014; Galeotti et al., 2016) under favourable orbital configurations (i.e. low eccentricity and low-
500 amplitude change in obliquity resulting in dampened seasonality, Coxall et al., 2005). However,
501 mechanisms involved in the Eocene atmospheric pCO_2 decrease are still debated. It has been
502 hypothesized that even a shallow opening of the Drake Passage gateway could have influenced the
503 evolution of pCO_2 prior to the Oi-1 and the onset of the ACC (Eagles et al., 2006). However, increases
504 in continental weathering are arguably the most important mechanism for this long-term atmospheric
505 pCO_2 drawdown (e.g. Raymo and Ruddiman, 1992; Zachos et al., 1999; Ravizza and Peucker-
506 Ehrenbrink, 2003; Van der Flierdt, 2011; Elsworth et al., 2017). Yet, the mechanisms involved in that
507 continental weathering increase remain ill-constrained. Additionally, the Antarctic ice sheet itself
508 may have provided a positive feedback by impacting the ocean circulation, chemistry, and
509 productivity and so atmospheric pCO_2 (e.g. Miller et al., 2009) while the albedo effect of the growing
510 Antarctic ice-sheet could have contributed to cooler seawater temperature (e.g. DeConto et al., 2007).
511 Therefore, to better constrain the mechanisms involved in the Eocene-Oligocene pCO_2 drawdown
512 and the triggering of the Oligocene glaciation itself remain a prime topic for further studies to better
513 constrain the Earth's global climate feedbacks.

514 The Antarctica's gateways opening is also of a major interest not only for ACC onset itself but for the
515 whole thermohaline circulation, which cannot initiate without southern connections between Pacific,
516 Atlantic and Indian oceans (e.g. Stow, 2001; Katz et al., 2011; Straume et al., 2020; Toumoulin et al.,
517 2020). Indeed, "the modern four-layer ocean structure and increased export of Northern Component
518 Water to the South Atlantic has been related to the development of the Antarctic Circumpolar
519 Current" (in Vahlenkamp et al., 2018), the latter having been allowed by the Drake and Tasmanian
520 gateways opening (e.g. Stow, 2001; Katz et al., 2011). The seawater masses stratification that we
521 evidence through Mg/Ca molar ratios and Ce/Ce* anomalies and that we interpret as signing the
522 Drake Passage gateway opening in the 31-26 Ma interval could have participated to the establishment

523 of the Antarctic Bottom Water (AABW) (Katz et al., 2011; Stow, 2001). Such proto-AABW
524 formation could then have promoted the Atlantic Meridional Overturning Circulation (AMOC),
525 hence initiating a global thermohaline circulation (Katz et al., 2011; Elsworth et al., 2017). Therefore,
526 the date that we propose in this contribution for the Drake Passage gateway opening and ACC
527 initiation also possibly signs the premises of a modern-like thermohaline circulation during the 31-
528 26 Ma time interval.

529

530 5. Conclusion

531 In summary, Mg/Ca molar ratios, Ce/Ce* anomalies and $^{87}\text{Sr}/^{86}\text{Sr}$ isotope compositions acquired on
532 both benthic and planktonic foraminifera attest to a connection between the PO and SAO from 31 Ma
533 onwards, reflecting the Drake Passage gateway opening and its progressive deepening to 26 Ma.
534 These results, in adequacy with the last tectonic-sedimentary synthesis on the Drake Passage gateway
535 and with our revised interpretation of the Pacific ϵ_{Nd} record (ϵ_{Nd} collapse ~ 30 Ma), provide a better
536 timing of the Drake Passage gateway opening. Since the Tasmanian gateway was already open to
537 deep seawater circulation at that time, the Drake Passage gateway opening between 31 Ma and 26
538 Ma likely led to the establishment of the ACC during this time interval. Also, this opening of the
539 Drake Passage gateway likely played a crucial role in the modern-like thermohaline circulation onset
540 and the AABW establishment. Finally, the antecedence of the Oligocene glaciation onset (33.7 Ma),
541 relative to the Drake Passage gateway opening to deep seawater circulation, and ACC onset (31-26
542 Ma) confirms that the ACC alone did not trigger the Oligocene glaciation even if it may have played
543 and still plays a central role into the stabilization of the following icehouse climate. Thus, the abrupt
544 Oligocene glaciation onset would have been triggered by $p\text{CO}_2$ drawdown below threshold values
545 (below ~ 750 ppm), whether or not coupled with favourable orbital configurations, as previously
546 proposed and modelled. To identify the contributions of the various forcing leading to major $p\text{CO}_2$
547 decline, such as tectonic, continental weathering, AMOC and nutrient supply, and to understand their
548 influence on the Earth's climate dynamics is thus fundamental as much for the past as for the future.

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847 *Figure 1: (a, b) Compilation of Drake Passage gateway opening ages reported in literature (Latimer and Filippelli,*
 848 *2002; Lawver and Gahagan, 2003; Livermore et al., 2005, 2007; Pfuhl and McCave, 2005; Brown et al., 2006; Eagles*
 849 *et al., 2006; Scher and Martin, 2006; Barker et al., 2007; Lagabrielle et al., 2009; Egan et al., 2013; Borrelli et al., 2014;*
 850 *Scher et al., 2015; Sarkar et al., 2019) (soft colours are for shallow depth aperture and dark colours for deep aperture,*
 851 *curves in b are kernel density obtained using the Past 3 software (Hammer et al., 2001). The deep aperture of the*
 852 *Tasmanian gateway between 35.5 and 33.5 Ma (Stickley et al., 2004) is also reported (in brown). The polar-centered map*
 853 *(Martinson, 2012) schematizes the clockwise Antarctic Circumpolar Current following the Tasmanian and Drake*
 854 *Passage gateways opening (in blue). Submarine areas above 3000 m depth are represented in light grey and land in*
 855 *black. Studied ODP/IODP legs are also reported as well as the Agulhas Ridge and the Kerguelen Plateau locations,*
 856 *which are further considered in the discussion.*

857 *Figure 2: Age models for samples (large coloured symbols) from the IODP Leg 320 Hole 1334 C (a) and ODP Leg 113*
 858 *Hole 689 B (b). Calibrations are based on a synthesis of magnetostratigraphic and biostratigraphic data symbolized by*
 859 *black dots (Barker et al., 1988; Pälike et al., 2010). The age data of the Leg 113 have been updated according to the*
 860 *Geological Timescale 2012 (GTS2012). Regressions are computed using the packages changepoint (Killick and Eckley,*
 861 *2014) and segmented (Vito and Muggeo, 2003) on the R software (R Development Core Team., 2019). Uncertainties*
 862 *(represented in shaded colours) correspond to the 95 % confident interval of regressions.*

863 *Figure 3: Secondary electrons SEM microphotographs of hand-picked foraminifera (a, b, c, d) and residual fractions*
 864 *remaining after the cleaning and the acid dissolution procedure (e, f). The samples include planktonic (Globigerina; a,b)*
 865 *and benthic (Nodosaria and Gyroidinoides, respectively c and d) foraminifera. No evidence of secondary calcite*
 866 *recrystallization is noticeable, attesting of the pristine character of tests. (e) Residue after cleaning and dissolution*
 867 *procedure showing undissolved and discarded quartz grains (Qtz), amorphous silica from radiolarian tests (Si), pyrite*
 868 *(Pyr) and clay minerals (Clays). (f) Undissolved and discarded clay platelets.*

869 *Figure 4: Clay-CaCO₃ mixing models for Si/Ca (a) and Al/Ca (b) (μmol/mol) ratios attesting of the absence of noticeable*
 870 *clay contamination after our cleaning and digestion protocol (always less than 0.005 % of potential clay contamination,*
 871 *less than 0.002 % for 95% of the samples). Major element contents of illite, smectite and kaolinite used for calculations*
 872 *are from Deer et al. (2013). Box-and-whiskers plots showing the distribution of foraminifera samples from the 1334C*
 873 *(yellow) and 689B (red) sites were obtained using the Past3 software (Hammer et al., 2001).*

874 *Figure 5: PAAS normalized REE spectra of the planktonic and benthic foraminifera tests from site 1334C (a) and 689B*
 875 *(b). PAAS normalizing values are from Barth et al. (2000) with Tm from Kamber et al. (2005). Modern seawater REE (x*
 876 *10⁶) spectrum are represented for comparison (data are from Grenier et al., 2013 for the equatorial PO and from*
 877 *Hathorne et al., 2014 for the SAO).*

878 *Figure 6: Evolution of Ce/Ce*_N (a), Mg/Ca (b) and ⁸⁷Sr/⁸⁶Sr ratios (c) recorded by the benthic (light colour) and*
 879 *planktonic (dark colour) foraminifera from the studied sites 689B and 1334C along the Eocene-Oligocene interval (EOT).*
 880 *Because it has been shown that planktonic foraminifera could exchange REE with seawater along their rain-down*
 881 *through the water column and once on the seafloor (e.g. Roberts et al., 2010; Osborne et al., 2017), temporal evolutions*
 882 *of their Ce/Ce* anomaly cannot be interpreted as signing surface seawater redox state evolutions but an integrated signal*
 883 *of the whole water column. PAAS normalizing values are from Barth et al. (2000). The Mg/Ca data are compared to the*
 884 *benthic foraminifera δ¹⁸O data of (Kennett and Stott, 1990) of the site 689B (ODP Leg 113). Normalized variation curves*
 885 *are normalized prime derivatives of smoothing spline functions of the ⁸⁷Sr/⁸⁶Sr data generated with the R software (R*
 886 *Development Core Team., 2019). Error bars for both ages and ⁸⁷Sr/⁸⁶Sr ratios are similar to symbol size. The red and the*
 887 *yellow domains drawn for sites 689B and 1334C represent the envelopes of the maximal error bars. Compiled ⁸⁷Sr/⁸⁶Sr*
 888 *data for the SO are from (Barrera et al., 1991; Mead and Hodell, 1995; Zachos et al., 1999; Scher and Martin, 2004).*

889 *Figure 7: Schematic representation of the hypothesized mixed layered depth evolution at Maud Rise (SAO) in response*
 890 *to Drake Passage gateway opening based on numerical modelling studies of Ladant et al. (2018) and Zhang et al. (2020)*
 891 *and on our data (Fig. 6). This stratification scenario fits with (1) our Mg/Ca data indicating surface water warming and*
 892 *bottom water remaining cold between 31 and 27 Ma and (2) bottom SAO seawater oxygenation indicated by negative Ce*

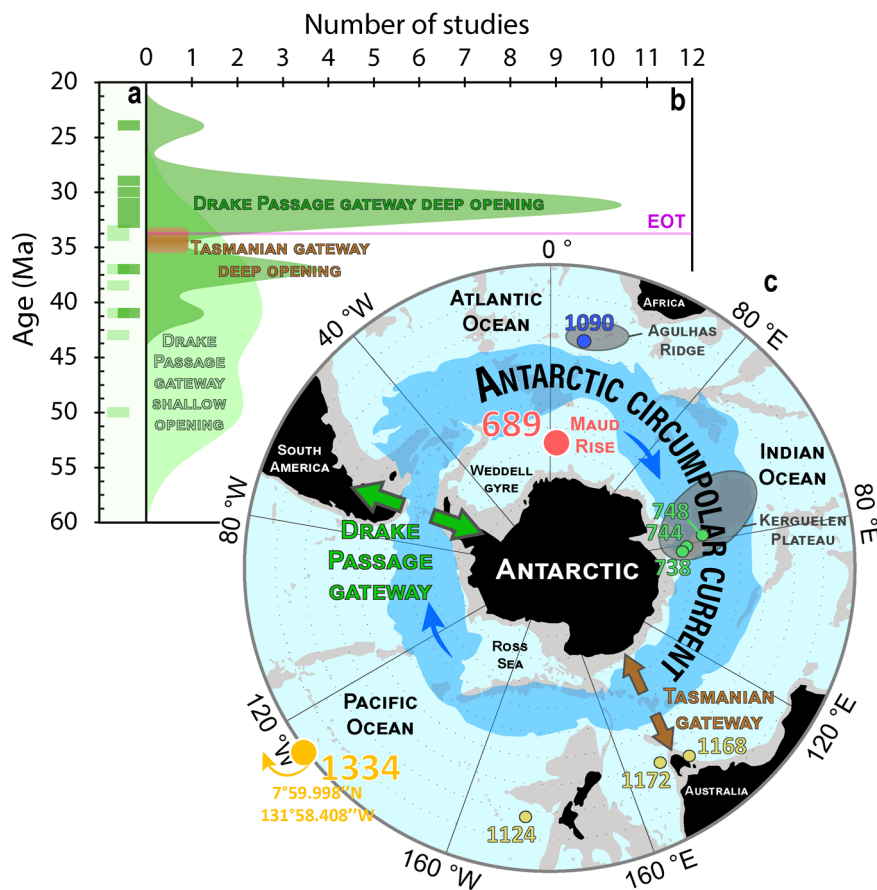
893 anomalies increase along the 31-26 Ma interval, up to reach similar value to deep pacific waters. SAO/PO seawater
 894 masses homogenization is also suggested by Sr isotopes ratios that are dissociated before 31 Ma and identical after.
 895 Approximate paleo-depths are from Barker et al. (1988).

896 Figure 8: (a) Compilation of benthic foraminifera $\delta^{18}O$ for the high latitudes of the Southern Ocean (e.g., Maud Rise and
 897 Kerguelen Plateau, in red), Southern Atlantic Ocean (in grey), and Pacific Ocean (in yellow) by Cramer et al. (2009).
 898 (b) Compilation of published fossil fish teeth ϵ_{Nd} data for the Maud Rise (in red, site 689), the Kerguelen Plateau (in
 899 green, sites 738, 744, 748), the Agulhas Ridge (in blue, site 1090) and the Pacific Ocean (in yellow, sites 192, 277, 464,
 900 465, 883, 884, 1124, 1168, 1172, 1208, 1209, 1211) during the Eocene-Oligocene interval (Thomas, 2004; Scher and
 901 Martin, 2004, 2006, 2008; Scher and Delaney, 2010; Scher et al., 2011, 2014, 2015; Hague et al., 2012; Huck et al.,
 902 2017; Wright et al., 2018). Punctual weathering flux increases are from Scher et al. (2011, 2014). Clay minerals changes
 903 are from Ehrmann and Mackensen (1992) and Diekmann et al. (2004).

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907 Figure 1

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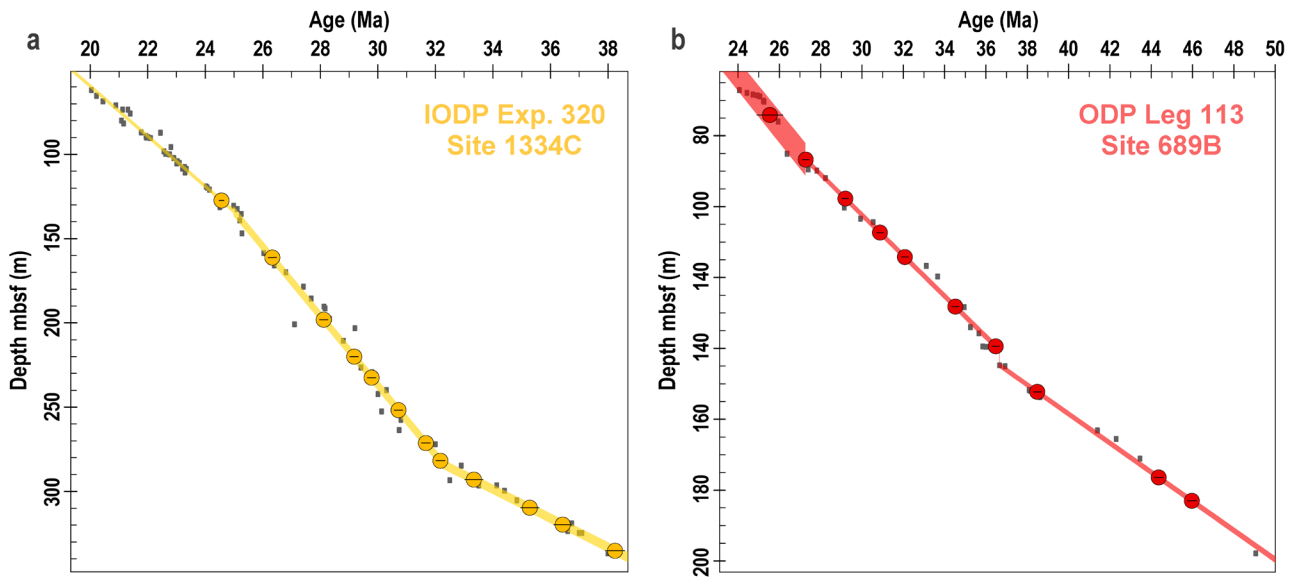
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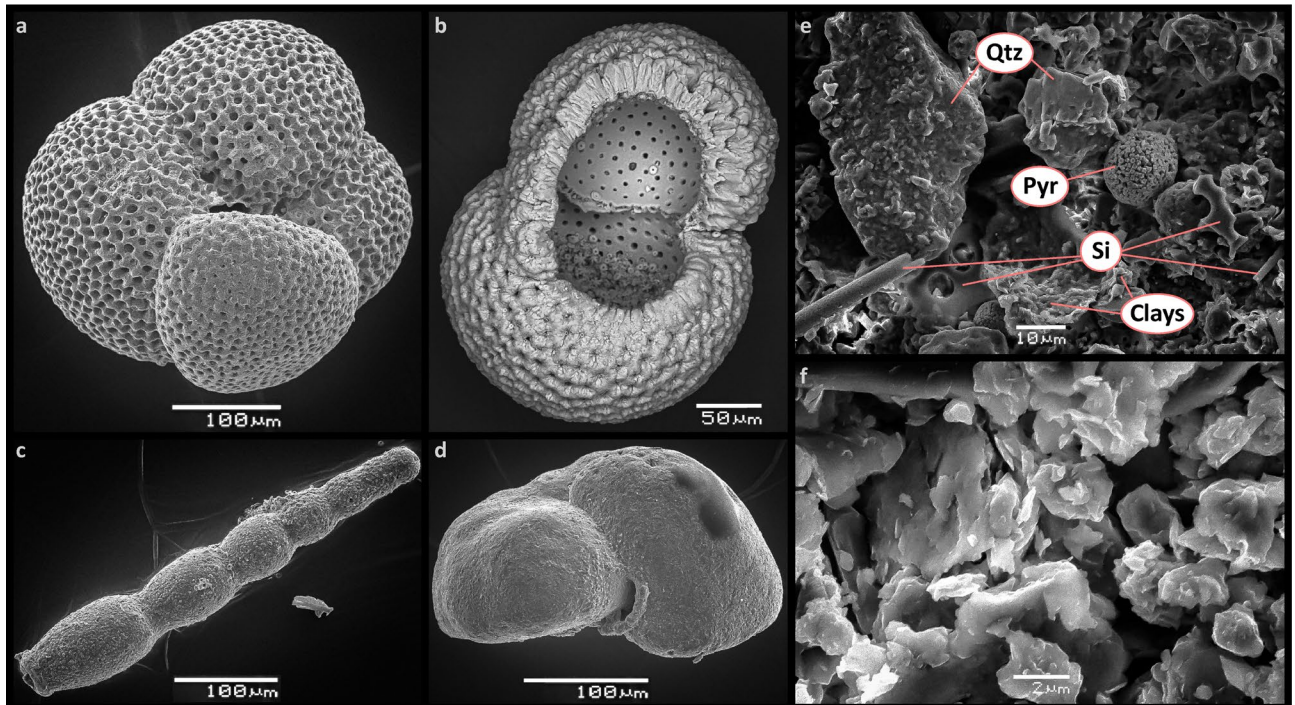


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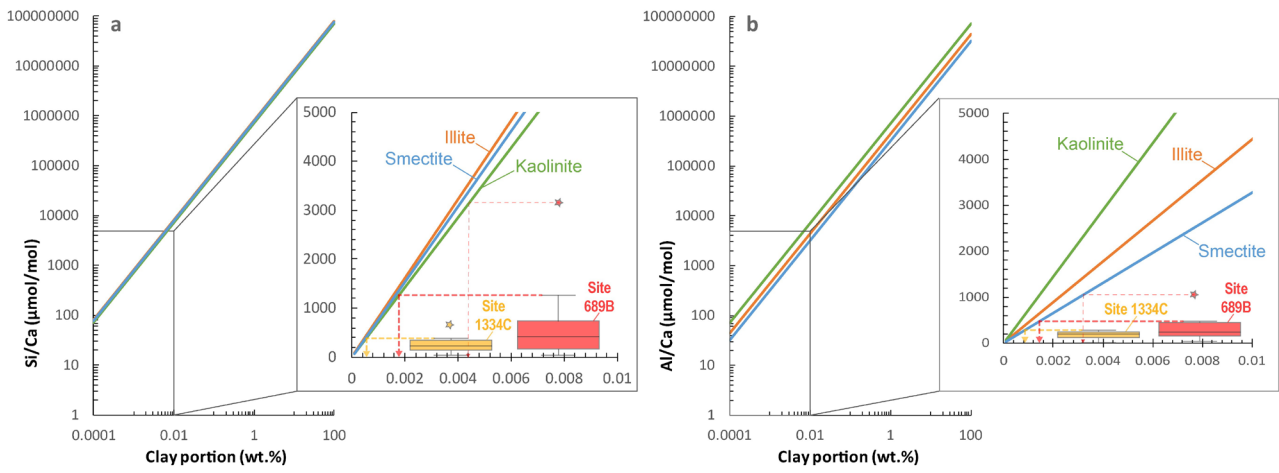
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919 *Figure 3*

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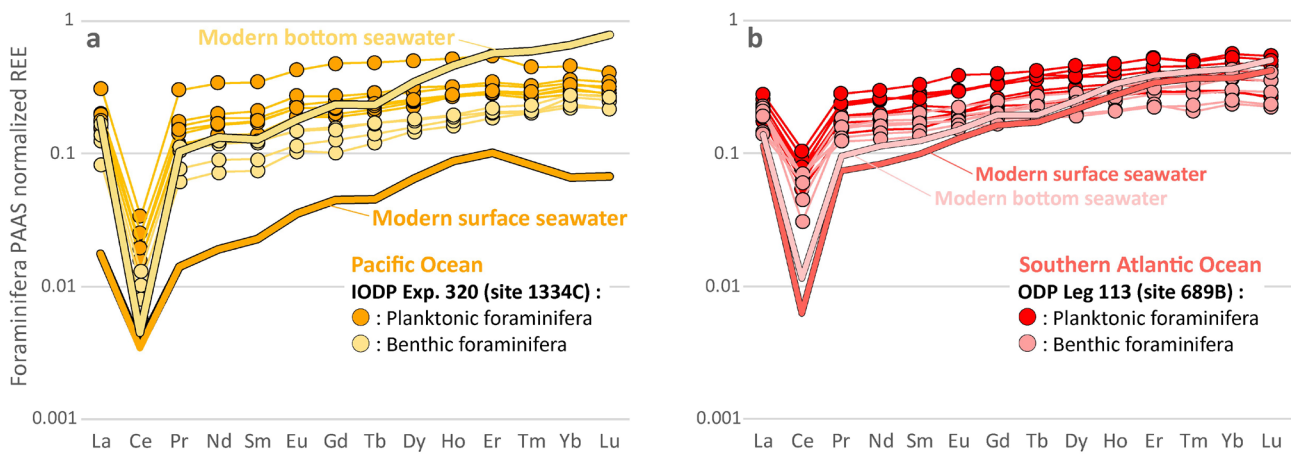
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Figure 4

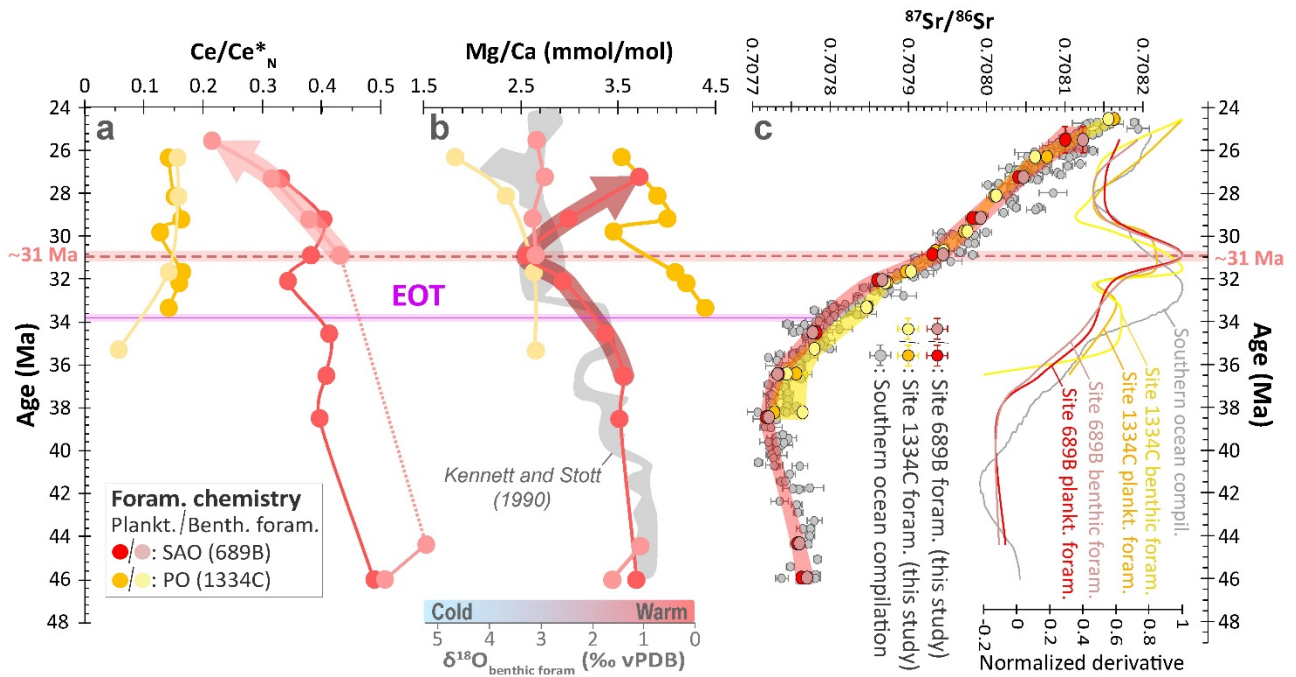


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Figure 5

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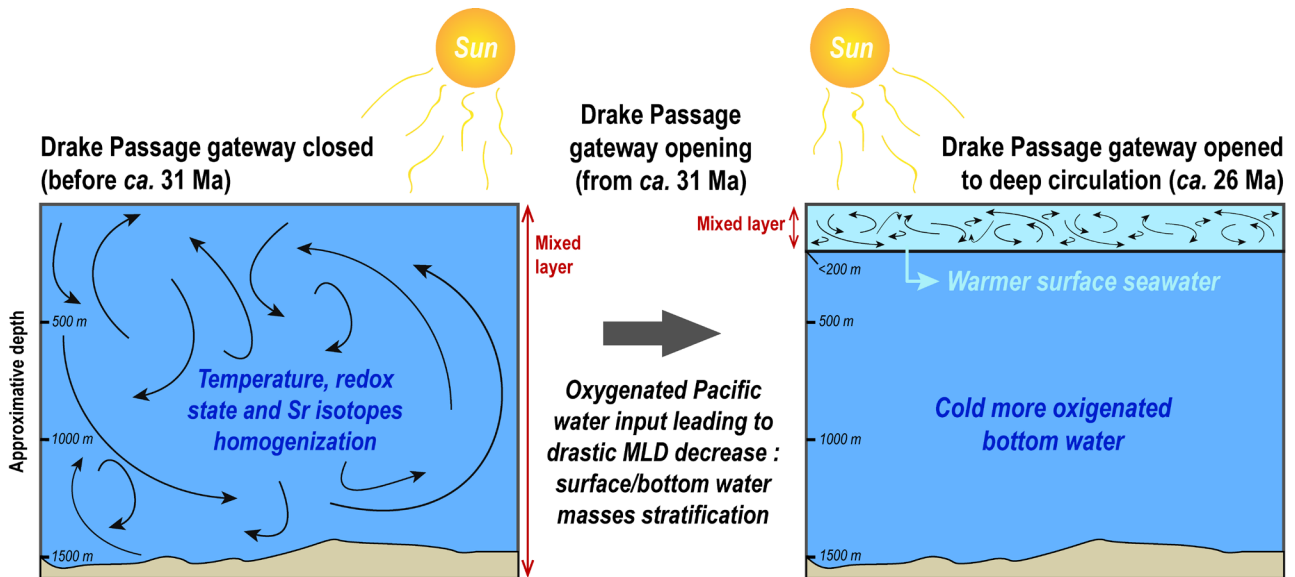
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940 Figure 6

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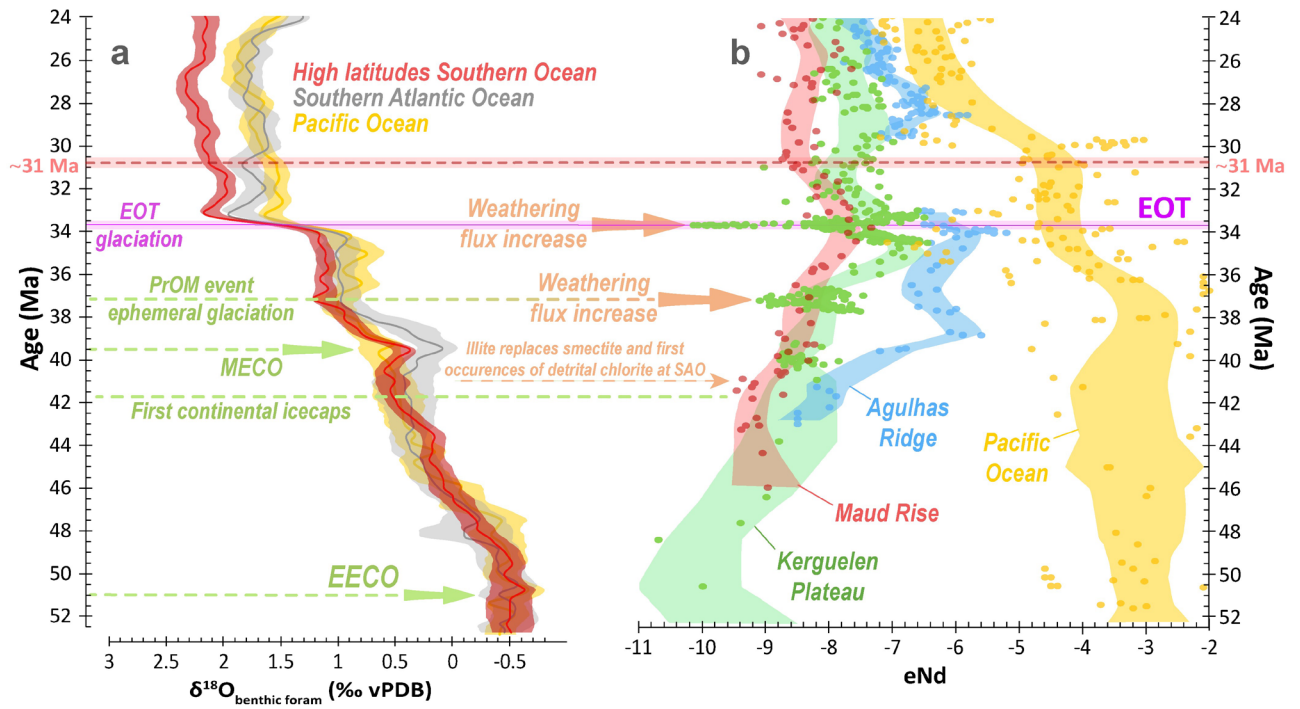
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950 *Figure 8*