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1 **Absence of a strong, deep-reaching Antarctic Circumpolar**
2 **Current zonal flow across the Tasmanian Gateway during**
3 **the Oligocene to early Miocene.**

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20

21 **ABSTRACT**

22 **The vigorous eastward flow of the Antarctic Circumpolar Current (ACC) connects all**
23 **major ocean basins and plays a prominent role in the transport of heat, carbon and**
24 **nutrients around the globe. However, the establishment of a deep circumpolar flow,**
25 **similar to the present-day ACC, remains controversial thereby obscuring our**

26 understanding of its climatic impact. Deciphering the chemical composition of
27 Circumpolar Deep Water (CDW) within the ACC can provide critical insights about its
28 development and evolution. Here we present new fossil fish teeth/bone debris neodymium
29 isotope (ϵ_{Nd}) records from Deep Sea Drilling Project (DSDP) Sites 278 and 274 in the
30 southwest Pacific Ocean, with the aim to trace changes in deep water masses across the
31 Tasmanian Gateway between the early Oligocene and early Miocene (~ 33-22 Ma). Site
32 274 provides the first Nd isotope record proximal to the Ross Sea during the Oligocene
33 (33.5-23.4 Ma). Its Nd isotope composition shows excursions to very radiogenic values,
34 $\epsilon_{Nd(t)} = -3.1$ and $\epsilon_{Nd(t)} = -3.7$, at 33.5 Ma and 23.8 Ma, respectively, in response to major
35 steps in Antarctic ice sheet expansion. A shift to lower, more unradiogenic $\epsilon_{Nd(t)}$ values
36 between 29.7 and 29.1Ma is linked to an increased influence of proto-CDW upwelling at
37 the site. In contrast, the Nd isotope record from Site 278 in the southern Emerald Basin
38 shows little variability ($\epsilon_{Nd(t)} = -6.0$ to -6.7) throughout the Oligocene and early Miocene
39 (30.9-21.8 Ma). Comparison with published data north of the ACC path, demonstrates
40 the presence of two deep water masses in the South Pacific prior to the inferred onset of
41 the ACC (33-30 Ma), one occupying depths between ~2500 and 3000 m ($\epsilon_{Nd(t)} = \sim -3$ to -5)
42 and a deep/bottom water mass (> 3000 m) with a more unradiogenic Nd isotope
43 composition ($\epsilon_{Nd(t)} = \sim -6$). Site 278 located close to the proto-polar front (proto-PF)
44 indicates that following the inferred onset of the ACC, deep waters bathing the southern
45 Emerald Basin remained more radiogenic in the Southwest Pacific compared to sites
46 along the proto-PF in the South Atlantic and Indian Ocean ($\epsilon_{Nd(t)} = \sim -8.1$). This indicates
47 a provinciality in Nd isotope compositions of deep waters along the proto-PF across the
48 Tasmanian Gateway. Our data are incompatible with the existence of a modern-like
49 homogenous (lateral and vertical) Nd isotope composition of CDW along the main flow
50 path of the ACC in all oceanic basins in the Oligocene to early Miocene. We attribute

51 **distinct Nd isotope compositions of deep waters across the Tasmanian Gateway to reflect**
52 **a less deep reaching and weaker ACC (proto-ACC) than today. Our findings suggest that**
53 **the modern strong and deep-reaching ACC flow must have been developed at a later**
54 **point in the Neogene.**

55

56 **Keywords:** ACC, CDW, neodymium isotope ratios, deep ocean circulation, Tasmanian
57 Gateway, Oligocene-early Miocene

58 **1. INTRODUCTION**

59 The Antarctic Circumpolar Current (ACC) is the world's largest and strongest ocean
60 current, transporting $\sim 136\text{-}162 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ of water along a $\sim 20,000$ km long path around
61 Antarctica (e.g., [Rintoul et al., 2014](#)). The absence of continental barriers around 60°S permits
62 the vigorous eastward flow of the ACC, which is driven by the strong westerly winds
63 (westerlies) and buoyancy forcing ([Rintoul, 2018](#)). Most of the circumpolar flow of the ACC
64 takes place along the Polar and the Subantarctic Fronts (PF and SAF, respectively), which
65 extend from the surface to the seafloor (i.e. deep-reaching), connecting the Atlantic, Pacific
66 and Indian Oceans, thus actively modulating the global ocean circulation ([Orsi et al., 1995](#);
67 [Sokolov and Rintoul, 2007](#); [Rintoul, 2018](#)). This interbasin connection of the ACC is a critical
68 feature of the modern global overturning circulation, which carries heat, carbon and nutrients
69 around the globe. It modulates the exchange of heat, moisture and carbon dioxide (CO_2)
70 between the ocean and the atmosphere and controls local and global marine primary
71 productivity ([Rintoul, 2018](#)).

72 Despite the critical role of today's strong deep-reaching ACC flow in the global ocean
73 circulation and Earth's climate, there is still an ongoing debate regarding the timing of its onset
74 and evolution towards a modern-like configuration throughout the Cenozoic period. Major

75 changes in paleogeography and seafloor paleobathymetry related to tectonic activity through
76 time, as well as changes in climate are likely to have controlled the initiation and evolution of
77 the ACC during the Cenozoic. For example, it is widely accepted that the ACC only initiated
78 after the opening of the two last major land barriers within its circum-Antarctic flow, the Drake
79 Passage and the Tasmanian Gateway, and their respective deepening to bathyal depths (>2000
80 m). Dating the time when the Drake Passage reached this depth is controversial, spanning from
81 the middle Eocene to the late Miocene (see [Dalziel, 2014](#) for discussion). In contrast, the
82 opening of the Tasmanian Gateway is better constrained ([Stickley et al, 2004](#)).

83 A Southern Ocean with closed tectonic gateways featured clockwise gyres in the South
84 Pacific and South Indian/Atlantic Ocean basins ([Huber et al., 2004](#)). Between ~49 Ma and 35.5
85 Ma, a shallow current flowing westward across the Tasmanian Gateway (from the Pacific to
86 the Indian Ocean) was proposed based on marine microfossils, organic geochemical records
87 and model simulations ([Huber et al., 2004](#); [Bijl et al., 2013](#)). Initial opening was followed by
88 intensified deepening of the seaway between ~35.5 and 30.2 Ma as indicated by the analysis
89 of sedimentological, micropaleontological and paleomagnetic data ([Stickley et al., 2004](#)).
90 Chemical water mass reconstructions suggest a westward flowing (i.e., from the Pacific to the
91 Indian Ocean) deep water current connecting both sides of the northern parts of the Tasmanian
92 Gateway between 32 and 30 Ma ([Scher et al., 2015](#)). A deep eastward flowing water connection
93 between the Indian and Pacific sectors of the Southern Ocean, was only established after 30
94 Ma, probably due to alignment of the northern part of the Tasmanian Gateway within the
95 latitudinal band of the westerlies ([Scher et al., 2015](#)). However, micropaleontological and
96 paleotemperature records suggest weak latitudinal paleotemperature gradients and warm
97 conditions prevailed in the Southern Ocean until at least the mid-Miocene ([Bijl et al., 2018](#);
98 [Hartman et al., 2018](#); [Salabarnada et al., 2018](#); [Sangiorgi et al., 2018](#)). Thus, the characteristics
99 of this nascent ACC remain poorly understood.

100 One possible way to reconstruct past connectivity along the ACC is to track the
101 evolution of Circumpolar Deep Water (CDW), using neodymium (Nd) isotope compositions
102 ($^{143}\text{Nd}/^{144}\text{Nd}$ ratio) of ambient seawater in fossil fish teeth and bones (e.g., [Wright et al., 2018](#)).
103 Circumpolar Deep Water is the most voluminous water mass in the Southern Ocean, occupying
104 the shallow to deep layers of the ACC in the Southern Ocean ([Orsi et al., 1995](#)). Today, CDW
105 has a vertically and horizontally homogenous Nd isotope composition, probably due to the
106 strong vertical and horizontal mixing along the frontal systems (e.g., [Stichel et al., 2012](#);
107 [Lambelet et al., 2018](#)). These findings indicate that the evolution of the Nd isotope composition
108 of CDW in the past may provide important insights into interbasinal connections/circulation in
109 the Southern Ocean.

110 Neodymium is incorporated into fish teeth at the sediment-water interface at the time
111 of deposition, remineralisation and burial ([Martin and Haley, 2000](#)). Post-burial alteration is
112 generally negligible relative to the early diagenetic rare earth element (REE) uptake, allowing
113 retention of the Nd isotope composition of bottom water signals (e.g., [Martin and Haley, 2000](#)).
114 Neodymium is a lithogenic element, introduced to the ocean at its interfaces with the solid
115 earth (e.g., [van Flierdt et al., 2016](#)). Water masses forming in the different oceanic basins are
116 distinct in their Nd isotope fingerprint, due to the distribution of geological ages and lithologies
117 around the globe, with younger volcanogenic material yielding more radiogenic (higher) Nd
118 isotope compositions, and older cratonic rocks being characterised by less radiogenic (lower)
119 Nd isotope compositions. ϵ_{Nd} denotes the deviation of a measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratio from the
120 chondritic uniform reservoir in parts per 10,000 ([Jacobsen and Wasserburg, 1980](#)). Because
121 seawater Nd has a relatively short residence time in the ocean (~500 to 1,000 years) ([Tachikawa](#)
122 [et al., 2003](#)), Nd can be carried from one ocean basin to another, and hence trace the provenance
123 of water masses.

124 Here we present new Oligocene to lower Miocene neodymium isotope data obtained
125 from fossil fish teeth and bones (fish debris) in sediments recovered by the Deep Sea Drilling
126 Project (DSDP) at Sites 278 and 274. Site 278 is strategically located within the main path of
127 the modern ACC in the southern Emerald Basin (Kennett et al., 1975; Rintoul et al., 2014),
128 while Site 274 is situated proximal to the Ross Sea in the Adare Basin (Figs. 1A; 1B; S1)
129 (Hayes et al., 1975). Our new data provide critical insights into the chemical evolution of CDW
130 in the Southwest Pacific in the time period following the initial (northern) opening of the
131 Tasmanian Gateway and leading up to the development of the modern-like ACC.

132 **2. Modern oceanographic setting of drill sites DSDP 274 and 278**

133 To study the history of the ACC, and particularly the chemical composition of deep
134 waters through the Oligocene to early Miocene, we used sediments recovered from Hole 278
135 (Kennett et al., 1975). Site 278 (56°33.42'S, 160°04.29'E, 3675 m water depth) is located on a
136 pelagic contourite drift in the southern Emerald Basin (Figs. 1A; 1B) (Kennett et al., 1975). At
137 present, the site is located close to the northern branch of the Polar Front (PF) and within the
138 main pathway of the ACC (Sokolov and Rintoul, 2007; Rintoul et al., 2014). As the ACC enters
139 the Southwest Pacific through the Tasmanian Gateway, its flow is constrained by the
140 Macquarie Ridge, which extends from depths of 5000 m to an average depth of 1500 m between
141 47°S and 56°S (Rintoul et al., 2014). The ACC passes over and through narrow gaps of the
142 Macquarie Ridge and south of the ridge crest around 56°S, before reaching the southern
143 Emerald Basin (Rintoul et al., 2014). From there, the ACC flows around the southern edge of
144 the Campbell Plateau and to the Bounty Trough, reinforcing the Deep Western Boundary
145 Current (DWBC). South of the Bounty Trough (46°S) the ACC turns to the east and continues
146 its journey across the South Pacific, while the DWBC continues north towards the equator in
147 the Pacific Ocean basin (Carter and McCave 1997). The ACC carries CDW eastwards from

148 the Southeast Indian Ocean into the Southwest Pacific Ocean. In the modern southern Emerald
149 Basin, CDW extends to the seafloor, bathing Site 278 (Figs. 1B; S1-S2) (Rintoul et al., 2014;
150 Lambelet et al., 2018).

151 Site 274 ($68^{\circ}59.81'S$, $173^{\circ}25.64'E$, 3305 m water depth) was drilled on the lower
152 continental rise, 250 km north-northeast of Cape Adare (Hayes et al., 1975) (Fig. 1B). It is
153 currently situated south of the Southern ACC Front (SACCF) (Figs. 1B; S1) (Sokolov and
154 Rintoul, 2007) and within the main outflow path of bottom water formed in the western Ross
155 Sea (Gordon et al., 2009). At a depth of 3305 m, the main modern water mass encountered is
156 Ross Sea Bottom Water (RSBW) (Fig. 1B; S1). Today, most of the dense shelf water on the
157 Ross Sea shelf resides in the Joides and Drygalski troughs, from where it is exported as RSBW
158 to the abyssal plains (Fig. 1B) (Gordon et al., 2009). In contrast to CDW, which can be found
159 all around Antarctica, bottom topography restricts the pathway of the Antarctic Bottom Water
160 (AABW), including RSBW, in the Southern Ocean (Orsi et al., 1995). Only a small portion of
161 RSBW formed in the western Ross Sea flows northward from its outlets, with most of it being
162 diverted westward against the continental slope into the deep Australian-Antarctic Basin
163 (Gordon et al., 2009). Through diapycnal mixing with overlying CDW, the hydrographic
164 properties of RSBW quickly erode along its westward pathway and are only detectable to
165 $\sim 140^{\circ}E$ along the Adélie-Wilkes Coast (Rintoul, 1998). In contrast, there is a larger fraction of
166 RSBW detected north of the Ross Sea, but this variety of RSBW forms predominantly in the
167 central Ross Sea and is channelled to the continental slope by the Glomar Challenger trough
168 and into the Hillary Canyon. Much of this bottom water enters the southeast Pacific Basin (Fig.
169 1B) (Gordon et al., 2009).

170

FIGURE 1

171 **2.1 Modern Nd isotope signatures of deep and bottom waters around the Tasmanian**
172 **Gateway**

173 Globally, North Atlantic Deep Water (NADW) and Pacific Deep Water (PDW)
174 constitute water mass endmembers in terms of their Nd isotope composition. Modern NADW
175 is characterised by relatively negative ϵ_{Nd} values ($\epsilon_{Nd} \sim -12$ to 13) (e.g., [Lambelet et al., 2016](#)),
176 resulting from the weathering of old cratonic rocks surrounding the North Atlantic. Pacific
177 Deep Water, in contrast, carries less negative (i.e. more radiogenic) ϵ_{Nd} values of ~ -4 (e.g.,
178 [Amakawa et al., 2009](#)), due to the influence of weathering of younger volcanogenic material
179 around the North Pacific. Modern CDW in the Southern Ocean is characterised by a relatively
180 homogenous Nd isotope composition along the flow path of the ACC across all ocean basins
181 and with depth ($\epsilon_{Nd} = -8.3 \pm 1.5$; $n = 158$) ([Figs. 2A](#)) ([Lambelet et al., 2018](#) and references
182 therein). In the western part of the Tasmanian Gateway along the $\sim 140^\circ E$ meridian, CDW
183 exhibits ϵ_{Nd} values of -8.6 ± 0.4 ($n=4$) around 2500-4300 m ([Lambelet et al., 2018](#)). In the
184 eastern part of the gateway, in the Macquarie Ridge region, close to Site 278, CDW has ϵ_{Nd}
185 values of -9.0 ± 0.2 and -9.2 ± 0.2 between 3201 and 4183 m, respectively ([Figs. 2A; S2](#))
186 ([Lambelet et al., 2018](#)). Ross Sea Bottom Water exhibits a more radiogenic Nd isotope
187 signature than CDW, probably acquired from Ross Sea shelf sediments and mixing with shelf
188 waters ($\epsilon_{Nd} = -6.5$ to -7.5 ; [Rickli et al., 2014](#); [Basak et al., 2015](#)) ([Fig. 2B](#)).

189 **FIGURE 2**

190 **3. MATERIALS AND METHODS**

191 **3.1 Study sites, age models, and paleodepths**

192 Our study focuses on the lower part of Hole 278 at DSDP Hole 278 (cores 34-26, 429-
193 329 meters below sea floor (mbsf)). Sediments from this interval consist of (i) siliceous
194 nannofossil chalk (429-386 mbsf), and (ii) detrital and nannofossil-bearing radiolarian-diatom
195 ooze, alternating with siliceous oozes (386-329 mbsf) (Kennett et al., 1975). A new age model
196 for the studied sections of Site 278 has been established based on the integration of new
197 magnetostratigraphic data with revised diatom, radiolarian and calcareous nannofossil
198 biostratigraphy, calibrated to the Geological Time Scale (GPTS) 2012 (Gradstein et al., 2012)
199 (Fig. 3; Supplementary Text; Tables S1-S4).

200 For DSDP Site 274, we focus on two distinct intervals (408.5-313.5 mbsf and 199.5-
201 180.5 mbsf). Shipboard core descriptions report that sediments between 408.5 and 328 mbsf
202 are composed of silty claystones with local presence of chert layers. Sediments between 323
203 and 180.5 mbsf consist of diatom-detrital silty clay and minor silty clay diatom ooze (Hayes et
204 al., 1975). Based on the initial age model, sediments between 408.5 and 180.5 mbsf are early
205 to late Oligocene in age (Hayes et al., 1975). New dinocyst constraints combined with new
206 magnetostratigraphic data and revised diatom biostratigraphy (updated from Hayes et al.
207 (1975)) calibrated to GPTS 2012 (Gradstein et al., 2012) provide an updated age model for the
208 studied sediments at Site 274 from ~33.7 to 23.3 Ma (Fig. S4) (Hoem et al., 2021).

209 We reconstructed the paleoposition of Sites 278 and 274 using the G-plates geodynamic
210 modeling (<http://www.gplates.org>; Müller et al., 2018), utilizing the plate circuit of Müller et
211 al. (2016). Paleodepths for the sites were obtained from the paleo-bathymetry grids generated
212 by the Earthbyte group (Scotese and Wright, 2018), incorporating the reconstructed
213 paleolocation of the two sites (Fig. S5). Our reconstructions show that Site 278 migrated
214 northwards from ~62.84°S (~30 Ma) to 61.4°S (~23 Ma); its paleodepth varied between 3500
215 and 4000 m throughout the Oligocene and early Miocene. Site 274 has migrated north from a
216 position of ~70.22°S in the early Oligocene (~33 Ma) to ~69.57°S at the end of the Oligocene-

217 Miocene transition (~23 Ma) and remained at ~2500-3000 m water depth throughout the
218 Oligocene.

219 **3.2 Neodymium isotope analyses**

220 Fish teeth and bones (hereafter termed fish debris) were handpicked from the > 63 µm
221 sediment fractions isolated by wet sieving. A total number of 32 samples were selected for fish
222 debris Nd isotope analyses (20 samples from Site 278 and 12 samples from Site 274) (**Table**
223 **S5**). All samples were treated with ultraclean 18 MΩ water (Milli-Q water) and methanol to
224 remove debris from surfaces and cavities following [Martin and Haley \(2000\)](#) in the MAGIC
225 laboratories at Imperial College London (see also [Huck et al., 2017](#)). Cleaned fish debris
226 samples were subsequently transferred into cleaned microcentrifuge tubes and dissolved
227 overnight in 50 µL of 2M HCl. Dissolved fish debris were loaded on Biorad cation exchange
228 resin (200-400 µm mesh) to separate the REEs from the sample matrix and Eichrom Ln-Spec
229 resin (50-100 µm bead size) to separate Nd from the other REEs.

230 Furthermore, seven samples were selected (two from Site 278 and five from Site 274;
231 **Table S5**) to determine the detrital Nd isotope composition in order to evaluate potential
232 contribution of the detrital sediments to the porewaters or overlying bottom water signature.
233 Samples were dried and gently homogenised using mortar and pestle. Approximately 500 mg
234 of homogenised material was subjected to a carbonate leaching procedure to remove biogenic
235 carbonate using 30 ml of 1.5% buffered acetic acid (modified from [Biscaye \(1965\)](#)). Due to
236 the significant carbonate content of the samples from Site 278 (38 to 43 %), approximately 1.5
237 g of dried sample and 70-75 ml of 1.5% buffered acetic acid were used. Exchangeable ions
238 were subsequently removed using 10 ml 1 M MgCl₂ solution. In a third step, ferromanganese
239 oxides and oxyhydroxides were removed using a weak reductive solution of 0.005 M
240 Hydroxylamine Hydrochloride (NH₂OH) 1.5% acetic acid and 0.03 M Na₂-EDTA for one

241 hour, followed by a stronger leaching step utilising 0.05 M NH₂OH for 17 hours. 50 mg of
242 leached and water washed detrital sediment was subsequently dried, weighted and digested on
243 a hotplate using a mixture of 1 ml of concentrated HNO₃, 0.8 ml HClO₄ and 2 ml HF. The
244 detrital samples were processed using the same ion chromatography as the fish debris samples.

245 Neodymium isotope ratios of fish debris and detrital sediment samples were determined
246 on a high-resolution Nu Plasma multiple collector inductively coupled plasma mass
247 spectrometer (MC-ICP-MS) at Imperial College London, operated in static mode. Instrumental
248 mass bias was corrected for using a ¹⁴⁶Nd/¹⁴⁴Nd ratio of 0.7219. All reported ¹⁴³Nd/¹⁴⁴Nd ratios
249 are corrected to a nominal JNd_i value of 0.512115 (Tanaka et al., 2000) using bracketing
250 standards. JNd_i standards were also used to monitor external reproducibility, and accuracy was
251 evaluated by processing USGS BCR-2 rock standards alongside samples, which yielded
252 average ¹⁴³Nd/¹⁴⁴Nd ratios of 0.512636 ± 0.000008 (n=23; 2s.d.), in agreement with the
253 published BCR-2 ¹⁴³Nd/¹⁴⁴Nd ratio of 0.512638 ± 0.000015 (Weis et al., 2006) (Table S5).

254 3.3 Rare earth element analyses

255 To correct for the decay of ¹⁴⁷Sm to ¹⁴⁴Nd within the fish debris over time, we used the
256 average Sm and Nd concentrations measured for two fish debris samples from Site 278
257 (¹⁴⁷Sm/¹⁴⁴Nd ratios of 0.1275 and 0.1256; ¹⁴⁷Sm/¹⁴⁴Nd_{average} = 0.1266) and three fish debris
258 samples from Site 274 (¹⁴⁷Sm/¹⁴⁴Nd ratios of 0.1532, 0.1921, and 0.2662; ¹⁴⁷Sm/¹⁴⁴Nd_{average} =
259 0.2038). The range of Sm/Nd ratios reported here are consistent with values from Oligocene to
260 Miocene fish debris material elsewhere (Huck et al., 2017; Wright et al., 2018; Evangelinos et
261 al., 2020). Corrections for in situ decay of ¹⁴⁷Sm amounted to 0.19 to 0.28 ε_{Nd} units for Site 278
262 and 0.02 to 0.14 ε_{Nd} units for Site 274; (t) denotes age-corrected samples (Table S5).

263 To investigate whether the Nd in the fish debris is of authigenic nature (i.e. seawater-
264 derived), the full suite of REE concentrations were determined on one fish debris sample from

265 Site 278 and three fish debris samples from Site 274 (Table S6). Rare earth element analysis
266 was performed at the Open University using an Agilent Technologies 8800 Triple-Quad
267 Inductively Coupled Plasma-Mass Spectrometer (ICP-MS). Analyses were standardized via a
268 suite of seven synthetic multi-element standards made up with certified plasma standard
269 solutions. An intermediate synthetic multi-element monitor standard, a fossil bone standard
270 (Chavagnac et al. 2007) and a 2% HNO₃ blank were run every 5th sample to monitor instrument
271 drift and precision. Precision was generally better than ±2% (1 s.d.). Both oxide interferences
272 (CeO⁺/CeO⁻) (<0.5%) and doubly charged species (Ce⁺⁺/Ce⁺) (<1.2%) were kept low. All REE
273 data were normalised to Post Archean Shale (PAAS) concentrations (Taylor and McLennan,
274 1985).

275 4. RESULTS

276 4.1 Age model DSDP Site 278

277 An age-depth model for the Oligocene and lower Miocene intervals of DSDP Site 278
278 is developed from the integration of new magnetostratigraphic data, calcareous nannofossil,
279 marine diatom, and radiolarian biostratigraphy, calibrated using the Geological Time Scale
280 (GPTS) 2012 (Gradstein et al., 2012) (Figs. 3; Supplementary Text; Tables S1-S4). We also
281 present an alternate age model that fit with the constraints of the data (for details see
282 supplementary material). The stratigraphic section cored at Site 278 from the bottom of the
283 sedimentary section in core 34 at ~429 mbsf up to sample 278-21R-1, 50 cm at ~282 mbsf
284 spans the time interval from ~ 31 to 17.7 Ma, with one unconformity noted between samples
285 278-31R-3, 35 cm and 278-31R-2, 139 cm (398.85 – 399.89 mbf), which removed a ~2.4 m.y.
286 time interval from 28.2 to 25.8 Ma. The top of this interval is bounded by an unconformity
287 between samples 278-21R-1, 50 cm and 278-20R-6, 50 cm (281.5 mbsf), which removed the
288 time interval 17.6 to 15.5 Ma (Fig. 3). Studied sediments (cores 34R-26R) are dated between

289 ~31 and 21.7 Ma, similar to the initial shipboard ages (Fig. 3), suggesting upper Oligocene to
290 lower Miocene sediments (Kennett et al., 1975).

291

FIGURE 3

292 4.2 Fish debris Nd isotope compositions

293 Fish debris Nd isotope data from Site 278 show little variability, ranging from $\epsilon_{Nd(t)} = -$
294 6.0 ± 0.2 to 6.7 ± 0.2 from the early Oligocene to the early Miocene (~ 30.9-21.8 Ma) (Fig. 4;
295 Fig. S6A; Table S5). The average $\epsilon_{Nd(t)}$ throughout the record is -6.4 ± 0.2 (n=20). In contrast,
296 fish debris Nd isotope data from Site 274 show more variability, and reach from maximum
297 values of -3.1 ± 0.3 at 33.5 Ma and -3.7 ± 0.3 at 23.8 Ma to relative minima of -5.0 ± 0.1 , at
298 ~33.2-29.7 Ma, -6.2 ± 0 at ~29 Ma, -7.3 ± 0.3 and -6.8 ± 0.2 at ~24.3 and 24.0 Ma, respectively
299 (Fig. 4; Fig. S6B; Table S5). The late Oligocene time period features a pronounced excursion
300 to more radiogenic $\epsilon_{Nd(t)}$ values by ~3 epsilon units between 24.0 and 23.8 Ma.

301 4.3 Bulk sediment Nd isotope compositions

302 The detrital samples at Site 278 exhibit $\epsilon_{Nd(t)}$ values of -10.4 ± 0.3 (~ 30.5 Ma) and -7.9
303 ± 0.3 (~ 21.8 Ma), respectively (Fig. 4; Fig. S6A; Table S5). Lower Oligocene detrital sediment
304 Nd isotope compositions from Site 274 exhibit $\epsilon_{Nd(t)}$ values of -4.2 ± 0.3 (~33.5 Ma), $-7.6 \pm$
305 0.3 (~33.2 Ma) and -8.5 ± 0.3 (~29 Ma). Two additional late Oligocene detrital sediment
306 samples yielded $\epsilon_{Nd(t)}$ values of -10.3 ± 0.3 and -8.5 ± 0.3 at ~24.0 Ma and 23.8 Ma, respectively
307 (Fig. 4; Fig. S6B; Table S5).

308 4.4 Fish debris REE patterns

309 Five fish debris samples from Site 278 and Site 274 yielded middle-REE enriched
310 patterns (Fig. 5; Table S6). Only one sample from section 278-33R-6 showed a negative Ce

311 anomaly, which together with the middle-REE enriched patterns represent a diagnostic feature
312 of seawater-derived Nd in fish teeth samples (Scher et al., 2011).

313 **FIGURE 4**

314 **FIGURE 5**

315 **5. DISCUSSION**

316 **5.1 Reliability of Nd isotopes from Sites 278 and 274**

317 Results of Oligocene to Miocene-aged fish debris samples at Site 278 (paleodepth:
318 3500-4000 m) are more radiogenic (less negative) in their Nd isotope composition ($\epsilon_{Nd(t)} = -6.0$
319 to -6.7) than modern CDW in the south Emerald Basin ($\epsilon_{Nd} = \sim -9$) at ~ 3200 - 4200 m water
320 depth (Lambelet et al., 2018). Results for Site 274 ($\epsilon_{Nd(t)} = -3.1$ to -7.3) show generally more
321 radiogenic values compared to ambient RSBW ($\epsilon_{Nd} = -6.5$ to -7.5 ; Rickli et al., 2014; Basak et
322 al., 2015) (Fig. 4). The first question to address is if the fish debris record a purely authigenic
323 signal.

324 Site 278 is located near the Macquarie Ridge, a relict volcanic mid-ocean ridge,
325 characterised by basalts and peridotites with an ϵ_{Nd} value of $+7$ to $+11$ (Dijkstra et al., 2009;
326 Conway et al., 2012). Interaction of volcanic sediments and/or hydrothermal products with
327 ambient seawater could have altered the ϵ_{Nd} values at Site 278 towards more radiogenic values.
328 However, hydrothermal activity has been shown to be a sink rather than a source of Nd near
329 ridges (Stichel et al., 2018). Moreover, fish teeth records from the Kerguelen Plateau show no
330 changes in their Nd isotopic composition associated with the local volcanic activity during the
331 timing of the northern Kerguelen Plateau emplacement (Wright et al., 2018). Volcanic glass
332 and rock fragments have been reported from close to the base of siliceous nannofossil chalk at

333 Site 278 (Kennett et al., 1975) (~31-28.2 Ma, based on our new age model). However, the lack
334 of any significant variability in our fish debris Nd record makes it difficult to reconcile any
335 major volcanic contribution to the porewater and hence to the fish debris Nd isotope signature
336 (Figs. 4; S7A).

337 To further evaluate the potential interaction between detrital sediments and pore
338 water/seawater Nd isotope composition, we can use the results from two detrital sediment
339 samples from the siliceous chalk unit (section 33R-6) and from the nannofossil-bearing diatom
340 ooze unit (section 26R-2) at Site 278. The detrital samples exhibit $\epsilon_{Nd(t)}$ values of -10.4 ± 0.3
341 and -7.9 ± 0.3 , respectively (Fig. 4; Table S5), within the range of terrigenous inputs from
342 Paleozoic granites and metasedimentary rocks ($\epsilon_{Nd(t)} = -4.9$ to -13.9) of the South Island, New
343 Zealand (Pickett and Wasserburg 1989). This observation is in line with the general deep water
344 flow in the area, suggesting that terrigenous sediments in the Southern Emerald Basin are
345 predominantly derived from the erosion of the New Zealand's Southern Alps via the Solander
346 Channel (Carter and McCave, 1997). Importantly, the measured detrital sediment composition
347 is more unradiogenic (i.e. more negative ϵ_{Nd} values) than the almost contemporaneous fish
348 debris signatures and hence cannot be responsible for the observed more radiogenic values. In
349 addition, REE patterns from a fish debris sample from section 33R-6 (~30.5 Ma) exhibit a
350 pattern enriched in middle-REEs and a prominent negative cerium (Ce) anomaly, which
351 represent diagnostic features of uncontaminated seawater origin in the fish debris (Fig. 5; Table
352 S6) (Scher et al., 2011; Huck et al., 2016, 2017). The fish debris sample from section 26R-2
353 (21.8 Ma) exhibit also a typical middle-REEs enriched pattern but records a slightly positive
354 Ce anomaly (Fig. 5). Positive Ce anomalies in fish debris have been linked to anoxic seawater
355 and/or pore water conditions, resulting in remobilization of REEs from authigenic or organic
356 matter coatings (Elderfield and Pagett, 1986; Freslon et al., 2014; Huck et al., 2016). High

357 biogenic productivity at Site 278 associated with proto-polar front upwelling processes
358 (Kennett et al., 1975) may have led to anoxic conditions at the seafloor.

359 Similar to Site 278, bulk sediment results from Site 274 exhibit a less radiogenic (i.e.
360 lower) Nd isotope fingerprint than fish debris when comparing results for the same samples
361 and samples taken very close to each other. The difference in $\epsilon_{Nd(t)}$ for three pairs of fish debris-
362 detrital sediment samples analyzed is -4.9, -3.5 and -2.4 ϵ_{Nd} units, while adjacent samples show
363 differences in $\epsilon_{Nd(t)}$ of -2.5 and -1.1 (Figs. 4; S7B; Table S5). This pronounced difference
364 between the $\epsilon_{Nd(t)}$ values of the fish debris and the detrital sediment samples at Site 274 strongly
365 supports the absence of significant exchange between the detrital sediments and the fish debris.
366 Overall, the more radiogenic $\epsilon_{Nd(t)}$ values of the fish debris, compared to the more unradiogenic
367 $\epsilon_{Nd(t)}$ values of the detrital samples, argues against a contribution of Nd from the detrital
368 sediments.

369 Furthermore, REE patterns of three fish debris samples at Site 274 yielded middle-REE
370 bulge patterns, including a slightly positive Ce anomaly (Figs. 5; Table S6). The close
371 proximity of Site 274 to the Antarctic continental margin and the high biogenic fraction in the
372 sediment (Hayes et al., 1975) may have caused the observed positive Ce anomaly resulting
373 from increased weathering inputs and/or remobilized REEs from authigenic or organic matter
374 coatings due to anoxic conditions in the pore waters (Elderfield and Pagett, 1986; Freslon et
375 al., 2014; Huck et al., 2016). Further work is required to fully understand the positive Ce
376 anomaly in our fish debris samples. Nevertheless, REE patterns from Site 278 and 274 are
377 similar to REE patterns from Eocene-Oligocene fish debris records from around the Tasmanian
378 Gateway, which have been interpreted to preserve a predominantly seawater signature, despite
379 positive Ce anomalies (Huck et al., 2016, 2017). We therefore suggest an authigenic nature of
380 the fish debris record from Sites 278 and 274 and use our new data to explore the chemical

381 fingerprint of proto-CDW and RSBW across the Tasmanian Gateway, and its implications for
382 the ACC evolution.

383 **5.2 Early Oligocene epoch (33 -30 Ma): Prior to the inferred onset of the ACC** 384 **throughflow via the Tasmanian Gateway**

385 Neodymium isotope values from our deeper Site 278 (paleodepth: ~4000 m) between
386 31 and 30 Ma, indicate that the southern Emerald Basin was bathed by an unradiogenic deep
387 water mass ($\epsilon_{Nd(t)} = \sim -6.4$) (Figs. 4; 6). These values converge with published Nd isotope
388 records from the deep South Pacific (Site U1370, paleodepth ~5000 m, Site 596 paleodepth
389 ~5000 m and Site 323, paleodepth ~4000 m; $\epsilon_{Nd(t)} = \sim -6.3$) (Thomas et al. 2014; McKinley et
390 al., 2019), suggesting that abyssal sites in the South Pacific were likely influenced by a
391 common deep/bottom water mass, such as South Pacific Deep Water (SPDW) (Fig. 6).
392 Temporal and spatial Nd isotope patterns and modelling results show that SPDW was a
393 prominent and persistent water mass of the deep (≥ 3500 m) South Pacific (between at least
394 $\sim 30^\circ\text{S}$ - 63°S) from 70 to 20 Ma, characterised by $\epsilon_{Nd(t)}$ of ~ -6 (Thomas et al. 2014; McKinley
395 et al., 2019; Sarkar et al., 2019). We therefore suggest that Site 278 (paleolocation: $\sim 63^\circ\text{S}$) was
396 bathed by SPDW between 33 and 30 Ma.

397 Neodymium isotope records consistent with model results indicate bimodal mode of
398 deep ocean circulation in the Pacific Ocean from ~ 70 to 25 Ma (Scher, 2014; Thomas et al.,
399 2014; Scher et al., 2015; Huck et al., 2017; McKinley et al., 2019; Sarkar et al., 2019).
400 Convection of SPDW occurred in the South Pacific (likely in the Ross Sea region), while North
401 Pacific Deep Water (NPDW) formation occurred in the North Pacific. Our $\epsilon_{Nd(t)}$ values from
402 Site 274, located proximal to the Ross Sea, however, range from -3.1 to -5 between 33.5 and
403 29.7 Ma and albeit of low resolution are more radiogenic (less negative), compared to both the
404 present-day RSBW Nd isotopic composition ($\epsilon_{Nd} = -6.4$ to -7.5 ; Rickli et al., 2014; Basak et al.,

405 2015) and the deep record from Site 278 (Figs. 4; 6). The average $\epsilon_{Nd(t)}$ value of ~ -4.6 for the
406 early Oligocene (~ 33.5 - 29.7 Ma) at Site 274 (paleodepth: ~ 2500 - 3000 m) overlaps with
407 published data from the East Tasman Plateau (Site 1172; average $\epsilon_{Nd(t)} = \sim -4.2$ paleodepth: \sim
408 2400 m) and Hikurangi Plateau (Site 1124; average $\epsilon_{Nd(t)} = \sim -4.6$; paleodepth: ~ 3000 m) (Scher
409 et al., 2015; Sarkar et al., 2019) (Fig. 6). Similarities in Nd isotope values and paleodepths
410 between the Antarctic margin record from Site 274 and the published records for the northern
411 parts of the Tasmanian Gateway (Sites 1172 and 1124; Scher et al. 2015), suggest that these
412 sites may have been influenced by a common water mass, occupying water depths between \sim
413 2500 and 3000 m. This more radiogenic signature is akin to northern sources Equatorial Pacific
414 Deep Water (Fig. 6) (Scher, 2014), which originated in the deep Equatorial Pacific, where
415 NPDW and SPDW mixed (Thomas et al., 2014). Neodymium isotope data from Sites 1172 and
416 1124 suggest that Equatorial Pacific water was present at 2500 - 3000 m, reaching at least to
417 southern latitudes of $60^{\circ}S$ - $65^{\circ}S$ between 36 and 30 Ma (Scher et al., 2015; Sarkar et al., 2019).
418 We here suggest that water from a northern source with a more Equatorial Pacific-like Nd
419 isotope signature reached to the Adare Basin (Site 274, $\sim 70^{\circ}S$) between 33 and 30 Ma.

420 Our results point to the presence of two distinct deep water masses in the late
421 Eocene/early Oligocene South Pacific: one occupying depths between ~ 2500 and 3000 m with
422 Nd composition ($\epsilon_{Nd(t)} = \sim -3$ to -5) and a second deep/bottom water mass (> 3000 m) with more
423 unradiogenic Nd composition ($\epsilon_{Nd(t)} = \sim -6$) (Fig. 6).

424 **FIGURE 6**

425 **5.3 Early Oligocene to early Miocene epoch (30-22 Ma): After the inferred onset of the** 426 **ACC throughflow via the Tasmanian Gateway**

427 Neodymium isotope values at Site 274 show a shift towards less radiogenic values
428 ($\epsilon_{\text{Nd}(t)} = -6.2 \pm 0.3$) between 29.7 and 29.1 Ma. Slightly less radiogenic $\epsilon_{\text{Nd}(t)}$ values are observed
429 around 24.3 Ma and 24.0 Ma (-7.3 ± 0.25 and -6.8 ± 0.15 , respectively) (Figs. 4; 6), resembling
430 modern RSBW values ($\epsilon_{\text{Nd}} = -6.4$ to -7.5 ; Rickli et al., 2014; Basak et al., 2015). The shift
431 between 29.7 and 29.1 Ma at Site 274 coincides with the shift observed in the two Nd isotope
432 records from the East Tasman Plateau (Site 1172 $\epsilon_{\text{Nd}(t)} = -4.8$ to -6.8) and the Hikurangi Plateau
433 (Site 1124 $\epsilon_{\text{Nd}(t)} = -5.6$ to -6.2) between 29.6 and 29 Ma, respectively (Scher et al., 2015) (Fig.
434 6). This latter shift has been linked to the northward migration and alignment of the Tasmanian
435 Gateway with the westerly winds (Scher et al., 2015). The change in the $\epsilon_{\text{Nd}(t)}$ values at Site
436 274 between 29.7 and 29.1 Ma does not coincide with any significant known change in the
437 Antarctic ice sheet that would have influenced the sediment input around the Ross Sea. Instead,
438 given that Site 274 was located south of the proto-PF during this period (Nelson and Cooke,
439 2001; Scher et al., 2015), we posit that the shift to more unradiogenic Nd contribution is
440 probably due to the onset of proto-CDW upwelling reaching the Adare Basin and Site 274.

441 Site 278, on the other hand, was located close to the proto-PF and proto-SAF between
442 30 Ma and 22 Ma and thus in the main pathway of the proto-ACC and proto-CDW (Nelson
443 and Cooke, 2001), explaining its invariant Nd isotope fingerprint (average $\epsilon_{\text{Nd}(t)} = -6.4$)
444 throughout the early Oligocene to early Miocene (29.7-22 Ma) (Figs. 4; 6). These values are
445 matched by the two published Nd isotope records (Sites 1172 and 1124; $\epsilon_{\text{Nd}(t)} \sim -6.5$) from the
446 eastern side of Tasmanian Gateway after the onset of ACC influence at this northern location
447 (Scher et al., 2015) (Fig. 5). All three records show more radiogenic Nd isotope compositions
448 than modern CDW in the Emerald Basin ($\epsilon_{\text{Nd}} = -9$; Lambelet et al., 2018) (Figs. 2A; S2). They
449 furthermore differ from records to the west of the Tasmanian Gateway. Throughout the
450 Oligocene (30-23 Ma) Nd isotope records along the proto-PF in the South Atlantic and the
451 South Indian Ocean are characterised by proto-CDW with a Nd isotope composition between

452 ~ -7 and -9 (Site 689: [Scher and Martin, 2004](#); Site 748 and 744: [Wright et al., 2018](#); Site 269:
453 [Evangelinos et al., 2020](#)) (**Figs. 6; 7**). Such values are similar to modern CDW ($\epsilon_{Nd} = -8.3 \pm 1.5$,
454 $n=158$) ([Lambelet et al., 2018](#)) (**Fig. 6**).

455 Comparison between our new and previously published data along the proto-ACC
456 pathway around the Tasmanian Gateway therefore reveals a provinciality in seawater Nd
457 isotopes with more radiogenic proto-CDW ($\epsilon_{Nd(t)} = -6$ to -7.5) on the Pacific side of the
458 Tasmanian Gateway, and less radiogenic proto-CDW ($\epsilon_{Nd(t)} = -7$ to -9) in the Indian and Atlantic
459 sectors of the Southern Ocean (**Figs. 6; 7**). This observation implies an absence of a
460 homogenous deep water mass across the Tasmanian Gateway during the Oligocene and early
461 Miocene, and suggests a more complex evolution of the deep ocean circulation across the
462 gateway than previously assumed.

463 We propose that the provinciality in water mass signatures across the gateway was due
464 to a less deep reaching and probably also weaker proto-ACC flow during the Oligocene to early
465 Miocene (**Fig. 7**). A shallower and weaker ACC flow entering the South Pacific would have
466 increased the amount of time that Pacific deep water could have mixed with the proto-CDW,
467 allowing for acquisition of a more radiogenic isotopic composition of proto-CDW. These
468 inferences are consistent with numerical simulations showing limited throughflow of the proto-
469 ACC due to Australasian paleogeography during the Oligocene ([Hill et al., 2013](#)) and weaker
470 global overturning circulation due to weaker westerly winds from the early to mid-Miocene
471 ([Herold et al, 2012](#)).

472 Recently collected sedimentological data, microfossil assemblages and past
473 reconstructions of sea surface temperatures from the Antarctic-Australian Basin and around
474 the Tasmanian Gateway, also infer a weaker proto-ACC frontal system, characterised by
475 intrusions of warm waters from northern latitudes from the Oligocene to the middle Miocene,
476 weak latitudinal paleotemperature gradients and absence of strong sea ice seasonality (34-11

477 Ma) (Bijl et al., 2018; Hartman et al., 2018; Salabarnada et al., 2018; Sangiorgi et al., 2018;
478 Evangelinos et al., 2020; Hoem et al., 2021). The question remains when the modern-like ACC
479 was established. Our data suggest that this occurred during the Neogene, but future work is
480 necessary to address the timing of its establishment and the wider implications of a weaker
481 Oligocene to Miocene ACC.

482 **FIGURE 7**

483 **5.4 Evidence of a glacial weathering event during the latest Oligocene (~25 Ma)**

484 The most pronounced ϵ_{Nd} excursion observed at Site 274 (from -6.8 ± 0.3 to -3.7 ± 0.3)
485 between 24.0 and 23.8 Ma is a peculiar feature of our new record and may reflect a local change
486 in seawater composition (Figs. 4; 6). Although only represented by a single data point, another
487 radiogenic isotope excursion is hinted at ~ 33.5 Ma ($\epsilon_{Nd(t)} = -3.1 \pm 0.3$; Figs. 4; 6). Considering
488 the relatively invariant Nd isotope compositions of Oligocene Pacific deep water (Scher, 2014;
489 McKinley et al., 2019) and proto-CDW (this study; Wright et al., 2018; Evangelinos et al.,
490 2020) during the Oligocene, we consider two potential causes for the radiogenic excursion at
491 Site 274: (i) increased boundary exchange/volcanic activity, and (ii) pulses in Antarctic ice
492 advance leading to increased weathering inputs to local seawater.

493 Volcanic activity and deposition of material from the McMurdo volcanic group ($\epsilon_{Nd} >$
494 0; see references in Cook et al., 2013) have been reported to occur predominantly after 24 Ma
495 (Roberts et al., 2013). In addition, there is little physical evidence of volcanic contributions to
496 the sediments at Site 274 (Hayes et al., 1975), an observation that is confirmed by our measured
497 detrital sediment $\epsilon_{Nd(t)}$ values of -8.5 ± 0.3 (Fig. 4). Furthermore, REE patterns indicate that
498 fish debris predominantly preserve a seawater signature at 23.8 Ma (Fig. 5). We therefore

499 exclude the possibility of boundary exchange/volcanic activity as a cause for the positive Nd
500 isotope excursion.

501 A more promising connection can be made with the timing of a major expansion of ice
502 from Antarctica. The Eocene-Oligocene transition is well documented as a major ice advance
503 on Antarctica (e.g., [Zachos et al., 1992](#); [Galeotti et al., 2016](#)). A seawater expression of
504 weathering inputs from this ice expansion has been documented in the Prydz Bay area, where
505 late Eocene and Eocene Oligocene Transition (EOT) fish teeth show a shift in seawater
506 composition consistent with increased Antarctic weathering inputs ([Scher et al., 2011](#); [Bohaty
507 et al., 2012](#); [Wright et al., 2018](#)). Even though only documented by a single data point, the very
508 radiogenic $\epsilon_{Nd(t)}$ value of -3.1 ± 0.3 , our Site 274 exhibits the most extreme Nd isotope signature
509 ~ 33.5 Ma ([Figs. 4; 6](#)). A shift towards more radiogenic values may indicate erosion and
510 delivery of more radiogenic material exposed upstream in the Transantarctic Mountains, such
511 as basalts and sills of the Ferrar Large Igneous Province (ϵ_{Nd} : -3.5 to -6.9; see [Cook et al., 2013](#)
512 for a summary) or rocks like the Granite Harbour Intrusives in the Gabbro Hills ($\epsilon_{Nd} = \sim -2.3 \pm$
513 1.6 , $n=4$; [Borg et al., 1990](#)).

514 Interestingly, our data clearly define a second major seawater Nd isotope excursion
515 between 24.0 and 23.8 Ma, which appears to be closely associated with one of the most
516 significant climate transitions since the EOT resulting in a major Antarctic ice sheet expansion
517 (~ 24.5 -24 Ma; [Levy et al., 2019](#)). [Kulhanek et al. \(2019\)](#) described the first occurrence of ice
518 proximal glaciomarine sediments from DSDP Site 270 on the shelf of the Ross Sea at this time.
519 In addition disconformities are observed at another Ross Sea drill site proximal to the
520 Transantarctic Mountains (CRP-2/2A; [Naish et al., 2008](#)), and in seismic data ([Sorlien et al.,
521 2007](#)), which are consistent with the first major expansion of marine ice sheets across the Ross
522 Sea continental shelf (see also [Levy et al., 2019](#)). We consider it very likely that large changes
523 in the Antarctic cryosphere, i.e. large ice advance in this case, would have changed significantly

524 the erosional input from Antarctica to the Ross Sea margin, and hence would have imprinted a
525 radiogenic Nd isotope signal in local seawater.

526

527 **6. CONCLUSIONS**

528 New Oligocene to Miocene seawater Nd isotope data from Site 274 in the Adare Basin, Ross
529 Sea, reveal two major shifts in seawater chemistry in response to major steps in Antarctic ice
530 sheet expansion at the Eocene/Oligocene boundary (33.7 Ma) and between ~24.5 and 24.0 Ma.
531 In contrast, Site 278 in the southern Emerald Basin shows an invariant Nd isotopic composition
532 ($\epsilon_{Nd(t)} = -6.0$ to -6.7) throughout the Oligocene and early Miocene (31-22 Ma). Comparison
533 with previously published data suggests the presence of two deep water masses in the South
534 Pacific prior to the inferred onset of the ACC (34-30 Ma), one occupying depths between
535 ~2500 and 3000 m with Nd composition $\epsilon_{Nd(t)} = -3$ to -5 and a deep/bottom water mass (>3000
536 m) with more unradiogenic Nd composition ($\epsilon_{Nd(t)} \sim -6$). Following the onset of the ACC and
537 the opening of the Tasman Gateway (30-22 Ma), the seawater Nd isotope record at Site 278,
538 remained more radiogenic than the proto-CDW in the South Atlantic and Indian Ocean. We
539 attribute the provinciality in the Nd isotope composition of deep waters across the Tasmanian
540 Gateway to a weaker, less deep reaching ACC (proto-ACC) than today. Our finding implies
541 that the modern strong and deep-reaching ACC, which creates a near homogenous Southern
542 Ocean Nd isotope fingerprint must developed at a later point in the Neogene.

543

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545

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558

559 **Data availability**

560 The datasets to this article are available in the Supplement.

561

562 **Appendix A. Supplementary data:** Supplementary figures and tables

563

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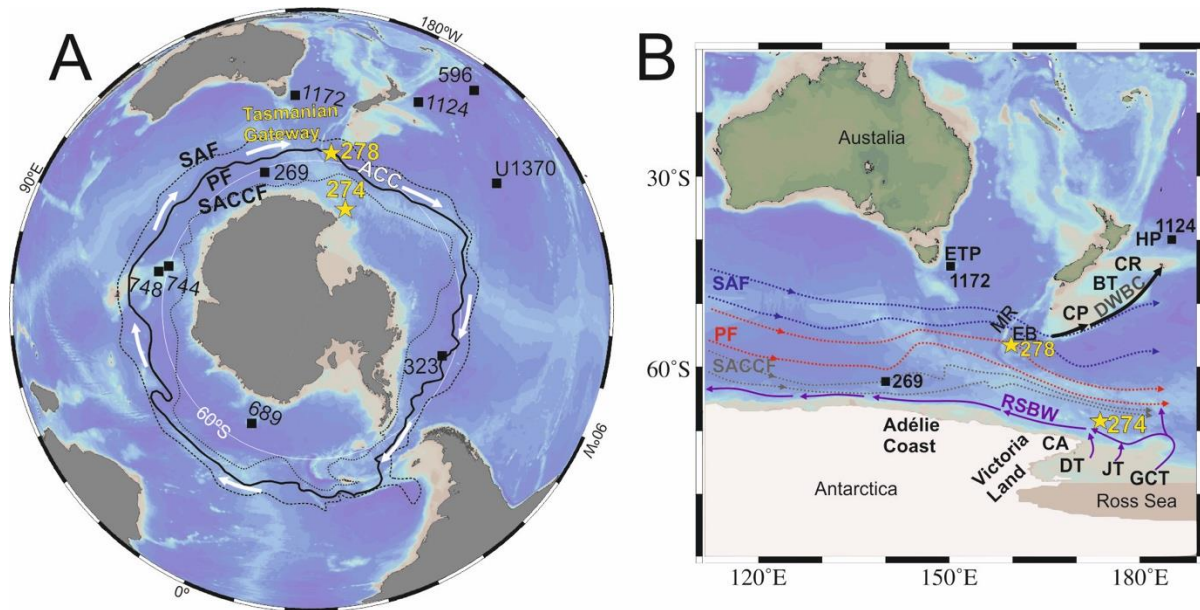
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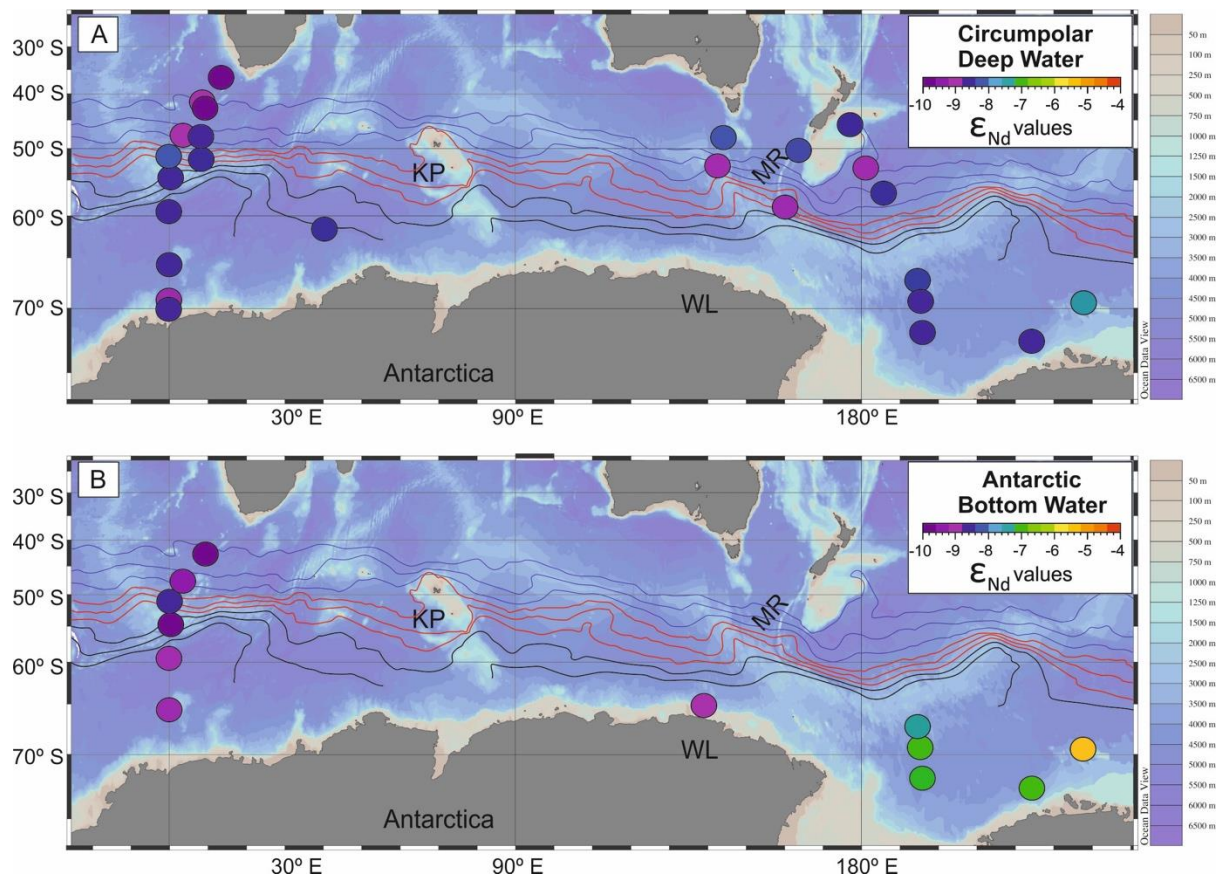
792 **List of Figures**



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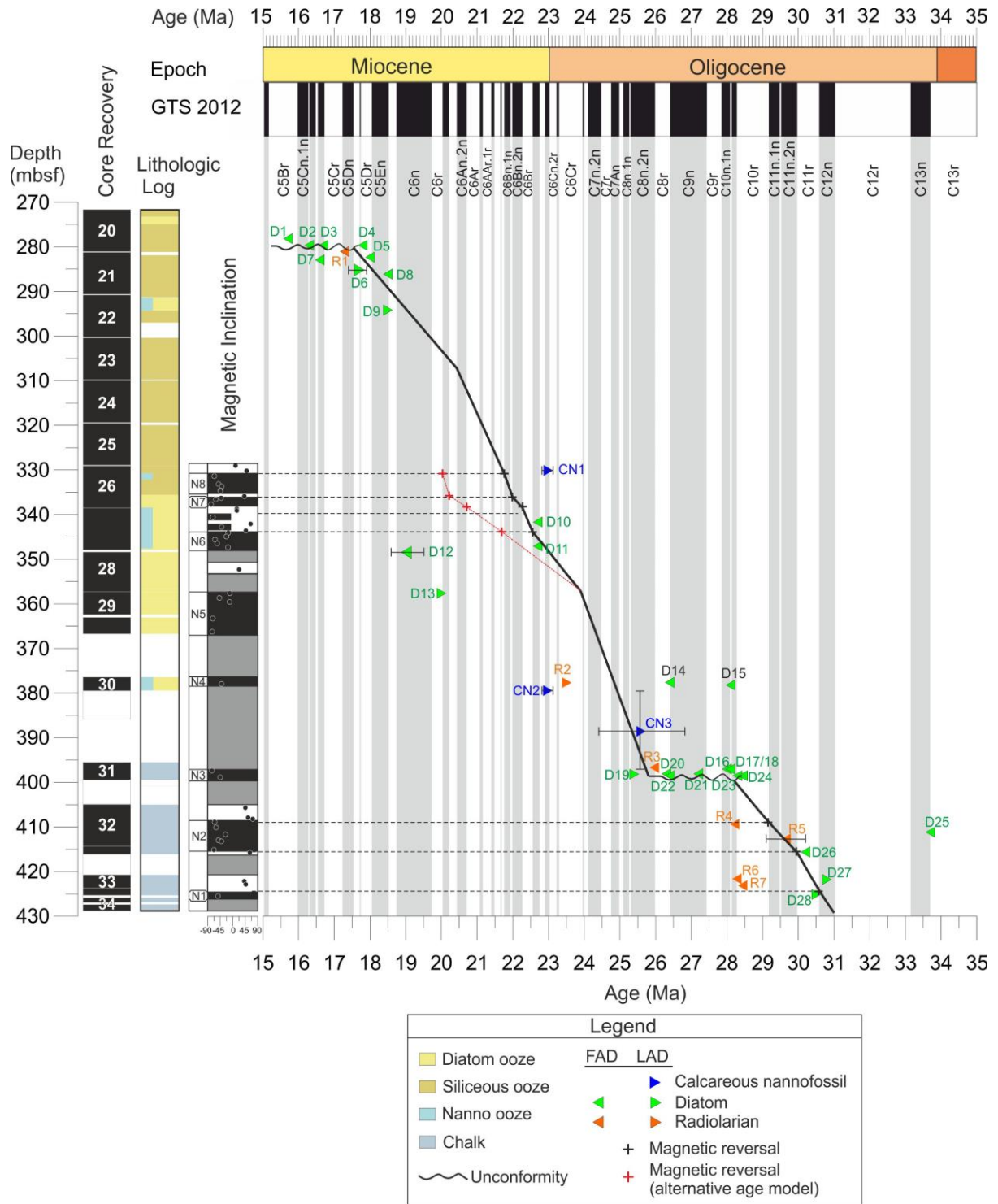
795 **Figure 1.** Overview map of modern Southern Ocean configuration (Ocean Data View,
796 (Schlitzer, 2016; version 4.79; <http://odv.awi.de>). **A:** Antarctic Circumpolar Current (ACC)
797 frontal system, study sites (ODP Sites 274 and 278) (yellow stars) and sites referred to in this
798 study (black squares). SAF: Subantarctic front (black dashed line), PF: Polar front (black line),
799 SACCF: Southern Antarctic Circumpolar Current front (black dotted line) adapted from Orsi
800 et al. (1995). **B:** Major modern deep ocean water circulation paths across the Tasmanian
801 Gateway adapted from Sokolov and Rintoul. (2007) and Gordon et al. (2009). The main
802 transport of Circumpolar Deep Water along the ACC takes place around the PF and SAF, Ross
803 Sea bottom water (RSBW) (purple) feeds Antarctic Bottom Water. Deep Western Boundary
804 Current (DWBC) (black arrows), MR: Macquarie Ridge, EB: Emerald Basin, ETP: East
805 Tasman Plateau, CP: Campbell Plateau, BT: Bounty Trough, CR: Chatham Rise, HP:
806 Hikurangi Plateau, CA: Cape Adare, JB: Joides Trough, DB: Drygalski Trough, GCT: Glomar
807 Challenger Trough.

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811 **Figure 2.** Modern neodymium (Nd) isotope composition for deep and bottom waters in the
 812 Southern Ocean (between 0°E and 110°W). Figures are made with Ocean Data View (Schlitzer,
 813 2016; version 4.79; <http://odv.awi.de>). **(A):** Nd isotope composition of Circumpolar Deep
 814 Water (CDW). **(B):** Nd isotope composition of Antarctic Bottom Water (AABW) (Stichel et
 815 al., 2012; Rickli et al., 2014; Garcia-Solsona et al., 2014; Basak et al., 2015; Lambelet et al.,
 816 2018; Amakawa et al., 2019). Antarctic Circumpolar Current (ACC) fronts (mean positions)
 817 adapted from Sokolov and Rintoul, 2007. SAF (blue), PF (red), SACCF (black), MR:
 818 Macquarie Ridge, WL: Wilkes Land, KP: Kerguelen Plateau



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820 **Figure 3.** Revised age-depth model for Site 278. Stratigraphy of Site 278, from left to right:

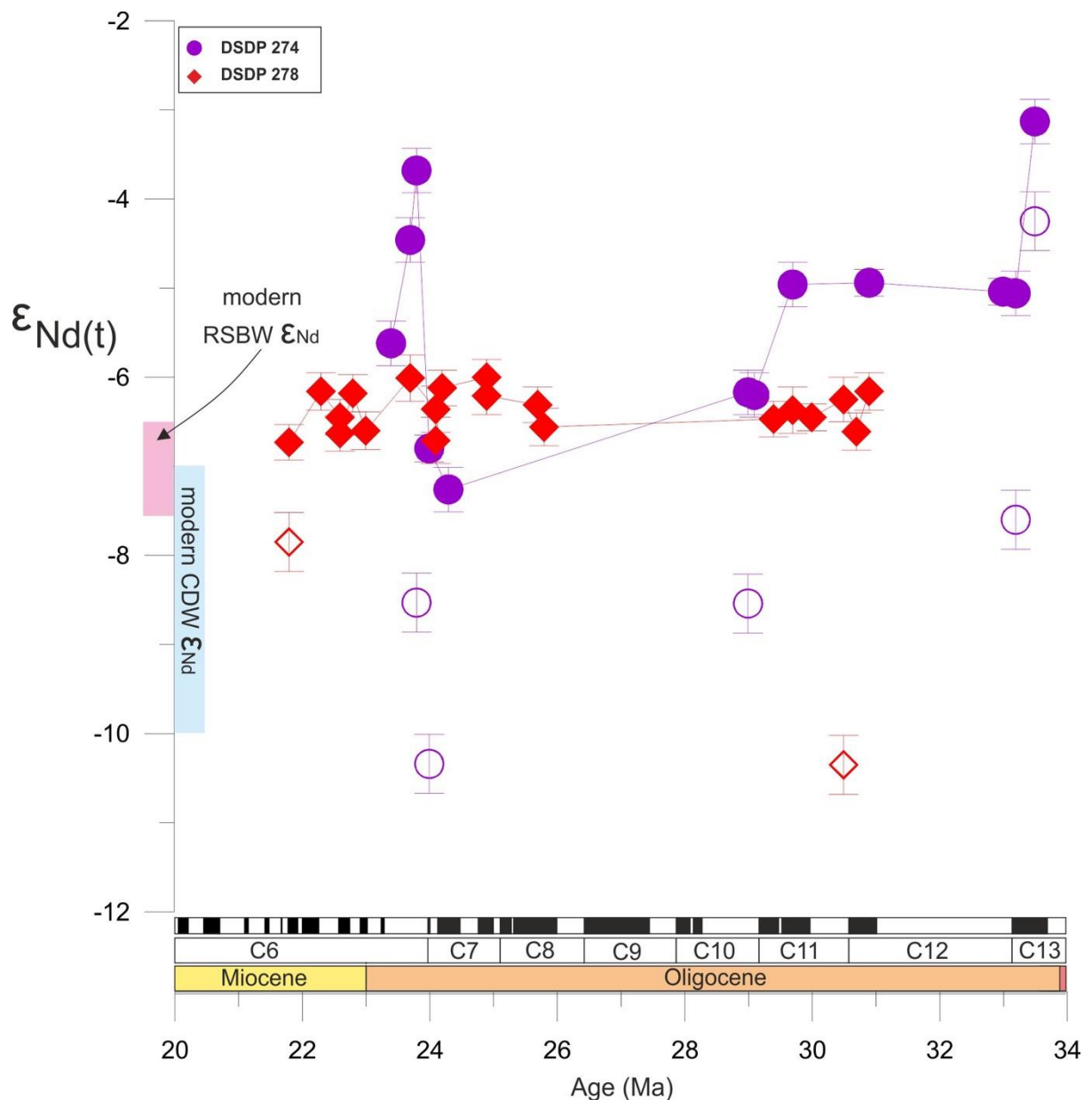
821 depth in meters below sea floor (mbsf), core recovery, lithologic log based on initial shipboard

822 sedimentological descriptions and local polarity based on inclination values. Grey vertical

823 stripes mark the normal polarity zones of the geomagnetic polarity timescale (normal=black,

824 reversed=white) (Gradstein et al., 2012). Black/red crosses= magnetostratigraphic constraints,

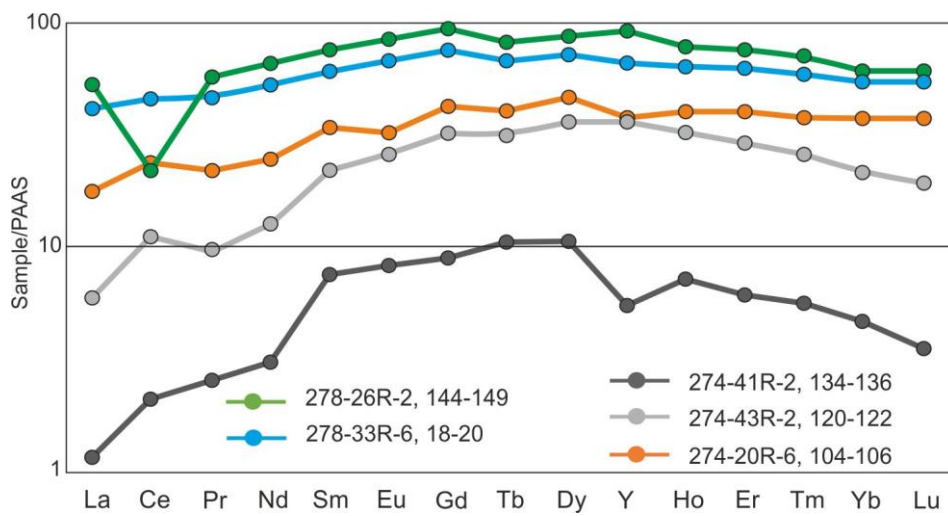
825 green triangles= diatom biostratigraphic constraints, blue triangles = calcareous nannofossil
 826 biostratigraphic ranges, orange triangles= radiolarian biostratigraphic constraints. FAD=first
 827 appearance datum; LAD= last appearance datum . Horizontal error bars on biostratigraphic
 828 events indicate total age range for the given datum. Vertical error bars indicate the depth range
 829 for the given datum. Datum and biozone labels correspond to those in Tables: S1, S2, S3 and
 830 S4. The back line following the different constraints shows our preferred chronology, and in
 831 red a possible alternate.
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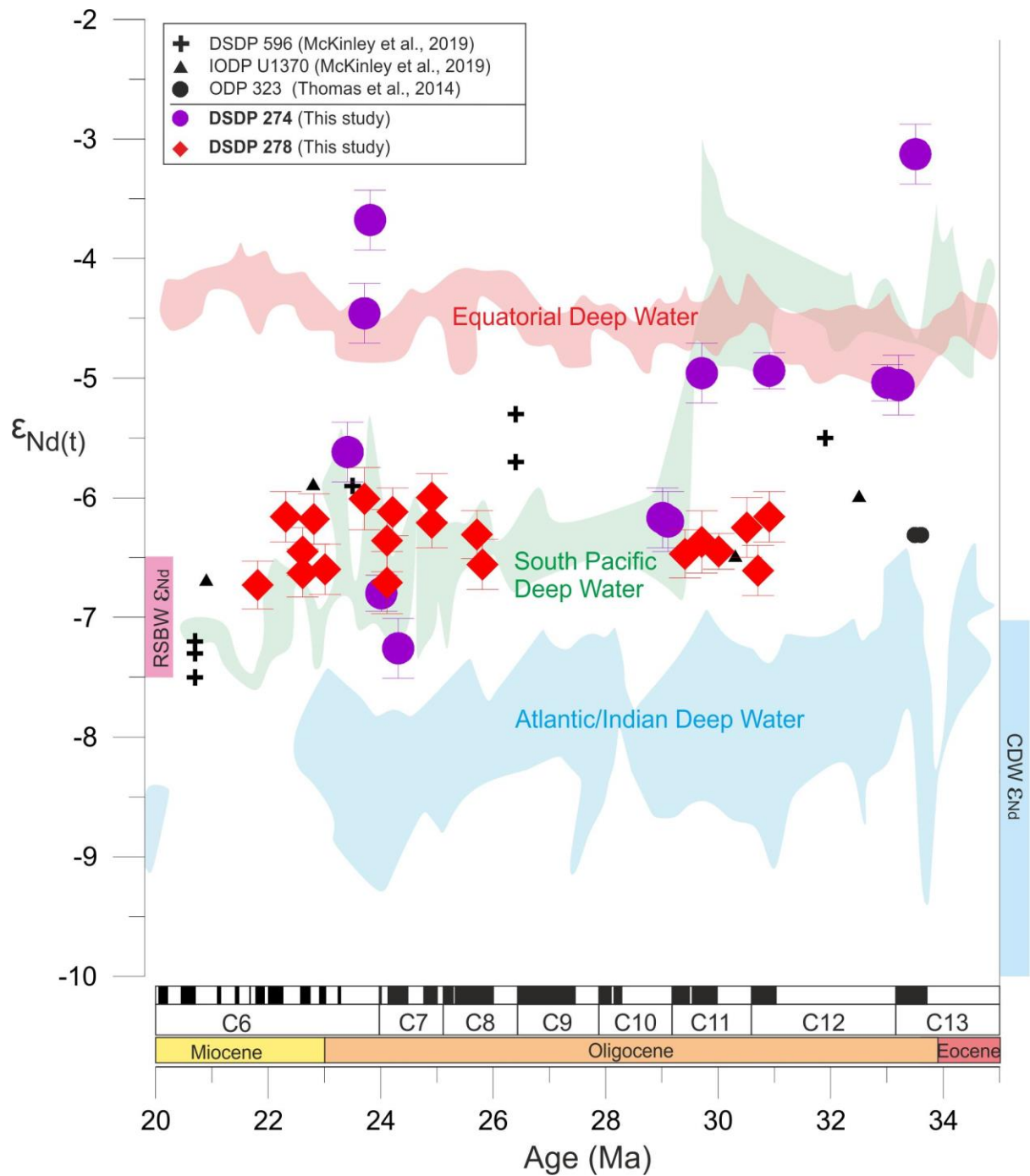
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836 **Figure 4.** Fish debris $\epsilon_{Nd(t)}$ records generated in this study from DSDP Sites 274 (purple circles)
 837 and 278 (red diamonds). Detrital sediment $\epsilon_{Nd(t)}$ values from DSDP Sites 274 (purple open
 838 circles) and 278 (red open diamonds). Present-day Ross Sea Bottom Water (RSBW) ϵ_{Nd}
 839 endmember ranges shown in magenta (Rickli et al., 2014; Basak et al., 2015). Present-day
 840 Circumpolar Deep Water (CDW) ϵ_{Nd} endmember ranges shown in blue (Lambelet et al., 2018
 841 and references therein).

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 848 **Figure 5.** Rare earth element (REE) patterns normalised to Post Archean Shale (PAAS)
 849 concentrations (Taylor and McLennan, 1985) for DSDP Sites 278 and 274.

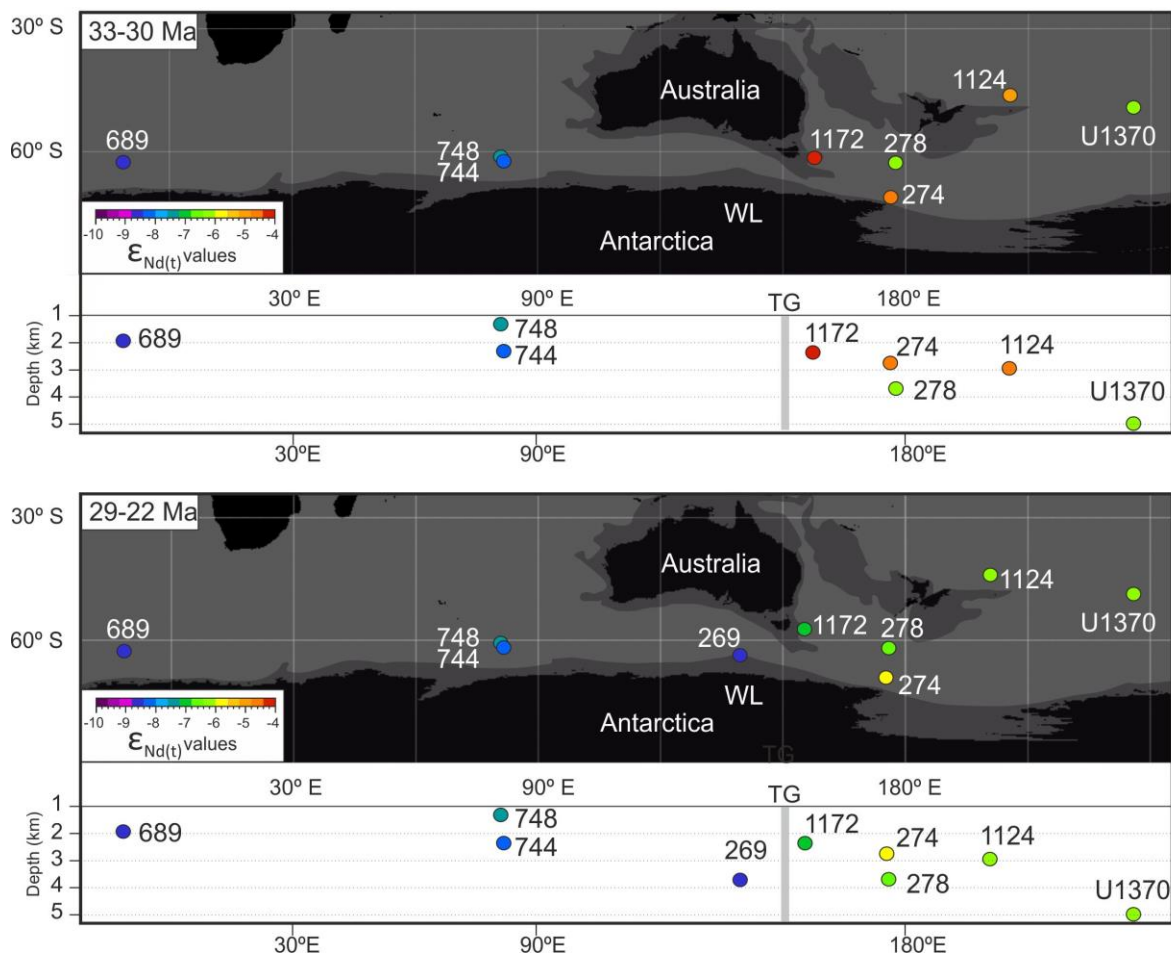


850

851 **Figure 6.** Comparison of fish debris ϵ_{Nd} records generated in this study from DSDP Sites 274
 852 and 278 with those derived from sites along the proto-CDW path in the Atlantic, Indian and
 853 Pacific sectors of the Southern Ocean. Atlantic/Indian Ocean locations include ODP Site 689
 854 (Maud Rise; [Scher and Martin, 2004](#)), ODP Sites 748 and 744 (Kerguelen Plateau; [Wright et al., 2018](#)), and DSDP Site 269 (off Adélie Coast; [Evangelinos et al., 2020](#)) (light blue shading).
 855
 856 ODP Site 1124 (Hikurangi Plateau) and ODP Site 1172 (East Tasmanian Plateau; [Scher et al.,](#)

857 2015) are combined in the green shading. Pacific Equatorial Age Transect (PEAT) Nd isotope
 858 results from the Equatorial Pacific are based on IODP Sites U1331, U1332, U1333, U1334,
 859 and U1335 (Scher, 2014) (red shading). The only data shown as individual symbols other than
 860 our new results are South Pacific DSDP Site 596, IODP Site U1370 (McKinley et al., 2019)
 861 and ODP Site 323 (Thomas et al., 2014). Blue shaded ranges refer to the present-day Nd
 862 signature of the CDW and magenta shaded rangers refer to the present-day Nd isotopic
 863 composition of the RSBW (Rickli et al., 2014; Basak et al., 2015; Lambelet et al., 2018 and
 864 references therein).

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866

867 **Figure 7.** Map of neodymium (Nd) isotopic data discussed in the main text plotted against
 868 paleo-latitude and longitude (upper panels) and paleo-depth (lower panels) from 33 Ma to 30
 869 Ma and from 29 to 22 Ma. Plate tectonic reconstructions and paleolocation of study Sites 278
 870 and 274, and other sites discussed in the main text were adapted from G-plates, based on the

871 global plate motion model from [Muller et al. \(2016\)](#). Continent with present-day coastlines
872 (black), grey boundary around continents shows areas of non-oceanic crust (e.g. continental
873 shelves). Circles represent site locations and are coloured based on their average $\epsilon_{Nd(t)}$ values
874 for each period considered. Grey band indicates the position of the Tasmanian Gateway (TG).

875

876 **Supplementary Information**

877 **Supplementary text**

878 **Age model for Site 278**

879

880 An age-depth model for the Oligocene and lower Miocene intervals of DSDP Site 278
881 is developed from the integration of new magnetostratigraphic data, calcareous nannofossil,
882 marine diatom, and radiolarian biostratigraphy, calibrated using the Geological Time Scale
883 (GPTS) 2012 ([Gradstein et al., 2012](#)) ([Fig. 3](#); [Tables S1-S4](#)). The initial age model prepared by
884 the Shipboard Scientific Party on DSDP Leg 29 ([Kennett et al., 1975](#)) was of coarse resolution
885 due in part to the infancy of diatom and radiolarian biostratigraphy. A compilation of age
886 models from Neogene intervals of DSDP drill holes presented by [Lazarus et al. \(1995\)](#) included
887 DSDP Site 278 and incorporated improved biostratigraphic age assignments as a result of 20
888 years of additional development and subsequent Ocean Drilling Program (ODP) drilling in the
889 Southern Ocean during ODP Legs 113, 114, 119, and 120. These later drilling efforts included
890 consistent integration of magnetic polarity stratigraphy in assessing the ages and correlation of
891 biostratigraphic datum levels. The compilation of [Lazarus et al. \(1995\)](#) provided the general
892 trend of age-depth for DSDP 278, but did not identify several short hiatuses of 2 – 4 m.y.
893 duration that are resolved in our analysis. We report here the age interpretation for the time
894 interval ~31 to ~18 Ma, and refer the reader to the report of [Evangelinos et al. \(in preparation\)](#)
895 for a revised age model for the time interval < 18 Ma at DSDP Site 278.

896 A total of 19 samples were examined for the identification of calcareous nannofossils.
897 Smear slides were prepared from unprocessed sediment and examined with a Leica DMRP
898 polarizing microscope at 1000 x magnification. Occasionally, x1250 and x1600 magnification
899 was also used for the identification of small specimens. At least 3 mm² (ca. 200 visual fields at
900 x1000) were scanned to provide semi-quantitative values. To detect the presence of rare or very
901 rare specimens an additional two to three slide traverses were observed. The nannofossil
902 assemblages were used to define the First and Last appearance datums (FAD and LAD)
903 Selected calcareous nannofossil bioevents were considered in moderate to well preserved
904 samples, Standard low to mid-latitude markers of the Martini (1971) and Okada and Bukry
905 (1980) were absent. We instead used the high southern latitude zonations developed by Wei,
906 1991 and 1992; Wei and Wise (1990, 1992); Monechi and Reale (1997); Watkins and Villa
907 (2000); Agnini et al. (2014). Biochronology was developed following Lourens et al. (2004)
908 and Raffi et al. (2006), using astronomically calibrated ages from magnetic reversals and
909 reference isotope stratigraphies, adjusted to the ATNTS2004 (Gradstein et al., 2004). Only 12
910 of these samples contained enough amounts of nannofossils to provide biostratigraphic
911 information ([Table S1](#)). In most of these samples calcareous nannofossils are abundant. The
912 lowest cores 278-34R to 31R consist of siliceous nannofossil chalk comprising 70 to 95%
913 calcareous nannofossils. Diatoms increase in number at the expense of calcareous nannofossils
914 in core 31R, and become the dominant sediment-forming microfossil in cores 30R and above.
915 Calcareous nannofossils are present in most intervals, but their abundance drops to trace
916 amounts or absent in some samples in cores 29R and 28R ([Kennett et al., 1975](#)).

917 Diatom biostratigraphic data for DSDP Site 278 were compiled from a variety of
918 published sources, as well as new data collected for this study. [Schrader \(1976\)](#) provided the
919 initial biostratigraphic characterization of 72 samples in this interval between samples 278-
920 34R-3, 50 cm to 278-21R-1, 50 cm (his dataset continues up to core 278-9R-1, 50 cm),

921 reporting 58 diatom taxa in this lower interval. [Weaver and Gombos \(1981\)](#) compared the
922 stratigraphy of DSDP Sites 278 and 266 in an attempt to reconcile divergent records of diatom
923 ranges, with the resultant interpretation of a hiatus within core section 15R-2 in DSDP Site 278
924 that resulted in truncated ranges. [Harwood \(1982\)](#) examined 20 samples across the Oligocene-
925 Miocene interval between samples 278-34R-1, 135 cm and 278-25R-2, 44 cm in the attempt to
926 apply the diatom zonation proposed by [Gombos and Ciesielski \(1983\)](#) from DSDP Leg 71 in
927 the South Atlantic to the DSDP Site 278 record, and documented the occurrence of 69 taxa.
928 Based on the stratigraphic abundance distribution of *Rocella* spp. taxa, [Harwood \(1982\)](#)
929 interpreted a hiatus between samples 278-31R-3, 139-141 cm and 278-3R1-2, 139-141 cm,
930 which was also noted in [Fenner \(1984\)](#), along with the ranges of 19 other diatoms in DSDP
931 Site 278. 35 new samples were examined by Harwood for the present study in the interval
932 between samples 278-34R-1, 142 cm and 278-26R1, 7 cm, wherein 112 diatom taxa were
933 noted, along with 17 other siliceous microfossil taxa (silicoflagellates, ebridians, endoskeletal
934 dinoflagellates, and a range of unrecorded chrysophycean cysts). The diatom and other
935 siliceous microfossil record require further documentation, as it is clear that the DSDP 278
936 record is an important reference section for diatom biostratigraphy. [Schrader \(1976\)](#) appears to
937 have examined strewn slides from chemically-treated sediment residues, whereas [Harwood](#)
938 [\(1982\)](#), following the methodology of [Gombos and Ciesielski \(1981\)](#), examined slides of size
939 fractions from $>37\ \mu\text{m}$ and $<37\ \mu\text{m}$, as well as raw strewn, chemically-treated sample residues.
940 Similarly, in preparation for the present study, diatom samples were sieved through a smaller,
941 20 μm mesh sieve. Sieving results in the recovery of a greater number of whole and identifiable
942 taxa, and is responsible for the recording of numerous taxa not reported by [Schrader \(1976\)](#),
943 including many marker taxa used in the [Gombos and Ciesielski \(1983\)](#) zonation. In addition,
944 the recent compilation of biostratigraphic events developed for IODP Expedition 374 in the
945 Ross Sea ([McKay et al., 2019](#)) was consulted to provide ages for diatom events noted in this

946 study. This compilation drew from output from the Constrained Analysis (CONOP) model
947 output associated with the study of ODP Site 744 as presented in [Farmer \(2011\)](#) and [Florindo
948 et al. \(2013\)](#), which builds upon the composite ordering of events and model age output for
949 Southern Ocean data sets presented by [Cody et al. \(2008, 2012\)](#) ([Table S2](#)).

950 Radiolarian biostratigraphic event data of first and last appearance datum levels are
951 interpreted from Table 1 of [Petrushevskaya \(1975\)](#). Taxon names presented in [Table S3](#) reflect
952 those presented by Petrushevskaya, as well as names in current use. Age information and
953 taxonomic revisions are drawn from the works of [Abelmann \(1990\)](#), [Caulet \(1991\)](#), [Takemura
954 \(1992\)](#), and [Takemura and Ling \(1997\)](#), as well as from the age model output provided by the
955 CONOP analyses from ODP Site 744. Assignment of radiolarian datum events to magnetic
956 polarity chrons are drawn from the radiolarian range data of select taxa presented in [Abelmann
957 \(1990\)](#) from ODP Sites 689 and 690 in conjunction with age models for these sites provided
958 by [Florindo and Roberts \(2005\)](#).

959 In order to provide independent age constraints to biostratigraphy we studied the
960 magnetic stratigraphy of DSDP Site 278. The interpreted magnetic chrons are presented in
961 [Table S4 and in Figures 3](#). Magnetic reversals in the samples were identified through stepwise
962 demagnetisation experiments by alternating fields (AF) using the 2G magnetometer (2G-
963 SRM750 Superconducting Rock Magnetometer) at the Paleomagnetic Laboratory of Barcelona
964 (CCiTUB-CSIC). A total of 47 samples provided reliable inclinations. Most of the samples
965 were successfully demagnetized after 80 mT, although some of them likely with higher
966 coercivity minerals were demagnetized after 120-150 mT. Core orientation is not
967 reconstructed, and magnetic inclinations are used to determine magnetic polarities ([Fig. S3A](#))
968 In general samples depict normal and reversed polarities ([Fig. S3B](#)). We have identified 7
969 reversal events shown on [Figure 3 and Supplementary Table S4](#).

970 The stratigraphic section cored at DSDP Site 278 from the bottom of the sedimentary
971 section in core 34R at ~429 mbsf up to sample 278-21R-1, 50 cm at ~282 mbsf spans the time
972 interval from ~ 31 to 17.7 Ma, with one unconformity noted between samples 278-31R-3, 35
973 cm and 278-31R-2, 139 cm (398.85 – 399.89 mbsf), which removed a 2.4 m.y. time interval
974 from 28.2 to 25.8 Ma. The top of this interval is bounded by an unconformity between samples
975 278-21R-1, 50 cm and 278-20R-6, 50 cm (281.5 mbsf) which removed the time interval 17.6
976 to 15.5 Ma.

977 Sediments between 396.95-428.45 mbsf (Cores 34R-31R) contain calcareous
978 nannofossils *Reticulofenestra daviesii*, *Cyclicargolithus floridanus*, *Chiasmolithus altus*,
979 *Reticulofenestra bisect/stavensis*, and *Cyclicargolithus abisectus*. In this interval occasional
980 reworked taxa have been observed. The interval between 379.5 and 414.41 mbsf is assigned to
981 be between NP25 (23.13-26.84 Ma) and NP23 (29.62-32.02 Ma) due to the co-occurrence of
982 *Chiasmolithus altus* (absent above 379.5 mbsf) and *Cyclicargolithus abisectus* (regular above
983 410.41 mbsf). The interval below 422.74 mbsf is attributed to an interval between biozones NP
984 23-NP 16?, due to the absence of *Reticulofenestra umbilicus* and the consistent presence of
985 *Reticulofenestra bisecta/stavenis* (Martini et al.1970; Watkins and Villa, 2000; Young et al.,
986 2017, Kulhanek et al., 2019). The LAD of diatoms *Pyxilla eocena* and *Hemiaulus rectus* var.
987 *twista* at 422.09 and 425.43 mbsf, respectively indicate an age near 30.8 to 30.5 Ma at the
988 bottom of DSDP 278. Recognition of the FAD of diatom *Rocella vigilans* var. A at 421.85
989 mbsf (30.2 Ma) and the FAD of radiolarian *Cyrtocapsella longithorax* at 417.45 mbsf (31.5 to
990 32.5 Ma) further constrains the age of core 33R, and allows assignment to the polarity Chrons
991 C12n and C11n (Fig. 3; Tables S2-S4). Accordingly, the magnetic reversal at 425.44 mbsf (N1)
992 is interpreted as the top of Chron C12n (30.59 Ma). Guided by the biostratigraphic datum it is
993 possible to place the base of C11n.2n and top of C11n.1n (Fig. 3, N2; Tables S4). These tie-
994 points and derived sedimentation rate can be extended complementing the age model up to

995 section 31R-3, where a cluster of diatom first appearance datum levels is noted, indicating the
996 presence of a hiatus. The FAD of diatoms '*Coscinodiscus*' *rhombicus* (28.45 Ma), *Azpeitia*
997 *gombosi* (28.3 Ma), *Rocella gelida* (26.4 Ma) in a sample from 398.85 mbsf, the FAD of
998 diatoms *Triceratium groningensis* (27.2 Ma) and *Rocella schraderi* (~26.3 Ma) in sample from
999 398.39 mbsf, and the FAD of radiolarian *Stylosphaera radiosa* (~26.0 Ma) in a sample from
1000 ~397 mbsf identify this break in time. The LAD of diatom *Rocella vigilans* (25.4 Ma) in a
1001 sample from 398.39 mbsf identifies when sedimentation resumed after the erosion or non-
1002 deposition associated with the hiatus (Fig. 3). Accordingly, the interval of normal polarity (N3)
1003 measured between 398.86 and 397.34 mbsf is interpreted to correlate to Chron C8n.2n (Table
1004 S4).

1005 A major change in the sedimentation occurs in core 30R (378.91-379.75mbsf) marked
1006 by the high abundance of siliceous microfossils (Kennett et al., 1975). This interval is
1007 characterized by the presence of calcareous nannofossils *Reticulofenestra daviesii*,
1008 *Cyclicargolithus floridanus*, *Discoaster deflandrei*, while regular taxa recorded above, such as
1009 *Reticulofenestra bisecta/stavensis* and *Chiasmolithus altus* are absent. This assemblage is
1010 assigned to zone NP25 (23.13-26.48 Ma) (Martini et al., 1970; Young et al., 2017), late
1011 Oligocene; however, due to the absence of characteristic marker species in the studied samples,
1012 we cannot discard an assignment of this interval to NP25-NN1. Despite the weak and low
1013 inclinations at this interval, we tentatively assign the normal polarity zone (N4) found at 377.94
1014 mbsf to Chron C7An, suggesting an age of 24.8 Ma. A long interval of normal polarity
1015 identified in five samples from 366.28 to 357.66 mbsf (N5; Table S4) is interpreted as
1016 correlative to Chron C7n.2n (Fig. 3). One sample of reversed polarity at 352.32 mbsf is here
1017 interpreted as Chron C6Cr.

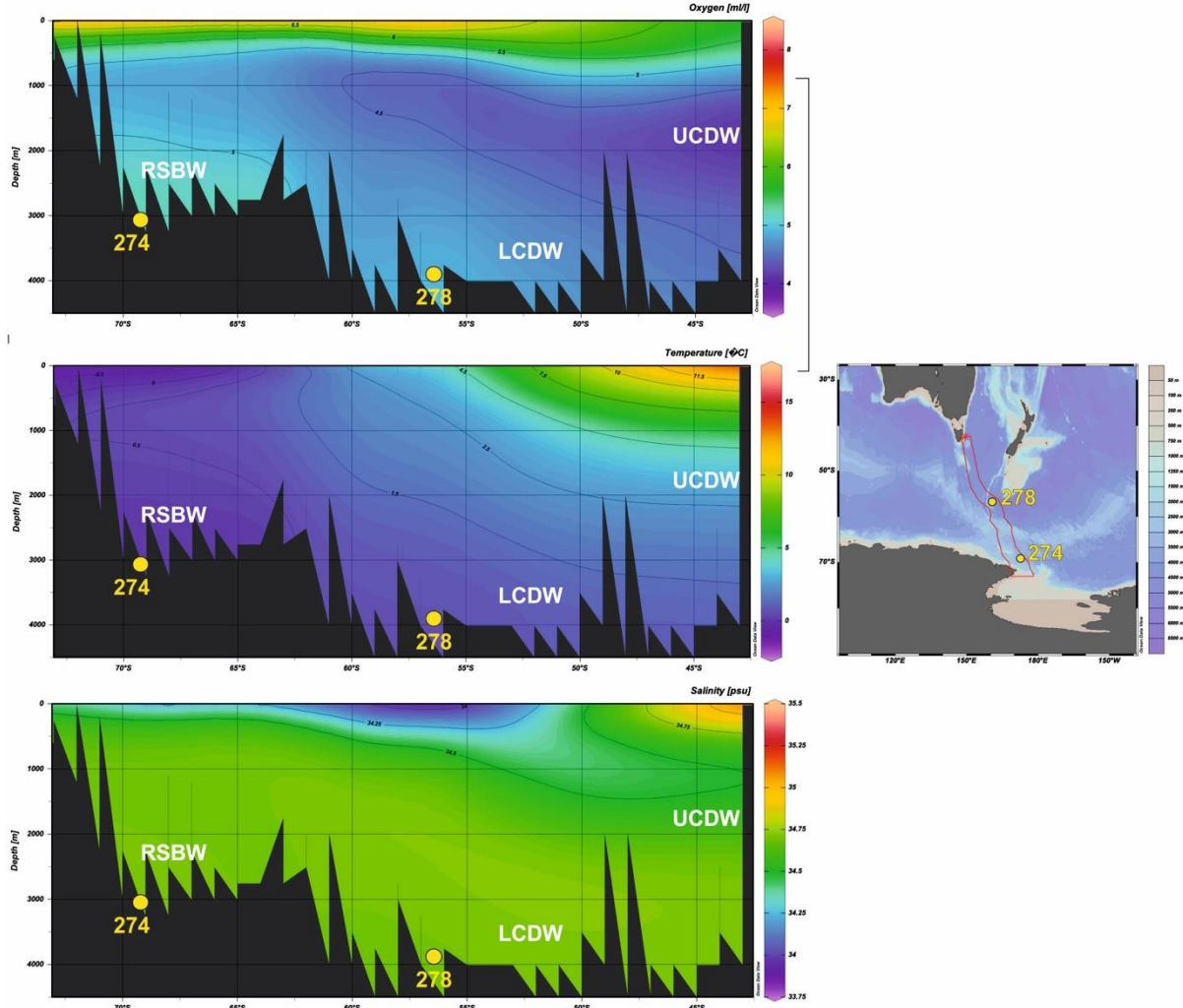
1018 The FAD of diatom *Raphidodiscus marylandicus* (22.7 Ma) is noted in a sample at
1019 347.39 mbsf and *Thalassiosira spumellaroides* (22.7 Ma) in a sample at 342 mbsf. These

1020 diatom assemblages together with the presence of the calcareous nannofossil *Cyclicargolithus*
1021 *floridanus* and absence of *Reticulofenestra bisecta/stavensis* and other markers recorded in the
1022 lower part of Site 278, support an assignment of this interval of DSDP 278 to be assigned to
1023 the interval to biozones NN1 (22.82-23.13 Ma) and NN2 (19-22.82 Ma) (Martini et al., 1970;
1024 Young et al., 2017; Kulhanek et al., 2019). This is consistent with the diatom record, although
1025 no precise markers were observed. The biostratigraphic constraints of cores 27R and 26R here
1026 guide the magnetostratigraphic correlation. The reversal at 337.76 mbsf (base of N7) is
1027 interpreted as the base of Subchron C6Bn.2n. Above, the reversed magnetozone is assigned to
1028 C6Bn.1r whilst the top of Chron C6Bn.1n is ascribed to the reversal occurring at 331.44 mbsf
1029 (Fig.3; Table S4). Because of the wide range provided by biostratigraphic constraints for the
1030 interval encompassing cores 29 to 26, an alternative magnetostratigraphic correlation is
1031 allowed. In such case, magnetozone N6 could correspond to Subchron C6Bn.2n and C6Bn.1n;
1032 N7 to C6An.2n; and N8 to C6An.1n. Nevertheless, we consider this correlation less likely as
1033 it implies a change in the sedimentation trend towards lower values, unlike the rest of our
1034 interpretation. In addition, the fit pattern between our local reversals and the GPTS is not as
1035 good as the one using our preferred model.

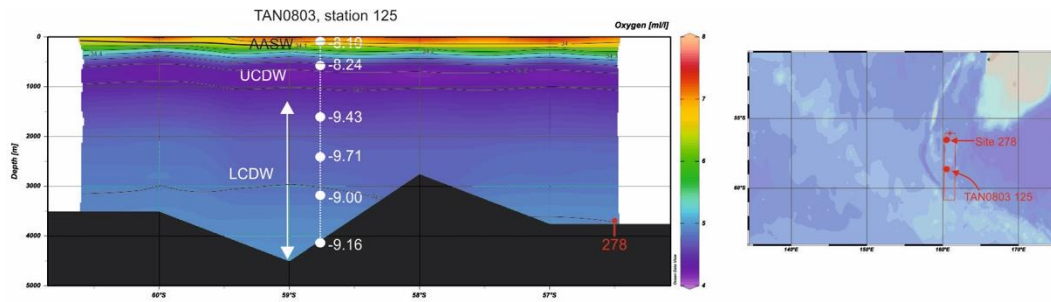
1036 Above the polarity reversals of core 26R, there is limited chronostratigraphic control
1037 up to cores 22R and 21R where numerous diatom events are noted. The LAD of *Thalassiosira*
1038 *spumellaroides* (18.5 Ma) noted in a sample at 294.5 mbsf and the FAD of *Fragilariopsis*
1039 *pusilla* (18.5 Ma) at 286.5 mbsf indicate the age at this level. Similarly, the close association
1040 of the FAD of *Crucidenticula ikebei* (18.0 Ma) at 282.66 mbsf and the LAD of *Thalassiosira*
1041 *praeфрага* (17.4 – 17.9 Ma) at 285.25 mbsf provides age control for this interval. Between
1042 cores 21R and 20R, we interpret the existence of a hiatus based on the coincident FADs of
1043 several diatoms in the sample at 280.01 mbsf: *Araniscus lewisianus* (17.8 Ma), *Denticulopsis*

1044 *maccollumii* (16.7 Ma), *Nitzschia grossepunctata* (15.6 Ma), and radiolarian FAD of
1045 *Eucyrtidium punctatum* (17.3 Ma) in a sample at 281.60 mbsf.

1046 **Supplementary Figures**
1047



1048 **Figure S1.** Maps showing the location of the sites and the distribution of dissolved oxygen,
1049 temperature and salinity versus depth along the Tasmanian to Ross Sea transect. Data derived
1050 from HYDROGRAPHIC ATLAS OF THE SOUTHERN OCEAN (Olbers et al., 1992). Maps
1051 were produced using Ocean Data View (Schlitzer, 2016). UCDW= Upper Circumpolar Deep
1052 Water; LCDW= Lower Circumpolar Deep Water; RSBW= Ross Sea Bottom Water.
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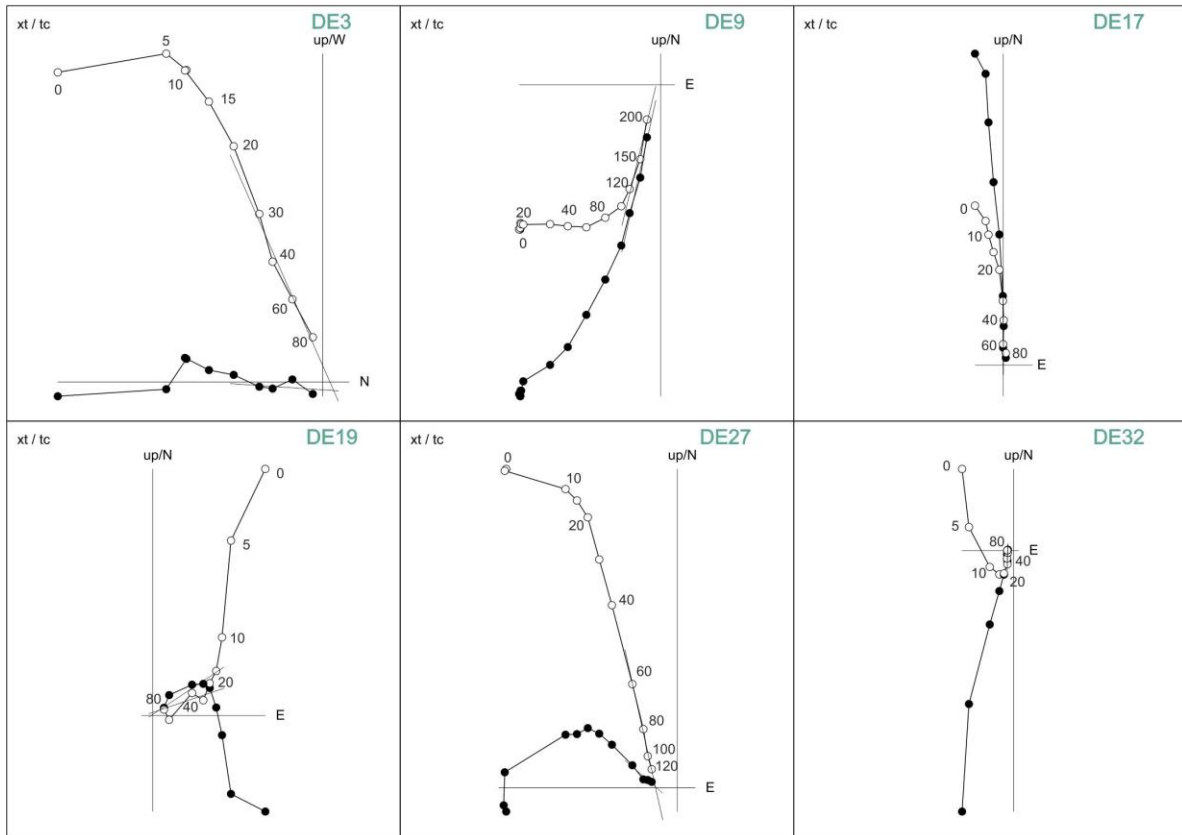
Figure S2. Neodymium isotopic composition from station TAN0803 station 125 (Lambelet et al., 2018) versus depth. Map and section were created using Ocean Data View software (Schlitzer, 2016). Antarctic Surface Water= AASW; UCDW= Upper Circumpolar Deep Water; LCDW= Lower Circumpolar Deep Water.

A) Demagnetization Diagrams (Alternating Field)

EXCLOR: dec = 183.8 inc = -66.3
int = 302 mad = 6.5

EXCLOR: dec = 192.6 inc = 41.9
int = 1038 mad = 2.1

EXCLOR: dec = 359.6 inc = -39.8
int = 119 mad = 4.4

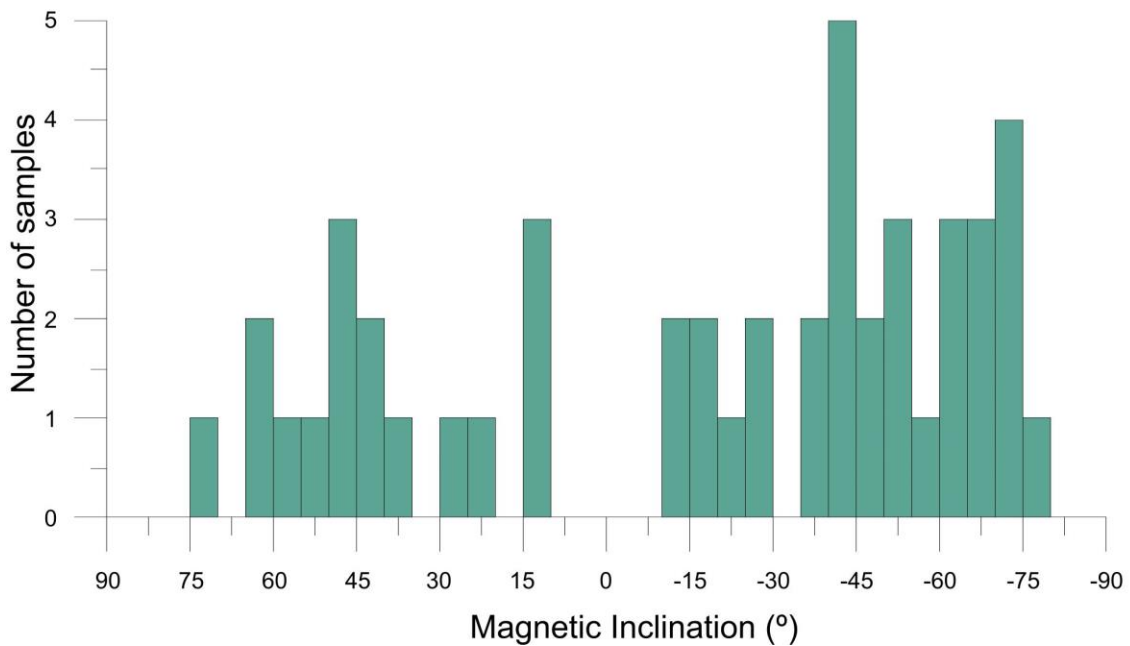


EXCLOR: dec = 56.3 inc = -15.9
int = 57 mad = 12.7

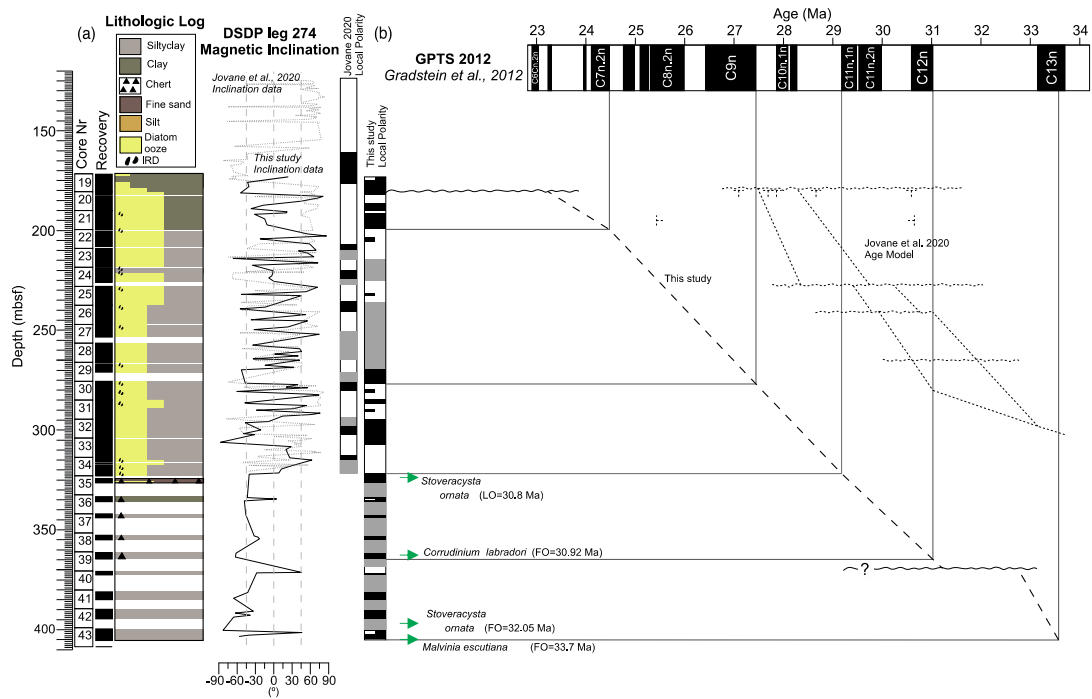
EXCLOR: dec = 311.0 inc = -73.7
int = 428 mad = 3.7

EXCLOR: dec = 181.4 inc = 45.5
int = 557 mad = 3.4

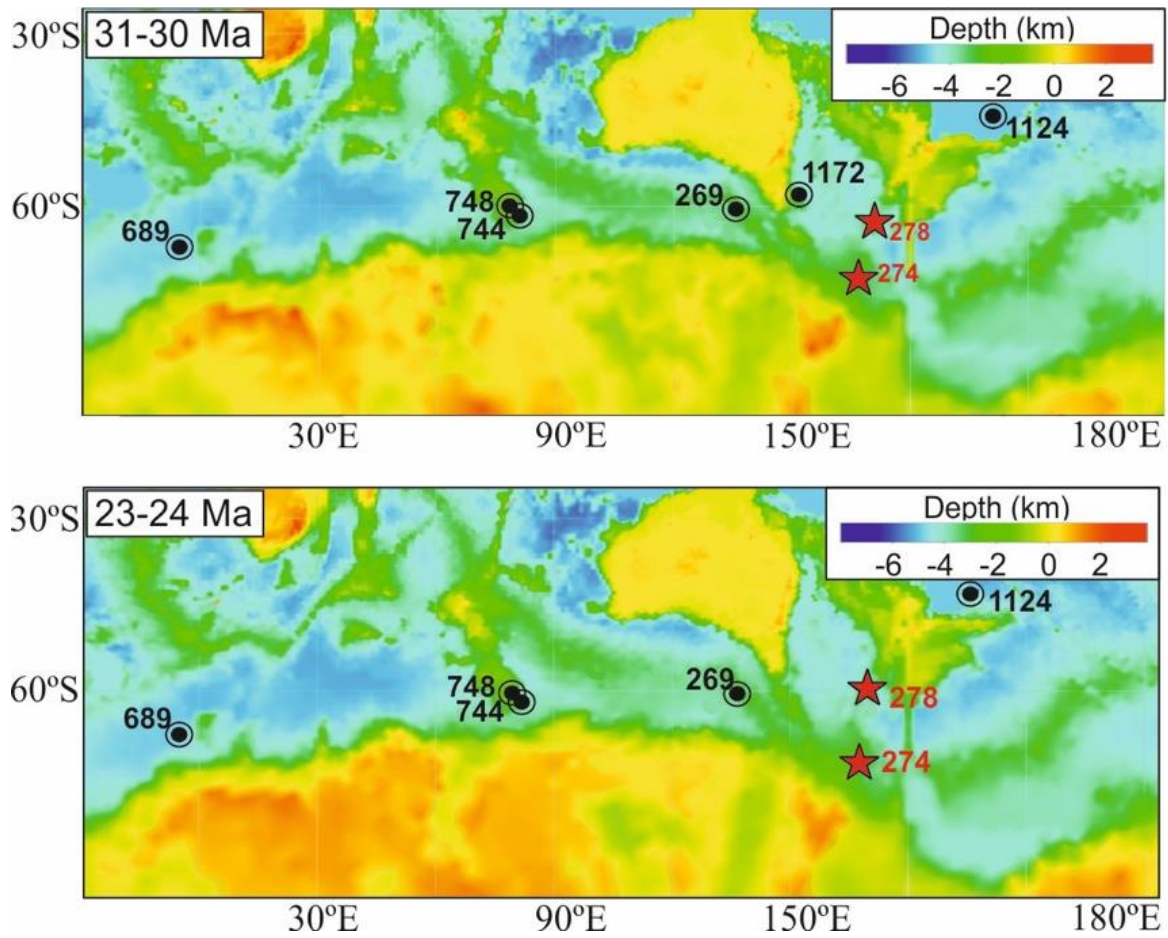
B) Distribution of Inclination



1065 **Figure S3: A: Demagnetization Diagrams B: Distribution of Inclination from Site 278**
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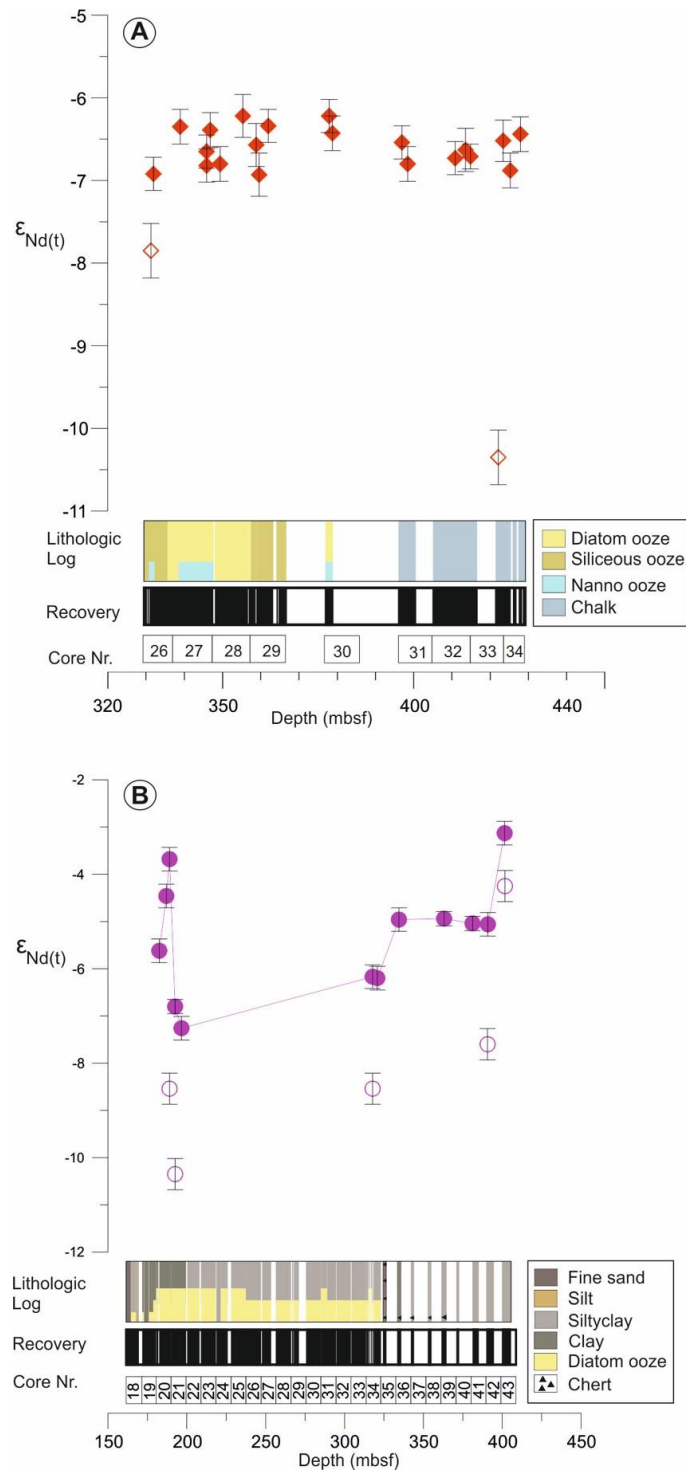


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 1069 **Figure S4.** Age mode for Site 274 from Hoem et al. (2021). (a) Core numbers, core recovery
 1070 and lithologic log are based on Hayes et al. (1975). (b) Magnetic correlation for Site 274 with
 1071 comparison to Jovane et al. (2020) (dotted lines). LO= Last occurrence and FO= First
 1072 occurrence (see Hoem et al. (2021) for details).
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Figure S5. Paleobathymetry and paleotopography reconstructions of the Southern Ocean derived from Paleo-digital elevation model by [Scotese and Wright \(2018\)](#). Reconstructions of the paleolocation of study sites and sites discussed were adapted from G-plates based on [Muller et al., 2016](#) global plate motion model.



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Figure S6: Neodymium isotopic composition ($\epsilon_{Nd(t)}$) of fish debris and detrital sediment samples from this study. **A:** Fish debris $\epsilon_{Nd(t)}$ values (red diamonds) and detrital sediment $\epsilon_{Nd(t)}$ values (red open diamonds) at Site 278. **B:** fossil fish debris $\epsilon_{Nd(t)}$ values (purple circles) and detrital sediment $\epsilon_{Nd(t)}$ values (purple open circles) at Site 274.

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