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Millennial-scale variations of the Holocene North Atlantic mid-depth gyre inferred from radiocarbon and neodymium isotopes in cold water corals

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35 values (from -15.4 ± 0.3 to 13.3 ± 0.2) suggesting a westward contraction of the mid-SPG and a higher
36 proportion of ENAW. The mid-Holocene long-term change in ϵNd is characterized by millennial
37 variations of up to 2.5 epsilon units well marked during the Late Holocene, indicating that eastward
38 extension of the mid-SPG coeval with warm periods in northern Europe (e.g. the Medieval Climatic
39 Anomaly and the Roman Warm Period). Most of the CWC-derived $\Delta^{14}\text{C}$ values match the global
40 ocean values indicating that the water masses bathing the corals were generally well ventilated; the
41 exceptions are a few short intervals of eastward extension of the mid-SPG, which are characterized
42 by lower $\Delta^{14}\text{C}$ during the late Holocene. We propose that these minor $\Delta^{14}\text{C}$ fluctuations in the Rockall
43 Trough may be related to local changes in the mixed layer depth or to variability in the advection of
44 water from the Labrador Sea where deep convection gives rise to thermocline waters that are
45 relatively depleted in terms of $\Delta^{14}\text{C}$.

46 The eastward extension of the mid-SPG between 8.8 and 6.8 ka BP is associated with the Holocene
47 climatic optimum, concurrent with a maximum strength of the Iceland-Scotland Overflow Water
48 (ISOW), which is indicative of an increase in the Atlantic Meridional Overturning Circulation (AMOC).
49 This period is followed by a transition in the North Atlantic circulation, which occurred around 6.8 ka
50 BP, and coincides with the onset of Labrador Sea Water formation, a decrease of the ISOW strength
51 and a contraction of the mid-SPG leading to a greater intrusion of saline subtropical water into the
52 subpolar Atlantic.

53

54 **Keywords:** Cold-water corals, Rockall Trough, Holocene, Nd isotopic composition, radiocarbon, North
55 Atlantic gyre dynamics.

56

57 **1. Introduction**

58 Although Holocene climate oscillations are small compared to the glacial and deglacial
59 variability, their study reveals key processes controlling the natural variability of the present warm
60 interglacial climate, which could be useful in evaluating the role of this variability in the future
61 climate. Variations in North Atlantic Ocean surface and mid-depth circulations are controlled by the
62 subtropical and subpolar gyres and have a significant influence on the properties of water entering
63 the Nordic Seas by redistributing heat and freshwater. However, the Holocene variability of these
64 North Atlantic surface and mid-depth circulations, and their link to deep water formation and
65 atmospheric forcing, are poorly documented.

66 The Holocene subpolar North Atlantic climate is characterized by an early to mid-Holocene “thermal
67 maximum” followed by progressive cooling induced by decreased insolation forcing (related to
68 orbital precession) (e.g. Marchal et al., 2002; Sarnthein et al., 2003). This climatic cooling reflects a
69 major reorganization of atmospheric and ocean circulation in the North Atlantic (e.g. O’Brien et al.,

70 1995; Came and Oppo 2007; Repschläger et al., 2017). Data- and model-based reconstructions show
71 a change of source for North Atlantic Deep Water (NADW) production which is associated with both
72 a decline in the production of Iceland-Scotland Overflow Water (ISOW) in the Nordic Seas during the
73 mid-Holocene (e.g. Rasmussen et al., 2002; Renssen et al., 2005; Kissel et al., 2013; Thornalley et al.,
74 2013) and the initiation of Labrador Sea Water (LSW) formation around 7 ka BP (Hillaire-Marcel et al.,
75 2001; Solignac et al., 2004; Hoogakker et al., 2011, 2015).

76 While Holocene variability of the lower branch of the AMOC is relatively well constrained, the
77 variations of the upper limb, which corresponds to surface and intermediate water masses
78 transported northward by the North Atlantic gyres, are not yet well defined (Ayache
79 et al., 2018). This is due to the complex spatial and temporal variability of the classic sea surface
80 temperature (SST) and sea surface salinity (SSS) proxies used to track past changes in the AMOC,
81 which are strongly influenced by the orbital-induced insolation trend and the disintegration of the
82 residual Laurentide Ice Sheet in the Early Holocene. This leads to a pronounced provincialism in the
83 records for surface physical properties of the North Atlantic (e.g. de Vernal and Hillaire-Marcel, 2006;
84 Renssen et al., 2009; Eynaud et al., 2018). In addition to the long-term trend, the Holocene is marked
85 by short-term (from centennial to millennial-scales) climate variability which is evidenced in the
86 deep-water circulations (e.g. Bianchi and McCave, 1999; Kissel et al., 2013) and in various records
87 throughout the North Atlantic realm (Bond et al., 1997; 2001; Debret et al., 2007; Sorel et al., 2012;
88 Desprat et al., 2013; Smith et al., 2016).

89 Observations over the last few decades indicate that the North Atlantic subpolar gyre (SPG)
90 circulation varied both in strength (Curry and McCartney, 2001; Häkkinen and Rhines, 2004, Zunino
91 et al., 2017) and shape (Bersch, 2002; Hátún et al., 2005; Berx and Payne, 2017). These changes have
92 played a major role in the redistribution of heat and freshwater from the surface to intermediate
93 depths, thereby influencing the properties of waters entering the Nordic Seas (e.g. Häkkinen et al.,
94 2011). Variations in the SPG strength and shape have been revealed by recent observations (e.g.
95 Flatau et al., 2003; Häkkinen and Rhines, 2004). Modeling studies also highlight potential rapid
96 changes due to internal advective positive feedback (Born et al. 2016) with a strong impact on
97 climate (Sgubin et al. 2017). Furthermore, it has been shown that there is a close link between the
98 SPG intensity and winds controlled by the North Atlantic Oscillation (NAO), illustrated by the
99 downturn in the SPG circulation after the mid-1990s due to the shift from a high to low NAO index
100 (Häkkinen and Rhines, 2004; Häkkinen et al., 2011). Over the Holocene, it has been shown that
101 certain climatic anomalies are related to SPG intensity (e.g. Thornalley et al., 2009; Colin et al., 2010;
102 Copard et al., 2012; Moffa-Sanchez and Hall, 2017). For instance, the warm Medieval Climatic
103 Anomaly has been linked to a greater eastward extension of the SPG in the North Atlantic (Copard et
104 al., 2012; Moffa-Sanchez and Hall, 2017). Conversely, the Little Ice Age has been linked to the

105 eastward contraction of the mid-depth SPG (Copard et al., 2012; Moffa-Sanchez and Hall, 2017).
106 Overall, these studies have shown that SPG intensity has undergone millennial-scale changes
107 comprising variations in the heat and salt budgets at intermediate depths, which may have
108 influenced deep-water formation in the Nordic Seas. These millennial oscillations have been
109 associated with a change in the main wind field resembling short-term changes related to NAO
110 fluctuations (e.g. Copard et al., 2012; Wassenburg et al., 2016; Moffa-Sanchez and Hall, 2017; Zunino
111 et al., 2017) even though the atmospheric pressure gradient controlling the NAO is difficult to assess
112 at periodicities that exceed the instrumental record (Ortega et al. 2015). Reconstructing millennial
113 scale SPG dynamics (strength and position) during the Holocene is thus necessary to provide a more
114 comprehensive picture of the internal oscillations of the ocean and to constrain their link to the
115 North Atlantic deep ocean circulation and climate evolution.

116 In this study, we have measured the Nd isotopic composition (ϵNd) and radiocarbon content
117 ($\Delta^{14}\text{C}$) on precisely U-Th dated cold-water coral (CWC) fragments of *L. pertusa* and *M. oculata*, which
118 were collected from a location in the SW Rockall Trough on the eastern edge of the SPG. Upper to
119 mid-depth water mass circulation (above 900 m water depth) in the North Atlantic is characterized
120 by distinct ϵNd values ranging from -10 for water masses originating from the subtropical Atlantic to -
121 15 for water masses originating from the Labrador Sea (Lacan and Jeandel, 2004; Lambelet et al.,
122 2016; Dubois-Dauphin et al., 2017). Therefore, as CWC ϵNd reflects the ambient seawater ϵNd
123 (Copard et al., 2010; van de Flierdt et al., 2010), analyses of ϵNd in this archive can be used to
124 establish the origin of the water masses (e.g. Colin et al., 2010; Copard et al., 2012; Montero-Serrano
125 et al., 2011, 2013; Wilson et al., 2014). This has already permitted the tracking of past changes in the
126 intensity of the mid-depth SPG (mid-SPG) in the NE Atlantic at different time scales (Colin et al., 2010;
127 Copard et al., 2012; Montero-Serrano et al., 2011, 2013). In addition, $^{14}\text{C}/^{12}\text{C}$ measurements of U-Th
128 dated CWC are useful for determining the age of the surrounding intermediate water mass and for
129 reconstructing the ventilation age of the water masses (Adkins et al., 1998). The ventilation age
130 recorded in CWC depends on (i) ^{14}C of surface water related to atmospheric ^{14}C through air-sea gas
131 exchange and time of equilibrium between the atmosphere and water masses; (ii) the time that has
132 elapsed since water masses reaching the CWC mounds have become isolated from the atmosphere,
133 and (iii) vertical and horizontal mixing processes involving water masses with different ^{14}C signatures.
134 Consequently, the combination of $\Delta^{14}\text{C}$ and ϵNd values analyzed on CWC samples from the SW
135 Rockall Trough can produce a unique time series of intermediate water provenance and ventilation in
136 the NE Atlantic.

137

138 **2. Material and hydrological settings of the SW Rockall Trough**

139 **2.1 Cold-water coral samples**

140 The CWC investigated in this study were collected from sediment cores Corals2012-39PC
141 (55.452°N - 15.870°W, 742 m water depth; length 413 cm) and Corals2013-18PC (55.523°N -
142 15.644°W, 790 m water depth; length 397 cm) retrieved during the 64PE360-Corals 2012 and
143 64PE377-Corals 2013 cruises, respectively. In addition, one living sample of *L. pertusa* was collected
144 with a box corer (M2000-25BX) in the same area in August 2000 (55.542°N - 15.6475°W, 773 m water
145 depth) (Table 1 and Fig. 1). The three coring sites are located about 16 km apart on two individual
146 carbonated mounds at the SW Rockall Trough margin (named the Logachev Mounds after the
147 research vessel R/V Prof. Logachev) (Fig. 1; Table 1). The sediment cores are composed of biogenic
148 fragments (mainly CWC fragments of *M. oculata* and *L. pertusa*) within a carbonate matrix. The base
149 of core Corals2012-39PC (below 385 cm) is composed of a cemented carbonate layer associated with
150 dropstones and volcanic glass, which have been deposited during the last glaciation (Bonneau et al.,
151 2018). The hard-ground was not observed in core Corals2013-18PC. Only large and well-preserved
152 specimens of *M. oculata* and *L. pertusa* were sampled. In total, 55 samples were analyzed for their
153 Nd isotopic compositions and trace element concentrations (45 samples from core Corals2012-39PC
154 and 10 samples from core Corals2013-18PC). In addition, a subset of 25 CWC samples from core
155 Corals2012-39PC was selected for radiocarbon (¹⁴C) analyses.

156 The chronological framework of the CWC samples investigated in this study is based on U-Th
157 dating performed on the same samples and reported in Table 1 (Bonneau et al., 2018). CWC
158 collected from cores Corals2012-39PC and Corals2013-18PC have been dated to the periods between
159 0.25±0.01 and 10.75±0.05 ka BP and between 1.29±0.03 and 8.65±0.09 ka BP, respectively (Table 1).
160 Dating of CWC has revealed continuous growth of CWC throughout the Holocene. The vertical CWC
161 accumulation rates in core Corals2012-39PC, estimated from the relationship between the U-Th age
162 of CWC and its depth within the core, range between ~15 cm ka⁻¹ and ~150 cm ka⁻¹ (mean values of
163 33 cmka⁻¹) (Bonneau et al., 2018).

164

165 **2.2 Hydrological settings**

166 The cores that are the focus of this study, Corals2012-39PC and Corals2013-18PC, are located
167 in the upper part of the permanent thermocline between a thick, sub-surface water mass and the
168 colder, fresher LSW identified at 1200 m in the SW Rockall Trough (White and Dorschel, 2010). They
169 are influenced by intense near-bottom currents induced by diurnal tidal movement (Mohn et al.
170 2014). These currents are locally important for food particle supply, replenish nutrients and oxygen
171 to the CWC reef, and prevent the corals from being buried beneath sediments (Van Haren et al.,
172 2014). These diurnal oscillations induce movements of the thermocline and have been highlighted by

173 daily temperature variations of up to ~ 3 °C at the thermocline depth (Van Weering et al., 2003;
174 Duineveld et al., 2007; Mienis et al., 2007; Van Haren et al., 2014).

175 This setting is embedded in the large-scale cyclonic SPG circulation that characterizes the
176 North Atlantic from the surface to at least 700 m water depth (Lavender et al., 2005). The basin-scale
177 cyclonic SPG is defined by strong currents along the northern and western boundaries (in the
178 northern Iceland Basin, along the Reykjanes Ridge, and in the Irminger and Labrador Seas) and a
179 meandering North Atlantic Current (NAC) to the south (Fig. 2; Lavender et al., 2005). On the eastern
180 boundary of the basin, the northward flow of the mid-SPG splits, with one branch flowing north into
181 the Iceland Basin, and the other entering the Rockall Trough where an anticyclonic circulation is
182 observed (Lavender et al., 2005).

183 The Rockall Trough is marked by strong wintertime mixing of the near-surface layers. This
184 mixing reaches a depth of 500-700 m (Holliday et al., 2000) and up to 1000 m during severe winters
185 (New and Smythe-Wright, 2001). The resulting upper-layer water mass is a mixture of saline Eastern
186 North Atlantic Water (ENAW), which is transported poleward along the Western European margin
187 from the Northern Bay of Biscay by the Shelf Edge Current (SEC) (Ellett et al., 1986; Pollard et al.,
188 1996), and the fresher Modified North Atlantic Water (MNAW), which is carried by the North Atlantic
189 Current (NAC) (New and Smythe-Wright, 2001). The MNAW is formed by the mixing of the Western
190 North Atlantic Central Water (WNACW), which flows from the Caribbean Sea, with the Sub-Arctic
191 Intermediate Water (SAIW), which is derived from the Labrador Current.

192 Below the near-surface layer, the warm, saline Mediterranean Sea Water (MSW) propagates
193 northward along the European margin as an eastern boundary undercurrent, typically at depths of
194 1000-1200 m. Lozier and Stewart (2008) showed that the northward penetration of the MSW along
195 the eastern boundary of the North Atlantic is characterized by temporal variability in relation to the
196 NAO index and SPG expansion, implying the possibility of short-lived penetration of MSW into the
197 eastern Rockall Trough. The inflow of Labrador Sea Water (LSW) into the Rockall Trough,
198 characterized by a marked salinity minimum, occurs at water depths ranging from 1200 to 1900 m
199 (New and Smythe-Wright, 2001).

200 Hence, CWCs collected on the Logachev Mound area are ideally located to track past changes
201 in subpolar and subtropical water masses transported into the Rockall Trough.

202

203 **3. Methods**

204 The CWC samples were cleaned following the procedure described by Copard et al. (2010) in
205 order to remove any sources of contamination (sediment, organic matter, Fe-Mn coatings). First, the
206 surface of the CWC skeleton was mechanically cleaned using a diamond-blade saw in order to
207 remove any visible Fe-Mn oxide and hydroxide coatings. The CWC fragments were further cleaned

208 using diluted HCl and a series of ultra-sonic baths. Dried samples were crushed into a homogeneous
209 fraction (coarse sand grain-size) and finally divided into 3 aliquots of ~600 mg, ~50 mg and ~15 mg
210 for Nd isotopic composition, Nd concentration and radiocarbon analyses, respectively.

211

212 **3.1 Neodymium concentration and isotope analysis**

213 Nd concentrations were measured on cleaned coral samples (20–30 mg),
214 dissolved in supra-pure 3 N HNO₃, using a quadruple ICP-MSXseriesII (Thermo FisherScientific)
215 at the Laboratoire des Sciences du Climat et de l'Environnement (LSCE, France) and following the
216 procedure described in Montero-Serrano et al. (2013). Briefly, sample and standard solutions were
217 systematically adjusted at 100 ppm Ca through dilution, without further chemistry. To compensate
218 signal derivation by a few percent during a day, a standard (JcP-1) was run every five samples.
219 Instrumental bias was taken into consideration by a bracketing technique giving an analytical
220 uncertainty of 5% (2σ) for Nd concentrations.

221 Nd was purified from cleaned coral samples (~600 mg) using Eichrom TRU-Spec and
222 Ln-Spec resins following the detailed analytical procedures described in Copard et al. (2010). The
223 ¹⁴³Nd/¹⁴⁴Nd ratios were analyzed using the ThermoScientific Neptune^{Plus} Multi-Collector Inductively
224 Coupled Plasma Mass Spectrometer (MC-ICP-MS), hosted at the LSCE. The bracketing correction
225 method outlined by Dubois-Dauphin et al (2017) was applied. For the Nd isotope analyses, sample
226 and standard concentrations were matched at 4 to 10 ppb. Mass-dependent fractionation was
227 corrected by normalizing ¹⁴⁶Nd/¹⁴⁴Nd to 0.7219 and applying an exponential fractionation law. During
228 the analytical sessions, every set of two samples was bracketed by analyses of the La Jolla Nd
229 standard solution, which is characterized by certified values of 0.511858±0.000007 (Lugmair et al.,
230 1983). The offset value between results and certified values of La Jolla was
231 inferior to 0.4 εNd units for all of the analyses presented in this study. The
232 analytical errors reported herein correspond to the external two sigma standard deviation (based on
233 repeat analyses of the La Jolla standard for the different analytical sessions). For a few coral samples
234 the internal two sigma standard deviation is used as it is higher than the external two sigma standard
235 deviation. The analytical errors obtained ranged from 0.2 to 0.5 epsilon units (Table 1). The analytical
236 blank values for Nd were <4 pg, which represents less than 0.1% of the minimum Nd yield from CWC
237 skeletons used in this study. As a result no blank correction was applied.

238

239 **3.2 Radiocarbon dating**

240 Approximately 15 mg samples of clean coral powder were treated with 0.01N HNO₃ for 15
241 min, rinsed with MilliQ water and dried. Each sample was then converted to CO₂ in a semi-
242 automated carbonate vacuum line (Tisnérat-Laborde et al., 2001). The CO₂ was reduced into graphite

243 using hydrogen in the presence of iron powder. Accelerator mass spectrometry (AMS) ^{14}C dating was
244 performed using the AMS equipment at ARTEMIS (LM14C - Saclay, France, Cottureau et al., 2007).
245 Blanks were obtained on a *L. pertusa* sample older than 100 000 yrs (U-Th age), and yielded a mean
246 activity of $0.0022 \text{ F}^{14}\text{C}$ (apparent age older than 48 000 yrs). Radiocarbon results are reported as
247 conventional ^{14}C ages in yr BP and in terms of $\Delta^{14}\text{C}$ defined as the per mil (‰) deviation of a sample's
248 $^{14}\text{C}/^{12}\text{C}$ ratio from that of a 19th century wood standard after correction of blank, fractionation and
249 correction for decay obtained from U-Th ages (cal yr BP) (Stuiver and Polach, 1977) (Table 1). The
250 marine reservoir ages (R) were calculated as the offset of the measured ^{14}C age of coral from
251 contemporaneous atmospheric ^{14}C age based on IntCal13 (Reimer et al., 2013). The reservoir offsets
252 (ΔR) were estimated by calculating the difference between the marine radiocarbon age of surface
253 water based on the Marine13 curve and the ^{14}C age of coral at that same time.

254

255 **4. Results**

256 The Nd concentrations obtained for fossil deep-sea corals of both cores, Corals2012-39PC
257 and Corals2013-18PC (*L. pertusa* and *M. oculata*), range from 7.1 to 55.0 ppb (Table 1). Such range of
258 Nd concentrations are in accordance with Nd concentrations previously obtained on CWC from the
259 SW Rockall Trough (7.3-184.3 ppb) (Colin et al., 2010; Copard et al., 2012). Copard et al. (2010) have
260 shown that such Nd concentration ranges are not linked to the presence of Nd contamination from
261 manganese-oxide and iron hydroxide coatings. Struve et al. (2017) have recently suggested that
262 authigenic phosphate phases (hypothesized apatites) could be important carriers of skeletal Nd in
263 cleaned CWC. This could explain the large range of Nd concentrations observed on the aragonite
264 skeleton.

265

266 One living sample of *L. pertusa* collected in box core M2000-25BX in August 2000 provides an
267 ϵNd of -13.6 ± 0.4 (Table 1), which is indistinguishable from ambient seawater ϵNd of -13.5 ± 0.2
268 (Dubois-Dauphin et al., 2017; Fig. 2a). This finding confirms that cleaned CWC skeletons record
269 ambient seawater ϵNd (Copard et al., 2010; van der Flierdt et al., 2010).

270 ϵNd obtained on *L. pertusa* and *M. oculata* collected from Corals2012-39P and Corals2013-
271 18PC range from -12.2 ± 0.3 to -16.6 ± 0.4 (Table 1). The coral ϵNd records display a decreasing trend
272 (from -13.2 ± 0.2 to -14.9) between 10.7 and 9.1 ka BP. The time interval from 8.8 to 6.8 ka BP is
273 associated with low ϵNd values (from -14.3 ± 0.5 to -15.4 ± 0.3) and is followed by a progressive
274 increase of the ϵNd from -15.4 ± 0.3 to -13.3 ± 0.2 during the time interval between 7 and 5 ka BP. One
275 coral sample dated to 8.6 ka BP displays an unradiogenic ϵNd value of -16.6 ± 0.4 . The interval
276 spanning the last 5 ka BP is marked by more radiogenic ϵNd values ranging from -14.3 ± 0.2 to
277 -12.2 ± 0.3 .

278 Coral ϵNd values obtained in this study are combined with published data from CWC
279 fragments collected from several carbonate mounds within a limited area of the SW Rockall Trough
280 and at a similar water depth (between 725 and 790 m water depth) (Colin et al., 2010; Copard et al.,
281 2012) (Fig. 2). This new dataset, presented in Figure 3, is composed of 100 coral ϵNd values spanning
282 the last 11 ka. The new results reproduce the variability observed in the previous coral ϵNd record
283 (Colin et al., 2010), thus confirming the large-scale character of Nd isotope changes in the SW Rockall
284 Trough. The new data imply a roughly twofold increase of time resolution compared to existing
285 Holocene coral ϵNd data from the Rockall Trough. Nevertheless, the time resolution is variable
286 throughout the Holocene, with lower resolution before 6 ka BP due to marked changes in CWC
287 abundance at the SW Rockall Trough margin (Bonneau et al., 2018). The Early Holocene (between 8.8
288 and 6.8 ka BP) stands out as period with unradiogenic ϵNd values. Our new CWC ϵNd stack confirms
289 a mid-Holocene transition from subpolar to subtropical Nd isotopic signatures (Colin et al., 2010), but
290 shows that this transition was rather smooth (lasted ~ 1.5 ka) and started just after 7 ka BP.
291 Superimposed on this long-term trend, millennial variations in the coral ϵNd of up to 2.5 epsilon units
292 can be identified, especially during the Late Holocene (Fig. 3).

293

294 The $\Delta^{14}\text{C}$ results obtained from CWC in this study (Table 1) plotted against U-Th dates, are
295 compared with the IntCal13 and Marine13 $\Delta^{14}\text{C}$ calibration curves (Reimer et al., 2013) from 0 to 10.1
296 ka BP and range from -72 ± 4 ‰ to $+70 \pm 9$ ‰ (Fig. 3). The decrease of CWC-derived $\Delta^{14}\text{C}$ observed
297 between the Early and Late Holocene is consistent with the Marine13 curve and is mainly linked to
298 the decrease of atmospheric $\Delta^{14}\text{C}$. Most of the CWC-derived $\Delta^{14}\text{C}$ values match the Marine13 $\Delta^{14}\text{C}$
299 calibration curve.

300 Eight CWC-derived $\Delta^{14}\text{C}$ values differ from the Marine13 calibration curve. These differences
301 are significant at 2 sigma according to the Chi-square test at a 95% significance level. One sample
302 (CWC sample from core Corals2012-39PC at 137.5 cm) yielded a higher $\Delta^{14}\text{C}$ value at 3.67 ka BP (-1.6
303 ± 5 ‰), which corresponds to an enrichment of 20 ‰ compared to the Marine13 curve at this time ($-$
304 21.5 ± 3.2 ‰). Seven CWC samples (CWC samples from core Corals2012-39PC at 38.5, 87.5, 109.5,
305 120.5, 169.5, 227.5 and 241.5 cm), whose respective ages are centered at 6.1, 4.4, 2.7, 2 and 0.7-1 ka
306 BP, are characterized by $\Delta^{14}\text{C}$ values that are lower than the Marine13 $\Delta^{14}\text{C}$ calibration curve (red
307 arrows in figure 3). In two cases, at about 6.1 ka and 2.7 ka BP, two CWC samples, results confirm this
308 decreasing trend in $\Delta^{14}\text{C}$. The lower $\Delta^{14}\text{C}$ (Fig. 3) values correspond to an older ^{14}C age for water and
309 thus reflect an increase in the marine reservoir age, which reaches up to 519 ± 35 years (Table 1). The
310 marine reservoir ages of these seven CWC samples are characterized by a mean value of 475 years
311 ($n=7$), which is higher than the mean reservoir age obtained on the other CWC samples investigated
312 in core Corals2012-39PC (367 years; $n=18$).

313

314 5. Discussion

315 5.1. Oceanographic controls on Rockall Trough seawater ϵNd

316 Two recent studies have provided new insights into the signature of the seawater Nd isotopic
317 composition in the eastern (Dubois-Dauphin et al., 2017) and western (Lambelet et al., 2016) North
318 Atlantic. Seawater ϵNd is found to be homogeneous (between -14 and -15) throughout the SPG for
319 the upper 1500 m water depth, suggesting a similar origin for the surface to intermediate subpolar
320 waters (Lambelet et al., 2016; Dubois-Dauphin et al., 2017). Likewise, the upper 1500 m of the
321 Rockall Trough water column are characterized by homogenous Nd isotopic compositions (from -
322 14.0 ± 0.2 to -13.3 ± 0.2 ; Dubois-Dauphin et al., 2017) (Fig. 2b). Thus, subsurface to intermediate water
323 masses are largely indistinguishable on the basis of their Nd isotope characteristics. Nevertheless,
324 when compared to ϵNd values for the SPG, those from the Rockall Trough seem to be influenced by a
325 more radiogenic end-member (Fig. 2b). Recent results obtained by Dubois-Dauphin et al. (2017)
326 along a North-South transect at 25°W , spanning the SPG and sub-tropical gyre (STG) interface,
327 highlight that water masses carried by the NAC from the surface to ~ 900 m water depth have an
328 unradiogenic Nd isotopic signature (from -15.1 ± 0.3 to -13.8 ± 0.3 ; Dubois-Dauphin et al., 2017). This
329 implies that MNACW ϵNd is strongly influenced by SAIW originating from the Labrador Sea, with
330 dissolved Nd concentrations that are twice those of WNACW in the western STG (Lambelet et al.,
331 2016). Consequently, this indicate that coral ϵNd record of the SW Rockall Trough cannot be
332 interpreted by a modification of proportions of water masses from SPG and STG waters transported
333 in the NAC as it was proposed by Colin et al. (2010). The more radiogenic end-member ϵNd observed
334 in the Rockall Trough can be only due to the contribution of the radiogenic NEAW which is
335 transported northward by the SEC (ϵNd signature of -11.3 ± 0.4 in the Bay of Biscay, 1σ , $n=6$; Rickli et
336 al., 2009; Copard et al., 2011). In addition, at greater water depth (around 950 m), the radiogenic
337 MSW (ϵNd around -10.6 ± 0.3 ; Dubois-Dauphin et al., 2017) is also carried northward along the
338 western European margin to the Porcupine Seabight (Dubois-Dauphin et al., 2017). The presence of
339 MSW in the Eastern Rockall Trough has been shown to respond to the dynamics of the mid-SPG, such
340 that mid-SPG contraction induces northward penetration of MSW (Lozier and Stewart, 2008). Hence,
341 by vertical mixing and recirculation (Holliday et al., 2000; Lavender et al., 2005) the MSW could also
342 induce a more radiogenic ϵNd signature to intermediate water masses of the Rockall Trough.

343 In summary, ϵNd values of intermediate waters at the SW Rockall Trough result mainly from
344 the mixing of unradiogenic MNACW (ϵNd from -15.1 ± 0.3 to -13.8 ± 0.3), which was carried by the
345 NAC, and radiogenic ENAW ($\epsilon\text{Nd} \approx -11.3$) and MSW (-10.6 ± 0.3), which propagated northward along
346 the European margin. Consequently, mid-depth coral ϵNd records from the SW Rockall Trough can be

347 linked to the spatial evolution of the interface between the mid-SPG waters and the ENAW/MSW
348 carried northward by the European slope current. An eastward expansion (westward contraction) of
349 the mid-SPG will induce a decrease (increase) in the relative proportion of the ENAW/MSW resulting
350 in unradiogenic (radiogenic) Nd isotopic signatures for intermediate water masses in the Rockall
351 Trough.

352

353 **5.2. The reorganization of mid-depth circulation in the North Atlantic during the Holocene**

354 The coral ϵNd record for the SW Rockall Trough displays significant variations, ranging from -
355 12.2 ± 0.3 to -16.6 ± 0.4 , over the Holocene, marked by a shift towards less mid-SPG influence between
356 7 and 5 ka and punctuated by abrupt events. Howe et al. (2016) suggested that increased detrital
357 input from the Laurentide Ice Sheet in the Early Holocene (Kurzweil et al., 2010) induced a
358 modification of the ϵNd of deep-water masses of the North Atlantic, which is well marked in cores
359 located at water depths greater than about 2500 m. Holocene seawater ϵNd records for the
360 intermediate water of the Labrador Sea are not available but we cannot rule out the possibility that
361 the ϵNd of the SPG had also been modified at the beginning of the Holocene. Nevertheless,
362 variations in the coral ϵNd of the SW Rockall Trough cannot be explained solely by changes in the Nd
363 isotopic composition of the water masses induced by lithogenic input from the Laurentide Ice Sheet
364 for several main reasons. First, whereas melting of the Laurentide Ice Sheet was greater in the early
365 Holocene than during the mid-Holocene, CWC ϵNd from the SW Rockall Trough is slightly more
366 radiogenic in the early Holocene than the mid-Holocene (time interval from 8.8 to 6.8 ka BP). Second,
367 observations of unradiogenic ϵNd from the early Holocene occur mainly at abyssal depths (around
368 4500 m, Howe et al., 2016), and are less clear at depths corresponding to NADW. Radiocarbon results
369 suggest fairly well-ventilated water (predominantly thermocline down to modern LSW) and hence
370 indicate little influence from abyssal ($\Delta^{14}\text{C}$ depleted) waters at our site. The variability we observe in
371 coral ϵNd is thus much more likely to be related to changes in the upper water column which,
372 according to Howe et al. (2016), does not seem to be as affected by this detrital input. Third, late
373 Holocene variations in coral ϵNd are of a similar magnitude (up to 2.5 epsilon units) as the
374 early/middle Holocene, but these late Holocene variations cannot be explained by lithogenic input
375 because the final demise of the Laurentide Ice Sheet occurred earlier (around 6.8 ka, Carlson et al.,
376 2008). Finally, the studied site is located on the eastern edge of the SPG implying marked variations
377 in the relative proportion of water masses from the SPG flowing into the SW Rockall Trough.

378 Therefore, based on the assumption that the sources of Nd for the water masses are
379 constant throughout the Holocene (i.e. taking into account the ϵNd and salinity of pure end-member
380 mid-depth waters from the SPG and STG), it is possible to estimate a salinity range at mid-depth of

381 about 1 PSU (from 35.0 to 36.0). This is in agreement with the range of Holocene sub-thermocline
382 salinity estimated from coupled Mg/Ca- $\delta^{18}\text{O}$ measurements obtained for planktic foraminifera from
383 the South Iceland Rise by Thornalley et al. (2009). Assuming constant ϵNd end-members for the mid-
384 depth SPG (-15) and STG (-11), we can estimate that the contribution of the SPG to the Rockall
385 Trough intermediate waters varied from around 20% (80% STG) to 100% (no influence of the STG)
386 over the last 11 ka. Our results confirm that the Early Holocene, between 8.8 and 6.8 ka BP, was
387 marked by an eastward expansion of the mid-SPG, compared to present-day conditions, in
388 agreement with Thornalley et al. (2009) (Fig. 4c). In addition, planktonic faunal assemblages (such as
389 *G. inflata* % reported in Fig. 4d) and oxygen stable isotopic records obtained on cores located in the
390 surroundings of the Faroe Islands suggest an eastward expansion of the NAC and reduced
391 contributions of warm saline waters from the STG to the Nordic Sea from 8 to 6 ka BP (Staines-
392 Urias et al., 2013). These results suggest that, between 8 and 6 ka BP, the upper layer of the NE
393 Atlantic was characterized by cold, fresh water implying an eastward extension of the SPG, which, at
394 mid-depth, is consistent with the high contribution of water originating from the mid-SPG estimated
395 on the basis of CWC ϵNd in the SW Rockall Trough.

396 Most of the $\Delta^{14}\text{C}$ values for CWC from the SW Rockall Trough are similar to those of the
397 Marine13 $\Delta^{14}\text{C}$ calibration curve indicating that the water surrounding the corals remained relatively
398 well ventilated, particularly before 6 ka BP (Fig. 3). The Li/Mg temperature record of the mid-depth
399 waters obtained by Bonneau et al. (2018) on the same CWC samples as those investigated in this
400 study is reported in the Figure 3 with a view to comparing it to ϵNd and $\Delta^{14}\text{C}$ records (Fig. 3). Whilst
401 the ϵNd record indicated large-amplitude changes in the water mass origin, Li/Mg reconstructed
402 temperatures exhibit relatively low variability (from 7 to 9°C) during the Holocene (about 2°C). This
403 may imply that winter convection was at least as deep as it is now (between 500 to 1000 m water
404 depth; Holliday et al., 2000; New and Smythe-Wright, 2001) without drastic changes in the vertical
405 temperature distribution. Nevertheless, the slightly lower Li/Mg temperature (7.0 ± 0.9 to 7.9 ± 0.9 °C)
406 observed in the Early Holocene (before 6 ka BP), compared to the Late Holocene (7.4 ± 0.9 to 8.8 ± 0.9
407 °C) (Bonneau et al., 2018), may be compatible with a greater proportion of colder SAIW (Pollard et
408 al., 1996) as is observed in the CWC ϵNd record for the SW Rockall Trough.

409

410 **5.3. Implications of the mid-Holocene long-term variations in SPG intensity on ISOW** 411 **formation and on ϵNd of the deep-ocean**

412 Only a handful of studies have addressed the evolution of the deep-limb of the AMOC in the
413 subpolar North Atlantic during the Holocene (e.g. Bianchi and McCave, 1999; Praetorius et al., 2008;
414 Kissel et al., 2013; Thornalley et al., 2013; Howe et al., 2016; Ayache et al., 2018). These studies have
415 identified several short-term variations in the intensity of North Atlantic deep-water, with amplitude

416 and timing depending greatly on the region investigated and the proxies used. Thornalley et al.
417 (2013) presented a stacked sortable silt record from the Reykjanes Ridge, which is indicative of the
418 Iceland–Scotland overflow strength during the Holocene. In figure 4b, we present the stacked
419 sortable silt record for comparison with the CWC ϵNd record (Fig. 4f). Results reveal a decrease in
420 Iceland–Scotland Overflow Water (ISOW) strength during the Mid- to Late Holocene, with maximum
421 Iceland–Scotland overflow strength occurring at ~ 7 ka; this maximum corresponds to low CWC ϵNd
422 values in the SW Rockall Trough. This suggests a link between the position of the mid-SPG and the
423 strength of the Iceland–Scotland overflow, which forms an important component of the deep branch
424 of the AMOC (Figs. 4b and 4f). In addition, to overcome complex spatial and temporal
425 variability of the classic SST and SSS proxies strongly influenced by the orbital-induced insolation
426 trend and the disintegration of the residual Laurentide Ice Sheet in the Early Holocene, Aya che et
427 al (2018) have extracted significant AMOC mode variations based on a
428 relatively large compilation of paleo-ocean SST data (HAMOC database;
429 Eynaud et al., 2018) from the North Atlantic Basin. They have identified a period
430 of enhanced AMOC mode in the period between 8 and 6.5 ka BP, which is
431 associated with a greater extension of the mid-SPG (low coral ϵNd). This period
432 of intensified AMOC is followed by a general weakening trend from around 6–7
433 ka to 2 ka, which is also consistent with the progressive westward contraction
434 and weakening of the mid-SPG observed in the Rockall Trough after 7 ka BP.

435

436 Before ~ 7 ka BP, water masses originating from the Labrador Sea were advected at lower
437 water depths (Hillaire-Marcel et al., 2013) and may have been linked to the surface and sub-surface
438 layers. These water masses may have entrained unradiogenic Nd isotopic composition from the
439 surface water of the Labrador Sea, which is marked by boundary exchange (Lambelet et al., 2016). At
440 intermediate depth (above 1000 m), we suggest that an intense SPG circulation would have brought
441 a greater volume of seawater into contact with the unradiogenic sediments of the Labrador Sea and
442 the Greenland margins (Jeandel et al., 2007). Thus, it is possible that such an intensification of SPG
443 circulation could also have increased the lithogenic Nd input from the boundary exchange. Such
444 increase of lithogenic Nd input needs then to be taken into account in the budget of Nd input to the
445 North Atlantic.

446 Today, the SPMW and LSW are entrained and mixed with the ISOW within the Iceland–
447 Scotland overflow. This entrainment of surrounding waters characterized by unradiogenic ϵNd values
448 modifies the pure ISOW signature (-8.2 ± 0.6 ; Lacan and Jeandel, 2004) to a value of -10.5 ± 0.5 after
449 overflowing the Iceland–Faroe ridge (Dubois-Dauphin et al., 2017). Based on ϵNd values and salinity,
450 Dubois-Dauphin et al. (2017) have estimated that ISOW flowing in the Iceland Basin is composed of

451 ~60% pure ISOW and of 40% entrained surrounding waters (LSW $\epsilon_{Nd} = -13.9 \pm 0.4$; and Modified North
452 Atlantic Water -MNAW $\epsilon_{Nd} = -13.1 \pm 0.3$; Lacan and Jeandel, 2005). The time interval from 8.8 to 6.8
453 ka BP is associated with an eastward extension of the mid-depth (CWC ϵ_{Nd} of the SW Rockall Trough)
454 and surface SPG (Thornaley et al., 2009; Staines- Uria, 2013). This increased flow of surface and sub-
455 surface waters, characterized by unradiogenic Nd isotopic signatures, in the Iceland basin may then
456 have been entrained by an intense ISOW (Thornaley et al., 2013). This thus would have contributed,
457 along with the Nd lithogenic input from detrital material of the Laurentide Ice Sheet (Howe et al.,
458 2016), to the unradiogenic Nd signature observed for the NADW in the Early Holocene in several
459 deep sites of the North Atlantic (Roberts et al., 2010; Böhm et al., 2015; Howe et al., 2016; Lippold et
460 al., 2016). Further modeling studies of the Nd isotopic composition are required in order to better
461 evaluate the lithogenic Nd inputs to the subpolar water and its impact on the Nd isotopic signature of
462 the North East Atlantic Deep Water (NEADW) during the Early Holocene through the entrainment
463 of SPMW deriving from an intensified SPG, into the ISOW.

464

465 **5.4. Millennial scale variations of the mid-SPG**

466 The coral ϵ_{Nd} record shows several short periods of eastward extension (low coral ϵ_{Nd}
467 values), which occurred approximately every 1500 ± 500 years, centered around 10.1, 8.6, 5.5, 4.5,
468 2.5, 1.8 and 0.8 ka BP (Fig. 3). For the Late Holocene, some of these time intervals (centered at 6.1,
469 4.4, 2.7 and 0.7-1 ka BP) are also associated with $\Delta^{14}C$ values that are 14 to 20‰ lower than the
470 Marine13 curve (Fig. 3). These differences, which are statistically significant, correspond to higher
471 marine reservoir ages that are between 430 ± 30 and 519 ± 35 yrs (Table 1). This millennial-scale
472 climate variability during the Holocene was first pointed out by Bond et al. (1997 and 2001)
473 (1470 ± 500 -year cycles) and is related to a combination of solar activity (1000 and 2500-year cycles)
474 and AMOC internal variations (1500-year cycles), which are particularly well marked for the last 5 ka
475 (Debret et al., 2007). This late Holocene time interval is also associated with large variations in the
476 CWC ϵ_{Nd} record obtained on the SW Rockall Trough (Figure 3).

477 Several processes may be responsible for the recorded $\Delta^{14}C$ variability. First, the $\Delta^{14}C$
478 variability may also be directly related to the changes in the relative advection of thermocline water
479 from the SPG (i.e. SAIW) and STG. Modern observations (pre-industrial ^{14}C estimates in GLODAP v1.1;
480 Key et al., 2004) and modeling studies covering the Holocene (Franke et al., 2008; Butzin et al., 2017)
481 indicate the existence of a considerable offset in radiocarbon contents between these sources. The
482 poorly stratified SPG is thoroughly mixed by deep convection resulting in the upper water having a
483 bomb-corrected $\Delta^{14}C$ of -65‰ (Key et al., 2004) and reservoir ages of around $550^{14}C$ years (Butzin et
484 al., 2017), which is slightly depleted/older with respect to modern global surface ocean (-58‰ and
485 $470^{14}C$ years of Marine13; Reimer et al., 2013). In contrast, the stratified thermocline water in the

486 STG is well equilibrated to the atmosphere, with modern surface $\Delta^{14}\text{C}$ of -45‰ (Key et al., 2004) and
487 reservoir ages of 300 ^{14}C years (Butzin et al., 2017) which is less depleted/younger than the average
488 modern surface ocean (Marine13; Reimer et al., 2013). Furthermore, in the modern ocean the
489 anomalous ^{14}C enriched water from the upper STG is observed to follow the NAC towards the
490 European Margin and the Rockall Trough (Key et al., 2004). The variability of $\Delta^{14}\text{C}$ relative to the
491 Marine13 curve observed in Rockall Trough coral record is of a similar magnitude as the offset
492 between the gyres and thus may simply be caused by changes in the relative advection of STG and
493 SPG thermocline waters as indicated by the coeval ϵNd excursions (Fig. 3). Analogous radiocarbon
494 variability related to dynamic changes in the STG influence has also been recorded in several
495 shallower records downstream from the Rockall Trough (Ascough et al., 2009; Eiríksson et al., 2011;
496 Wanamaker Jr et al., 2012).

497 However, given the position of the corals near the base of the modern mixed layer (Holliday
498 et al., 2000; New and Smythe-Wright, 2001), local variability in the mixed layer depth throughout the
499 Late Holocene may have driven the observed variability in radiocarbon contents. Over the last
500 decades, mixed layer depth has varied significantly in the Rockall Trough region and appears to be
501 principally forced by the position of the NAC and hence the extent of the SPG (Carton et al., 2008).
502 Wade et al. (1997) proposed that the SAIW, which is characterized by low salinity and temperatures
503 ($4^\circ\text{C} - 7^\circ\text{C}$, 34.7-34.9 PSU, Pollard et al., 1996) underneath the Central Water, plays an important role
504 in the stability of the NE Atlantic water column. They demonstrated that the presence of relatively
505 'pure' SAIW in the NE Atlantic prevents deep winter mixing due to enhanced stratification. In
506 contrast, the absence of a well-defined SAIW layer induces a weaker density gradient allowing deep
507 winter convection (Ellett et al., 1986; Holliday et al., 2000). As such, the incursion of SAIW in the
508 Rockall Trough during the Late Holocene episodes of SPG expansion, as indicated by unradiogenic
509 coral ϵNd , may have shoaled the mixed layer depth. The water masses around the studied CWC can
510 be then isolated from upper waters resulting in the observed $\Delta^{14}\text{C}$ depletion with respect to the
511 surface ocean (Marine13 $\Delta^{14}\text{C}$ calibration curve; Reimer et al., 2013).

512 Finally, the observed radiocarbon variability may also have originated from perturbations of
513 the convection processes in the Labrador Sea. Between 1885 and 1950 AD, $\Delta^{14}\text{C}$ fluctuations of 10‰
514 to 18‰ occurred in the surface of the NE Atlantic Ocean (Tisnérat-Laborde et al., 2010). They may
515 have originated in the Labrador Sea through the increased vertical convection and formation of
516 SPMW due to reinforced wind strengths in the high latitudes of the North Atlantic, as suggested by
517 Reverdin et al. (1997) for SSS anomalies. The surface freshwater intrusion in the Labrador Sea inhibits
518 deep convection as was observed during the great salinity anomaly in the 1960s (Reverdin et al.,
519 1997; Swingedouw et al. 2015). The demise of the Laurentide Ice Sheet ended at around 6.8 ka BP,
520 but several meltwater peaks are identified during the Late Holocene in the Labrador Sea and are

521 mainly attributed to the melting of the Greenland Ice Sheet (Fig. 4a) (Hoogakker et al., 2015). Taking
522 into consideration the error in age models, the $\delta^{18}\text{O}_{\text{seawater}}$ curve at Orphan Knoll (Labrador Sea) (Fig.
523 1) displays several events of SSS increase at about 4.2, 2.4, 1.8, and 1.0 ka BP (Hoogakker et al., 2015)
524 which correspond to an extension of the SPG (low coral ϵNd) (Figs. 4a and 4f). Such results are
525 consistent with a scenario where less freshwater input by the Eastern Greenland current to the
526 Labrador Sea could be responsible for a millennial perturbation of the LSW formation and SPG
527 circulation (Moffa-Sanchez and Hall, 2017). In this scenario, a decrease in freshwater input to the
528 Labrador Sea would enhance vertical convection resulting in relatively ^{14}C depleted upper SPG waters
529 (SAIW). Episodes of eastward extension of the SPG (low CWC ϵNd) may thus be responsible for the
530 advection of more ^{14}C -depleted SAIW in the SW Rockall Trough during the Late Holocene, in
531 agreement with our radiocarbon results. Further investigations will be necessary to constrain the
532 variability of the reservoir age within the subpolar Atlantic during the Holocene.

533 In summary, the relatively small $\Delta^{14}\text{C}$ variations recorded in the Rockall Trough do not allow
534 us to distinguish between these potential processes. Nonetheless, all processes highlight the
535 important role of the SPG dynamics which is consistent with the ϵNd record.

536

537 The intervals of eastward extension of the mid-SPG are not systematically correlated to those
538 inferred from SST and SSS reconstructions for surface and sub-surface water masses to the south of
539 Iceland by Thornalley et al. (2009); exceptions are the Early Holocene event involving the significant
540 eastward extension of SPG (between 9 and 6.8 ka BP) and the short events labeled 0, 1 and 2 in
541 figures 4c and 4f. Whereas SSS only records the evolution of inter-gyre mixing linked to the shape
542 and intensity of the SPG, the mid-depth ϵNd record also tracks the northward intrusion of the ENAW
543 by boundary current along the European margin. Nevertheless, both records are highly sensitive to
544 the shape and intensity of the SPG but they do not strictly match each other. According to Miettinen
545 et al. (2012) a regional SST see-saw exists between the Norwegian Sea and the subpolar North
546 Atlantic and is clearly identified for the LIA – MWP climate periods. This was associated with a
547 marked rerouting of the eastern and western branches of the surface limb of the AMOC (NAC) in the
548 NE Atlantic (Miettinen et al., 2012) giving rise to SST and SSS records which are not synchronous.
549 Nevertheless, the lack of regional synchronicity for short events, especially when they are derived
550 from SST records, has been pointed out in many studies (e.g. Larsen et al., 2013) and could be linked
551 to characteristics of the NAO itself (Miettinen et al., 2012).

552 The time interval from 1 to 0.68 ka BP, which is marked by a strong eastward extension of
553 the SPG, has been associated with the warm Medieval Climatic Anomaly and a subsequent
554 intensification of the surface limb of the AMOC (Copard et al., 2012; Wanamaker et al., 2012; Ortega

555 et al., 2015) (Fig. 4). The westward contraction of the weak SPG observed thereafter (between 0.68
556 and 0.2 ka BP) is coeval with the cold period of the Little Ice Age and may be linked to reduced AMOC
557 intensity. Furthermore, compared to the ϵNd values of CWC dated between 90 and 800 yrs BP (-12.8
558 on average; $n=19$), the unradiogenic ϵNd of modern CWC (dated to the last 50 years; -13.6 on
559 average, $n=15$) as well as that of seawater (Fig 2a) (-13.5 at 700 m in the Rockall Trough; Dubois-
560 Dauphin et al., 2017) suggest a greater extension of the mid-SPG during the last century (labeled as
561 event 0 in Figure 4f). This could be associated with global warming induced by considerable
562 anthropogenic greenhouse gas emissions and is consistent with an enhanced heat and salinity flux to
563 the Arctic Ocean supplied by the NAC (Spielhagen et al., 2011). The new ϵNd value record indicates
564 several other periods of major eastward extension of the SPG suggesting a strong millennial
565 variability in the intensity of the surface limb of the AMOC (Fig. 4f). The time interval between 2.1
566 and 1.7 ka BP is characterized by unradiogenic ϵNd values (-14.1 ± 0.3) and corresponds to a warm
567 period in Europe known as the Roman Warm Period (Moffa-Sánchez et Hall, 2017). The cold periods
568 corresponding to the Dark Ages Cold Period, between 1.2 and 1.7 ka BP, are associated with more
569 radiogenic ϵNd values (-12.4 to -13.4). The climatic history of Northern Europe before the Roman
570 Warm Period is less clear but the NE Atlantic region did witness a cold event at 2.8-3.5 ka BP
571 (Mayewski et al., 2004; Moffa-Sánchez et Hall, 2017), which is also associated with radiogenic ϵNd
572 values (-11.9). This extends the chronological framework of the results previously obtained by Copard
573 et al. (2012) on CWC of the SW Rockall Trough and confirms that warm intervals in the NE Atlantic
574 region and in surrounding land masses are associated with greater eastward extension of the SPG,
575 suggesting an intensification of the surface limb of the AMOC. We are not able to provide any
576 additional explanation for the origin of this oceanic variability, which may result from atmospheric
577 reorganization. However, unlike the large-scale mid-Holocene reorganization that could have
578 induced a northward shift of the westerlies and a greater extension of the SPG between 8.8 and 6.8
579 ka BP, millennial variability in the intensity and position of the westerlies remains a matter of debate.

580

581 **5.5. Coupling of the mid-SPG dynamics with the low latitudes**

582 The millennial variations in the coral ϵNd record obtained on the SW Rockall Trough are not
583 well correlated with other SST and SSS records for the NE Atlantic. However, the ϵNd data
584 correspond to SST variations observed at ODP site 658 in the eastern subtropical Atlantic off the
585 coast of Mauritania (deMenocal et al., 2000) (Fig. 4e). In addition to the mid-Holocene transition
586 observed in both records, the decrease in the coral ϵNd values suggests a greater eastward extension
587 of the mid-SPG, which is associated with a decrease of the SST at ODP site 658. Hydrographically,
588 these cold SST events could be linked either to an intensification of the Canary Current carrying cold

589 waters southwards along the western European margin or to an increase in regional upwelling
590 (deMenocal et al., 2000). This link between the high latitude climate and the variations of SST and
591 SSS in the Canary Current has been found as a fingerprint of freshwater release around Greenland in
592 so-called hosing experiments (Swingedouw et al., 2013). This teleconnection is described as being
593 related to the ocean circulation, the Canary Current being described as a freshwater leakage for low
594 SSS from the SPG to the STG. This link is generally consistent with the teleconnection between North
595 African and Northern Europe climates (e.g., Wassenburg et al., 2016). Cold events centered at 0.4,
596 0.8 and 2.6-3.3 ka BP for ODP Site 658 have been also observed in core OC437-7 24GGC located in a
597 northern position in the Eastern Tropical Atlantic (Morley et al., 2014). These events are associated
598 with a slight decrease in salinity, suggesting that this is not a regional process and may reflect a
599 greater southward heat transfer from high to low latitudes. Thus, we can infer from this correlation
600 between coral ϵNd values and low latitude SST records that time intervals of intensified mid-SPG and
601 AMOC might correspond to a decrease in the latitudinal thermal gradient in the North Atlantic. This
602 would have led to a major atmospheric reorganization during the mid-Holocene, with a northward
603 shift of the Hadley cell circulation and of the Inter Tropical Convergence Zone (ITCZ), as suggested by
604 modeling studies (Swingedouw et al., 2009; Marshall et al., 2014).

605

606 **6. Conclusions**

607 In this paper, we have analyzed the radiocarbon content and the Nd isotope composition of
608 precisely U-Th dated fragments of *L. pertusa* and *M. oculata* collected from the SW Rockall Trough
609 margin. These have been combined with previous results for the same region in order to characterize
610 mid-SPG dynamics and their potential impacts on NADW formation and climate in Northern Europe.

611 The coral ϵNd record exhibits pronounced variability, ranging from -12.2 ± 0.3 to -16.6 ± 0.4 ,
612 and tracks the position of the mid-SPG through time. As such, our new record reflects mixing of
613 waters from the SPG (-15) and from the sub-tropical Atlantic (-11) (mainly ENAW and MSW).

614 The time interval from 8.8 to 6.8 ka BP, which corresponds to the so-called Holocene thermal
615 maximum, has been linked to a greater eastward extension of the mid-SPG and a stronger AMOC.
616 This time interval was followed by a progressive contraction of the SPG between 7 and 5 ka BP. This
617 contraction was associated with a re-organization of thermohaline circulation and a progressive
618 decline in the production of the ISOW in the Nordic Seas during the mid-Holocene and the initiation
619 of LSW formation around 7 ka BP. After 5 ka BP, the CWC ϵNd value record displays millennial scale
620 variability with a greater eastward extension of the mid-SPG centered at 4.4, 2.5, 1.8 and 0.8 ka BP,
621 corresponding to the warm periods over northern Europe (such as the Medieval Climatic Anomaly
622 and the Roman Warm Period) and coinciding with an intensification of the AMOC. Radiocarbon
623 results show minor $\Delta^{14}\text{C}$ depletion (high reservoir ages) around 6.1, 4.4, 2.7, 2 and 0.7-1 ka BP. These

624 fluctuations, associated with a decrease of the ϵNd values, may result either from local variability of
625 the mixed layer depth or may be advected from the SPG where deep convection gives rise to
626 relatively $\Delta^{14}\text{C}$ depleted thermocline waters.

627 We suggest that the Early Holocene active state of the SPG intensity could be linked to an
628 entrainment of unradiogenic intermediate water masses originating from the SPG in the northern
629 deep-waters during its overflows from the Nordic Seas. This entrainment could have contributed to
630 the unradiogenic Nd isotopic signature observed for the NADW in the North Atlantic. Further
631 modeling experiments will be necessary to evaluate the impact of this process on the Nd isotopic
632 signature of the deep-water masses of the North Atlantic Ocean.

633 Overall, we propose that the ϵNd record for the SW Rockall Trough is a reliable record of SPG
634 extension in the NE Atlantic, independent of the intensity of the inter-gyre mixing. We have
635 identified a correspondence between mid-SPG extension, AMOC intensity and climate in Northern
636 Europe that needs to be further explored.

637

638

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651

652 **Figure captions:**

653

654 **Figure 1:** a) Bathymetric maps of the SW Rockall margin showing the locations of the studied cores
655 (Corals2012-39PC, Corals2013-18PC, M2000-25BX) and all other cores discussed in this study. b)
656 North Atlantic currents. The black arrows represent surface and intermediate currents. Dotted lines
657 represent deep circulation. The main water masses are labelled in italics. The gray arrows identify the

658 main directions of LSW propagation. AAIW, Antarctic Intermediate Water; DSOW, Denmark Strait
659 Overflow Water; ENACW, Eastern North Atlantic Central Water; ISOW, Iceland-Scotland Overflow
660 Water; LDW, Lower Deep Water; LSW, Labrador Sea Water; MSW, Mediterranean Sea Water; NADW,
661 North Atlantic Deep Water; SAIW, Subarctic Intermediate Water; SEC, Shelf Edge Current; WNACW,
662 Western North Atlantic Central Water. Map created with Ocean Data View Software (Schlitzer, R.,
663 Ocean Data View, <http://odv.awi.de>, 2015).

664

665 **Figure 2:** a) Potential temperature ($^{\circ}\text{C}$), salinity and ϵNd depth profiles for the water station ICE-CTD-
666 03 located in the Rockall Trough and collected in June and July 2012 during the ICE-CTD cruise on
667 board of the *R/V Thalassa* (Dubois-Dauphin et al., 2017). The water depth of the modern CWC
668 investigated in this study (M2000-25BX) is indicated by a gray horizontal band; b) Maps showing the
669 distribution of the salinity and the seawater ϵNd in the North Atlantic (Copard et al., 2011; Dubois-
670 Dauphin et al., 2017; Lambelet et al., 2016) as well as a schematic representation of the circulation at
671 750 m. The salinity distribution at 750 m of water depth has been also reported. The red star
672 indicates the position of the studied sites (locations of cores Corals2012-39PC and Corals2013-18PC).
673 Complete names of water mass accronyms have been provided in Figure 1

674

675 **Figure 3:** a) Comparison of deep-sea coral $\Delta^{14}\text{C}$ (‰) with the IntCal13 and Marine13 $\Delta^{14}\text{C}$ (‰)
676 calibration curves. The $\Delta^{14}\text{C}$ results obtained from CWC in this study are plotted with one item of
677 published data (Frank et al., 2004) against U-Th dates. The uncertainties of $\Delta^{14}\text{C}$ and of the U-Th ages
678 are plotted as 2σ error ellipses; b) Li/Mg temperature of CWC from cores Corals2012-39PC (black
679 circle) and Corals2013-18PC (open circle) (Bonneau et al., 2018); c) composite record of CWC ϵNd
680 obtained from studied cores (cores Corals2012-39PC and Corals2013-18PC) and from previous data
681 obtained from cores MD01-2454G and ENAM9915 (Colin et al., 2010; Copard et al., 2012). Positions
682 of cores have been reported in the Figure 1.

683

684 **Figure 4:** Comparison between the composite record of CWC ϵNd (f) and (a) $\delta^{18}\text{O}_{\text{seawater}}$ (‰ SMOW) at
685 Orphan Knoll in the Labrador Sea obtained from $\delta^{18}\text{O}$ and Mg/Ca derived $T^{\circ}\text{C}$ analysed on planktonic
686 foraminifera *G. bulloides* (cores HU91-045-093 and MD95-2024, 3488 m) (Hoogakker et al., 2015); (b)
687 Stacked data of SS values obtained from several cores located on the South Iceland Rise and Bjorn
688 Drift. This stacked data of sortable silt (SS) represents the relative strength of the Iceland–Scotland
689 overflow (ISOW) (Thornalley et al., 2013). Higher values for the SS mean size represent an increase in
690 the velocity of the ISOW (e.g. Bianchi and McCave, 1999; Thornalley et al., 2013) ; (c) Salinity of the
691 sub-thermocline water derived from the measurement of Mg/Ca– $\delta^{18}\text{O}$ on *G. inflata* in core RAPiD-12-

692 1K located south of Iceland; (d) Relative abundance of *G. inflata* (%) obtained for cores ENAM33
693 located east of the Faroe Islands near the Iceland Faroe gap (Staines-Urías et al., 2013); (e) SST
694 anomaly (°C) obtained on ODP site 658 located off Cap Blanc (Mauritania) (deMenocal et al., 2000).

695

696 **Tables caption:**

697 **Table 1:** Position, species and U-Th ages (Bonneau et al., 2018) of the deep-sea coral samples
698 investigated in this study. Results of Nd concentrations, Nd isotopic compositions, ¹⁴C age, Δ¹⁴C (‰)
699 and marine reservoir ages of the deep-sea coral samples investigated in this study are also
700 presented. $\epsilon\text{Nd} = \left[\left(\frac{{}^{143}\text{Nd}}{{}^{144}\text{Nd}} \right)_{\text{sample}} / \left(\frac{{}^{143}\text{Nd}}{{}^{144}\text{Nd}} \right)_{\text{CHUR}} - 1 \right] * 10000$, with the present-day
701 (¹⁴³Nd/¹⁴⁴Nd)_{CHUR} value of 0.512638 (Jacobsen and Wasserburg, 1980).

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Table 1

Depth (cm)	Species	$^{230}\text{Th}/^{234}\text{U}$ Age Cal (ka BP)	Nd (ppb)	$^{143}\text{Nd}/^{144}\text{Nd} \pm 2\sigma$	$\epsilon\text{Nd} \pm 2\sigma$	^{14}C age (yr BP) $\pm 1\sigma$	$\Delta^{14}\text{C}(\text{‰}) \pm 1\sigma$	MRE (yr $\pm 1\sigma$)
Coral2012-39PC								
(55.452N - 15.870W; 742 m water depth)								
0.5	<i>L. pertusa</i>	0.251 ± 0.014	41.3	0.511974 ± 0.000015	-13.0 ± 0.3			
1.5	<i>M. oculata</i>	0.276 ± 0.006	16.5	0.511995 ± 0.000028	-12.5 ± 0.5	750 ± 30	-58 ± 3	571 ± 25
4.5	<i>M. oculata</i>	0.473 ± 0.079		0.511998 ± 0.000010	-12.5 ± 0.2			
9.5	<i>L. pertusa</i>	0.663 ± 0.320		0.512014 ± 0.000014	-12.2 ± 0.3			
10.5	<i>M. oculata</i>	0.606 ± 0.036	22.0	0.511975 ± 0.000025	-12.9 ± 0.5	1120 ± 30	-64 ± 6	527 ± 25
15.5	<i>M. oculata</i>	0.740 ± 0.011	13.6	0.512007 ± 0.000024	-12.3 ± 0.5			
17.5	<i>L. pertusa</i>	0.719 ± 0.009	8.8	0.511987 ± 0.000018	-12.7 ± 0.4			
26.5	<i>L. pertusa</i>	0.854 ± 0.008	8.0	0.512005 ± 0.000024	-12.3 ± 0.5			
38.5	<i>L. pertusa</i>	1.020 ± 0.012	12.0	0.511963 ± 0.000019	-13.2 ± 0.4	1590 ± 30	-72 ± 4	455 ± 25
49.5	<i>L. pertusa</i>	1.475 ± 0.271		0.511983 ± 0.000010	-12.8 ± 0.2			
51.5	<i>L. pertusa</i>	1.447 ± 0.014	8.1	0.512000 ± 0.000024	-12.5 ± 0.5	1975 ± 30	-68 ± 4	387 ± 25
58.5	<i>L. pertusa</i>	1.431 ± 0.014	7.1	0.511982 ± 0.000018	-12.8 ± 0.4			
64.5	<i>M. oculata</i>	1.427 ± 0.017	13.4	0.511966 ± 0.000018	-13.1 ± 0.4			
73.5	<i>M. oculata</i>	1.854 ± 0.017	17.0	0.511913 ± 0.000018	-14.1 ± 0.4			
87.5	<i>L. pertusa</i>	2.008 ± 0.015	28.1	0.511965 ± 0.000014	-13.1 ± 0.3	2510 ± 30	-67 ± 5	430 ± 30
98.5	<i>M. oculata</i>	2.214 ± 0.010	7.2			2575 ± 30	-51 ± 4	334 ± 30
99.5	<i>M. oculata</i>	2.370 ± 0.029	13.5	0.511985 ± 0.000022	-12.7 ± 0.4	2660 ± 30	-43 ± 6	244 ± 30
105.5	<i>M. oculata</i>	2.308 ± 0.010	14.7	0.511939 ± 0.000019	-13.6 ± 0.4			
109.5	<i>M. oculata</i>	2.530 ± 0.015	16.8	0.511912 ± 0.000024	-14.2 ± 0.5	2910 ± 30	-55 ± 5	423 ± 30
120.5	<i>M. oculata</i>	2.786 ± 0.016	9.1	0.511956 ± 0.000019	-13.3 ± 0.4	3195 ± 30	-59 ± 5	485 ± 30
137.5	<i>L. pertusa</i>	3.673 ± 0.017	45.0	0.511959 ± 0.000011	-13.2 ± 0.2	3580 ± 30	-1.6 ± 5	179 ± 30
142.5	<i>M. oculata</i>	3.581 ± 0.015	11.0	0.511932 ± 0.000018	-13.8 ± 0.4	3745 ± 30	-33 ± 4	395 ± 25
149.5	<i>L. pertusa</i>	3.771 ± 0.069		0.511979 ± 0.000010	-12.8 ± 0.2			
151.5	<i>L. pertusa</i>	3.797 ± 0.014	24.5	0.511953 ± 0.000016	-13.4 ± 0.3	3865 ± 30	-21 ± 5	377 ± 30
169.5	<i>M. oculata</i>	4.436 ± 0.023	46.9	0.511906 ± 0.000011	-14.3 ± 0.2	4485 ± 30	-21 ± 6	495 ± 30
174.5	<i>L. pertusa</i>	4.527 ± 0.021	42.2	0.511927 ± 0.000014	-13.9 ± 0.3	4420 ± 30	-3 ± 6	369 ± 30
176.5	<i>M. oculata</i>	4.884 ± 0.020	11.1	0.511949 ± 0.000023	-13.4 ± 0.5			
198.5	<i>L. pertusa</i>	5.430 ± 0.039	27.3	0.511934 ± 0.000024	-13.7 ± 0.5	4990 ± 30	37 ± 8	361 ± 30

199.5	<i>M. oculata</i>	4.998	±0.228	0.511955	±0.000010	-13.3	±0.2				
209.5	<i>M. oculata</i>	5.258	±0.019								
217.5	<i>M. oculata</i>	6.023	±0.024	11.5	±0.000018	-14.5	±0.4	5575 ±30	35±6	262±30	
227.5	<i>L. pertusa</i>	6.110	±0.038	27.8	±0.000024	-13.8	±0.5				
235.5	<i>L. pertusa</i>	6.126	±0.022	15.6	±0.000018	-14.3	±0.4	5820 ±30	14±8	517±40	
241.5	<i>L. pertusa</i>	6.085	±0.017	8.8	±0.000018	-14.3	±0.4	5805 ±30	13±6	519±35	
251.5	<i>M. oculata</i>	6.670	±0.025	12.0	±0.000024	-14.6	±0.5	6240 ±30	31±7	377±40	
261.5	<i>M. oculata</i>	7.276	±0.033	12.9							
265.5	<i>M. oculata</i>	7.562	±0.037	16.3	±0.000021	-15.4	±0.4	6984 ±30	46±8	344±35	
282.5	<i>M. oculata</i>	8.680	±0.029	17.0	±0.000022	-15.0	±0.4	8185 ±35	31±8	282±40	
297.5	<i>M. oculata</i>	9.144	±0.031	21.3	±0.000026	-14.9	±0.5	8460 ±35	54±8	230±40	
307.5	<i>L. pertusa</i>	9.421	±0.057		±0.000024	-14.3	±0.5				
323.5	<i>M. oculata</i>	9.895	±0.032		±0.000024	-14.2	±0.5	9068 ±35	70±9	250±40	
332.5	<i>M. oculata</i>	10.079	±0.053	76.9	±0.000011	-15.0	±0.2	9325 ±35	60±12	457±40	
349.5	<i>L. pertusa</i>	10.024	±0.324								
354.5	<i>M. oculata</i>	10.105	±0.029	13.7	±0.000024	-15.0	±0.5	9270 ±35	70±9	419±45	
359.5	<i>M. oculata</i>	10.752	±0.047		±0.000011	-13.2	±0.2				
Corals2013-18PC											
(55.523N - 15.644W; 790											
m water depth)											
27.5	<i>L. pertusa</i>	1.293	±0.033		±0.000021	-13.0	±0.4				
92.5	<i>L. pertusa</i>	3.694	±0.017	23.0	±0.000013	-13.6	±0.2				
146.5	<i>L. pertusa</i>	5.227	±0.023	25.0	±0.000014	-14.1	±0.3				
178.0	<i>L. pertusa</i>	5.436	±0.056	22.4	±0.000018	-13.7	±0.4				
247.5	<i>L. pertusa</i>	6.069	±0.026	22.2	±0.000018	-13.9	±0.4				
288.5	<i>L. pertusa</i>	6.528	±0.078	45.3	±0.000014	-13.9	±0.3				
339.0	<i>L. pertusa</i>	7.175	±0.038	35.4	±0.000014	-15.4	±0.3				
359.0	<i>L. pertusa</i>	7.595	±0.036	55.0	±0.000014	-15.4	±0.3				
392.0	<i>L. pertusa</i>	8.650	±0.089	10.7	±0.000018	-16.6	±0.4				
M2000-											
25BX											
Top	<i>L. pertusa</i>	Living coral		9.7	±0.000018	-13.6	±0.4				







