



1 High variability of export fluxes along the North Atlantic 2 GEOTRACES section GA01: Particulate organic carbon export 3 deduced from the ^{234}Th method

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16 **Abstract.** In this study, we report Particulate Organic Carbon (POC) export fluxes estimated using the ^{234}Th -based
17 approach in different biogeochemical basins of the North Atlantic, as part of the GEOTRACES GA01 expedition
18 (GEOVIDE, May-June 2014). Surface POC export fluxes were deduced by combining export fluxes of ^{234}Th with the
19 POC to ^{234}Th ratio of sinking particles at the depth of export. Particles were collected in two size classes ($>53\ \mu\text{m}$ and
20 $1\text{-}53\ \mu\text{m}$) using *in-situ* pumps and the large size fraction was considered as representative of sinking material. Surface
21 POC export fluxes revealed latitudinal variations between provinces ranging from $1.4\ \text{mmol C m}^{-2}\ \text{d}^{-1}$ in the Irminger
22 basin where the bloom was close to its maximum peak, to $12\ \text{mmol C m}^{-2}\ \text{d}^{-1}$ near the Iberian Margin where the bloom
23 had already declined. In addition to the bloom staging, the variations of POC export fluxes were also related to the
24 phytoplankton community structure. In line with previous studies, the presence of coccolithophorids and diatoms
25 appeared to increase the POC export flux while stations dominated by pico-phytoplankton cells, such as cyanobacteria,
26 were characterized by lower fluxes. The surface POC export fluxes were then compared to *in-situ* and satellite primary
27 production (PP) in order to assess the export efficiency. This ratio strongly varied regionally and was generally low (\leq
28 14 %), except at two stations located near the Iberian margin (35%) and within the Labrador basin (38%), which were
29 characterized by unusual low *in-situ* PP. We thus conclude that the North Atlantic during this period was not as efficient
30 in exporting carbon from the surface, as described in recent studies. Finally, we estimated the flux of POC exported
31 100 m below the surface export depth in order to investigate the transfer efficiency along the section. This parameter



32 was also highly regional-dependent but the lowest attenuation of the POC flux was observed at stations where
33 coccolithophorids dominated.

34 1. Introduction

35 Through the sinking of particulate biogenic material, the biological carbon pump (BCP) plays a major role on the
36 sequestration of carbon-rich particles in the ocean interior. The North Atlantic harbors one of the most productive
37 spring phytoplankton bloom of the world's ocean (Esaias et al., 1986; Longhurst, 2010), generating an important pulse
38 of biogenic sinking particles (Buesseler et al., 1992; Honjo and Manganini, 1993; Le Moigne et al., 2013), which
39 accounts up to 18 % of the global BCP (Sanders et al., 2014). As a consequence, the North Atlantic has been identified
40 as an efficient ocean to export carbon to depth (Buesseler and Boyd, 2009; Herndl and Reinthaler, 2013).

41 The North Atlantic consists of several provinces (sensu Longhurst, 1995) characterized by distinct biogeochemical
42 and physical characteristics. For example, the low nutrient availabilities in the North Atlantic subtropical gyre limits
43 the biomass development (Moore et al., 2008), dominated by pico-phytoplankton such as cyanobacteria (Zehr and
44 Ward, 2002). Northward, in the Irminger and Labrador basins, the phytoplankton growth is strongly seasonally light-
45 limited (Riley, 1957) and the key parameter for alleviating these limitations is the progressive shoaling of the mixed
46 layer. There, diatoms dominate the phytoplankton bloom until the exhaustion of the silicic acid stock (Martin et al.,
47 2011). Then, an intense bloom of coccolithophorids develops (Poulton et al., 2010; Raitso et al., 2006). There, diatoms
48 dominate the phytoplankton bloom until the exhaustion of the silicic acid stock (Martin et al., 2011). Then, an intense
49 bloom of coccolithophorids develops (Poulton et al., 2010; Raitso et al., 2006). Between these basins, the west
50 European and Icelandic basins are like a transition zone where nutrients and/or light can limit the primary production
51 (Henson et al., 2009). In summer, at the end of the bloom, these basins can be iron-limited, and become “High Nutrient
52 Low Chlorophyll” regions (Blain et al., 2004; Moore et al., 2006; Sanders et al., 2005). The North Atlantic is thus a
53 heterogeneous basin in terms of limitations and phytoplankton communities. Ecosystem structure is thought to play an
54 important role on the fraction of particulate organic carbon (POC) which is exported from the surface ocean. Indeed,
55 Guidi et al. (2009) suggested that phytoplankton composition explained 68% of the variance in POC flux at 400 m.
56 High export efficiencies are reported in productive regions where diatoms dominate, but the exported material is
57 relatively labile and prone to remineralisation leading to low deep export flux and transfer efficiencies. Conversely, in
58 oligotrophic regions, where diatoms are largely absent, primary production is low and mostly regenerated.

59 Consequently, export efficiencies are low but the eventual exported material is likely to be refractory resulting in high
60 transfer efficiencies (Henson et al., 2012; Lam et al., 2011; Lima et al., 2014; Marsay et al., 2015). Phytoplankton size
61 structure has also been shown to be an important factor in controlling the POC export fluxes. Guidi et al. (2015)
62 highlighted that the exported POC was more refractory and the remineralisation depth was greater when the fraction
63 of micro-phytoplankton decreased or the fraction of pico-phytoplankton increased.

64 According to the impact of these biogeochemical factors on the POC export, the efficiency of the North Atlantic to
65 transfer POC to the deep ocean can be questioned. In this context, we investigated the derived-POC export fluxes using
66 the Thorium-234 (^{234}Th) approach in the different basins of the North Atlantic. ^{234}Th is a highly particle reactive



67 element, with a short half-life (24.1 days), which is widely used to explore particle export over short time events such
68 as phytoplankton blooms. A deficit of ^{234}Th with respect to its radioactive parent ^{238}U (conservative in seawater) is
69 usually observed in the upper water column where particles sink. In subsurface, excess of ^{234}Th compared to ^{238}U can
70 be observed pointing to a remineralisation of ^{234}Th -bearing particles (Savoie et al., 2004). A ^{234}Th flux can be
71 quantified and then converted into a POC flux by using the POC: ^{234}Th ratio of sinking particles at the depth of export
72 (Buesseler et al., 2006).

73 In this paper, we first explore the basin-wide variations of the ^{234}Th fluxes in order to define different export and
74 attenuation regimes in the North Atlantic. The POC export fluxes are then discussed with regards to the stage and
75 intensity of the bloom and the phytoplankton structure. Finally, using *in-situ* primary production, satellite primary
76 production and deeper POC export fluxes, we investigate the export and transfer efficiencies in the North Atlantic.

77 2. Methods

78 For the purpose of this work, 11 stations were investigated along the GEOTRACES section GA01 (GEOVIDE cruise,
79 15 May to 30 June 2014; R/V Pourquoi Pas?). The studied area crossed different basins: the Iberian basin, the west
80 European basin, the Icelandic basin, the Irminger basin and the Labrador basin (Fig.1).

81 2.1. Description of the regional basins

82 The Iberian basin (Stations 1 and 13) was characterized by oligotrophic conditions, with NO_3^- and $\text{Si}(\text{OH})_4$
83 concentrations under $1 \mu\text{mol L}^{-1}$ in the upper 40 m, despite the proximity of the Iberian margin, where sits a natural
84 upwelling (Costa Goela et al., 2016; Zúñiga et al., 2016; <http://marine.copernicus.eu/>), that potentially fuels the area
85 with nutrient-rich deep waters. Dissolved iron (dFe) concentrations were non-limiting, with concentrations in surface
86 waters varying between 0.22 and 1.0 nmol L^{-1} (Tonnard et al., 2018; this issue). In this basin, the total chlorophyll-*a*
87 (Chl-*a*) in the upper 200 m was the lowest averaging 0.26 mg m^{-3} and nano-phytoplanktonic species dominated but
88 with a mixed proportion of micro-, nano- and pico-phytoplanktonic species. The highest proportion of pico-
89 phytoplanktonic species was observed at Station 13 (34% of the total Chl-*a*; Tonnard et al., in prep.) with cyanobacteria
90 contributing for 12% of the total Chl-*a*.

91 The west European basin (Stations 21 and 26) was influenced by the North Atlantic Current (NAC): the southernmost
92 sub-branch evolving in a cyclonic eddy at Station 21 and the sub-arctic front (SAF) at Station 26. This front separates
93 cold and fresh waters from the subpolar region and the warm and salty waters from the subtropical region (Zunino et
94 al., 2017; this issue). Both stations were characterized by low surface $\text{Si}(\text{OH})_4$ concentrations ($\leq 1 \mu\text{mol L}^{-1}$), moderate
95 NO_3^- ($\geq 1 \mu\text{mol L}^{-1}$) and dFe concentrations (0.17 and 0.18 nmol L^{-1} in the upper 20 m). The total Chl-*a* concentrations
96 were higher in this basin reaching 0.45 mg m^{-3} ($n=2$). Micro-phytoplanktonic species dominated, contributing for more
97 than 45% of total Chl-*a*. The combined contribution of diatoms, dinoflagellates and haptophytes (including
98 coccolithophorids) was about 71%, with a higher contribution of diatoms (41%) compared to the two other taxa (4 and
99 26%, respectively; Tonnard et al., in prep.).



100 Within the Icelandic basin, Stations 32 and 38 were respectively influenced by the NAC northern branch and the
101 Eastern Reykjanes Ridge Current (ERRC; Zunino et al., 2017; this issue). The surface waters had low Si(OH)_4
102 concentrations ($1 \mu\text{mol L}^{-1}$), relatively high dFe concentrations ($> 0.45 \text{ nmol L}^{-1}$; Tonnard et al., 2018; this issue) and
103 high NO_3^- concentrations ($> 6 \mu\text{mol L}^{-1}$). Concentrations of total Chl-*a* reached 0.62 mg m^{-3} on average in the upper
104 200 m of Station 32 and 0.44 mg m^{-3} at Station 38, a value close to the one observed in the west European basin. Nano-
105 phytoplanktonic species dominated, up to 81% of the total Chl-*a* at Station 38, and this was associated to the highest
106 proportion of the haptophytes (56 and 55% at Stations 32 and 38; Tonnard et al., in prep.).

107 In the Irminger basin, Station 44 was located in the Irminger gyre while Station 51, located close to Greenland, was
108 influenced by the Eastern Greenland Current (EGC) guided by the continental slope (Zunino et al., 2017; this issue).
109 This basin was characterized by high surface Si(OH)_4 and NO_3^- concentrations ($> 6 \mu\text{mol L}^{-1}$) and moderate to high
110 dFe concentrations ($0.24\text{-}1.3 \text{ nmol L}^{-1}$; Tonnard et al., 2018; this issue). The highest total Chl-*a* concentrations were
111 measured in this basin, averaging 0.98 mg m^{-3} ($n=2$). Micro-phytoplanktonic species, and more specifically diatoms,
112 clearly dominated with a contribution of about 55 and 77% of the total Chl-*a* at Stations 44 and 51, respectively
113 (Tonnard et al., in prep.).

114 The Labrador basin (Stations 64, 69 and 77) was characterized by the subduction of the Labrador Seawater (LSW)
115 which was particularly intense (1700 m-deep convection) during the winter 2013-2014 (Kieke and Yashayaev, 2015).
116 Stations 64 and 77 were also influenced by the Western Greenland Current (WGC) and the Labrador Current (LC),
117 respectively. Macronutrients in the surface waters of the Labrador basin were characterized by a north to south gradient,
118 with high NO_3^- and Si(OH)_4 concentrations at Station 64 ($\geq 4.5 \mu\text{mol L}^{-1}$), decreasing gradually to the south with low
119 concentrations at Station 77 ($\sim 1 \mu\text{mol L}^{-1}$ and $< 1 \mu\text{mol L}^{-1}$, respectively). Moderate dFe concentrations were observed
120 in this area (between 0.23 and 0.30 nmol L^{-1} in the upper 20 m; Tonnard et al., 2018; this issue). As the Irminger basin,
121 micro-phytoplankton species (diatoms) dominated with a contribution averaging 57% of the total Chl-*a* (Tonnard et
122 al., in prep.). However, the total Chl-*a* concentrations were low (down to 0.25 mg m^{-3}), especially at Stations 64 and
123 69.

124 2.2. Total ^{234}Th and ^{238}U

125 Total ^{234}Th activities were determined from 4 L unfiltered seawater samples collected with Niskin bottles and stored
126 in polypropylene Nalgene bottles. Usually, 17 or 18 depths were sampled between the surface and 1000-1500 m, except
127 at Stations 26 and 77 where only 9 and 15 depths were sampled, respectively (Table S1). In addition, deep samples
128 ($n=15$; between 1000 and 3500 m) were taken for the calibration of low level beta counting (van der Loeff et al., 2006).
129 Deep water samples are generally considered to be in secular equilibrium regarding the ^{234}Th - ^{238}U pair. Seawater
130 samples were processed following the method developed by Pike et al. (2005). Briefly, samples were acidified at pH
131 2 using concentrated HNO_3 (suprapur grade, Merck), spiked with 1 mL of ^{230}Th yield monitor to estimate the ^{234}Th
132 recovery after the sample processing. After 12 hours of equilibration, pH was increased to 8.5 using concentrated
133 NH_4OH (suprapur grade, Merck). One hundred micro-liters of KMnO_4 and MnCl_2 (analytical grade, Merck) were
134 added to form a manganese oxide precipitate and, after 12 hours of equilibration, samples were filtered on quartz-
135 microfibre discs (QMA, Sartorius, $1 \mu\text{m}$ nominal porosity, 25 mm diameter). On board, filters were dried at $50 \text{ }^\circ\text{C}$



136 overnight, mounted on nylon holders, covered with Mylar and aluminum foil and ^{234}Th activity counted using low
137 level beta counters (RISØ, Denmark). Beta activity counting was continued until a relative standard deviation (RSD)
138 $\leq 2\%$ was reached. At the home-laboratory, residual beta activity was measured for each sample after a delay of six
139 ^{234}Th half-lives (~ 6 months) and these residual counts were subtracted from the gross counts obtained on-board. All
140 samples were then processed for Th recovery using ^{229}Th as a second yield tracer. Briefly, filters were dismantled
141 from the nylon holders and transferred to clean 30 mL teflon vials (Savillex). All samples were spiked with 50 μL of
142 ^{229}Th , dissolved in a 10 mL mix of 8M HNO_3 /1M H_2O_2 (suprapur grade, Merck), heated overnight at 60 °C and filtered
143 through Acrodisc® syringe filters (Pall, Nylon membrane, nominal porosity=0.2 μm , diameter=25 mm). Two
144 milliliters of the filtrate were pre-concentrated by evaporation and the residue diluted in 6 mL of 1.4 M HNO_3 (suprapur
145 grade, Merck). ^{230}Th and ^{229}Th concentrations were measured by SF-ICP-MS (Element 2, Thermo Scientific) in low
146 resolution mode. Each sample was analyzed 3 times and the precision of the ^{230}Th : ^{229}Th ratios averaged 1.2% (RSD),
147 which is within the range indicated by Pike et al. (2005). The total ^{234}Th recovery, involving all the steps described
148 above, was $91 \pm 14\%$ ($n=200$). Uncertainty on total ^{234}Th activity was estimated using error propagation and accounts
149 between 0.04 and 0.10 dpm L^{-1} .
150 The ^{238}U activity was deduced from salinity using the Equation 1, given by Owens et al. (2011):

$$151 \quad {}^{238}\text{U} = 0.0786 \times S - 0.315 \quad (1)$$

152 where ^{238}U is the ^{238}U activity in dpm L^{-1} and S is the salinity in psu.

153 **2.3. Particulate ^{234}Th and POC sampling and analysis**

154 Suspended particles were collected using *in-situ* large-volume filtration (100-1600 L) systems (Challenger Oceanics
155 and McLane pumps; ISP hereafter for “*in-situ* pumps”) through paired 142 mm-diameter filters: a 53 μm mesh nylon
156 screen (SEFAR-PETEX®; polyester) and a 1 μm pore size quartz-microfiber filter (QMA, Sartorius), respectively.
157 Two size fractions of particles were thus collected: the small size fraction (referred to as SSF hereafter, 1-53 μm) and
158 the large size fraction (referred to as LSF hereafter, $>53\ \mu\text{m}$). Filters were cleaned prior to the cruise as follows: PETEX
159 screens were soaked in 0.6M HCl, (Normapur, Merck) rinsed with Milli-Q water, dried at ambient temperature in a
160 laminar flow hood and stored in clean plastic bags; QMA filters were pre-combusted at 450 °C during 4 h and stored
161 in aluminum foils until use. *In-situ* pumps were deployed on a stainless steel cable between 15 and 800 m and the
162 pumping time was approximatively 2-3 hours (Table S2).

163 After collection, filters were processed on board. The 142 mm PETEX screen was cut into quarters using a clean
164 scalpel and two quarters were processed in this study. Large particles collected on the screen were rinsed off using
165 0.45 μm filtered seawater and re-filtered under a laminar flow hood on a silver filter (SterliTech, porosity=0.45 μm ,
166 diameter=25 mm) for the first quarter of the PETEX screen and on a GF/F filter (Whatman®, porosity=0.7 μm ,
167 diameter=25 mm) for the second quarter.

168 The QMA filters were sub-sampled with a perspex punch of 25 mm diameter.

169 Then, silver, GF/F and QMA filters were dried overnight (50 °C) and prepared for beta counting (see Section 2.2).

170 After counting the residual beta activity (~ 6 months later), samples were prepared for POC, PN analyses along with



171 their $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ isotopic compositions (here we present only POC data). In brief, filters were dismantled from filter
 172 holders and fumed with HCl vapor overnight inside a glass desiccator to remove the carbonate phase. After a drying
 173 step at 50 °C, samples were packed in precombusted (450 °C overnight) silver cups and analyzed with an elemental
 174 analyzer – isotope ratio mass spectrometer (EA-IRMS, Delta V Plus, Thermo Scientific). Acetanilide standards were
 175 used for the calibration. The detection limits and C blanks were respectively 0.63 and 0.80 μmol for Ag filters (n=11)
 176 and were 0.49 and 1.52 μmol for QMA filters (n=13).

177 2.4. Export fluxes of ^{234}Th

178 Thorium-234 activity in surface waters can be described using a simple mass balance equation (Savoye et al., 2006),
 179 which accounts for production from ^{238}U decay, ^{234}Th decay, sinking flux and transport as follow:

$$180 \quad \frac{dA_{Th}}{dt} = \lambda A_U - \lambda A_{Th} - P + V \quad (2)$$

181 where A_{Th} is the activity of total ^{234}Th in dpm L^{-1} ; A_U is the salinity-derived activity of ^{238}U in dpm L^{-1} , λ is the ^{234}Th
 182 decay constant (0.0288 d^{-1}); P is the net removal of ^{234}Th on sinking particles in $\text{dpm L}^{-1} \text{d}^{-1}$; V is the sum of the
 183 advective and diffusive fluxes in $\text{dpm L}^{-1} \text{d}^{-1}$.

184 Assuming steady state (constant total ^{234}Th activity with time) and neglecting the physical term V (Buesseler et al.,
 185 1992), the net export flux of particulate ^{234}Th can be determined using the following equation:

$$186 \quad P_z = \lambda \int_0^z (A_U - A_{Th}) dz \quad (3)$$

187 where P_z is the integrated flux of ^{234}Th from the surface to the depth z in $\text{dpm m}^{-2} \text{d}^{-1}$. Equation 3 has been solved for
 188 z representing the depth Eq at the base of the deficit zone (where ^{234}Th activity is back to secular equilibrium with
 189 ^{238}U) as well as for z representing the base of the Primary Production Zone (PPZ), i.e. the depth where *in-situ*
 190 fluorescence was only 10% of the surface value (see Section 3.1). The validity of the assumptions used for solving
 191 Equation 3 is discussed in Sections 4.1 and 4.2.

192 In Section 4.2, we attempt to calculate the ^{234}Th fluxes at the Eq depth by using a non-steady state (NSS) model (Savoye
 193 et al., 2006) that can be described as follows:

$$194 \quad P_z = \lambda \left[\frac{A_U (1 - e^{-\lambda \Delta t}) + A_{Th1} e^{-\lambda \Delta t} - A_{Th2}}{1 - e^{-\lambda \Delta t}} \right] \quad (4)$$

195 where Δt is the time interval between two visits of a single station; A_{Th1} and A_{Th2} are the ^{234}Th activities at the first and
 196 second visits, respectively.

197 In order to account for possible ^{234}Th excess relative to ^{238}U below Eq, we evaluated the ^{234}Th flux at the depth
 198 corresponding to Eq + 100 m, and compared this flux with the ^{234}Th flux at Eq (Black et al., 2017). The difference
 199 R100 is expressed as follows:

$$200 \quad R100 = P_{Eq} - P_{Eq+100} \quad (5)$$



201 where $R100$ is the flux reduction in $\text{dpm m}^{-2} \text{d}^{-1}$.

202 2.5. Scavenging fluxes of ^{234}Th

203 Using the data for particulate ^{234}Th activities next to total ^{234}Th activities, it is possible to describe the ^{234}Th activity
 204 with a 2-box model (Coale and Bruland, 1985). This model accounts for the partitioning of ^{234}Th between the
 205 particulate and dissolved phase. The mass balance equation for dissolved ^{234}Th can be written as follows:

$$206 \quad \frac{dA_{Thd}}{dt} = \lambda A_U - \lambda A_{Thd} - J + V_d \quad (6)$$

207 where A_{Thd} is the activity of dissolved ^{234}Th in dpm L^{-1} ; and J is the net removal flux from the dissolved to the particulate
 208 form (scavenging flux) in $\text{dpm L}^{-1} \text{d}^{-1}$. Here, ^{238}U is considered as a dissolved specie which produces ^{234}Th in the
 209 dissolved phase. The second mass balance equation describes the particulate ^{234}Th pool as follows:

$$210 \quad \frac{dA_{Thp}}{dt} = J - \lambda A_{Thp} - P + V_p \quad (7)$$

211 where A_{Thp} is the activity of particulate ^{234}Th (in dpm L^{-1}); the scavenging flux J described above becomes here the
 212 source term; P is the net removal flux of particulate ^{234}Th with sinking particles and already described with the one
 213 box model (Eq. 2 and 3). Using again the steady state assumption (constant activities for both particulate and dissolved
 214 ^{234}Th) and ignoring the physical terms (V_d and V_p), Equation 5 becomes:

$$215 \quad J_z = \lambda \int_0^z (A_U - A_{Thd}) dz \quad (8)$$

216 where J_z in $\text{dpm m}^{-2} \text{d}^{-1}$ is the net integrated flux of dissolved ^{234}Th to the depth z . In our case, the calculation was
 217 performed at the Eq depth for comparison with the ^{234}Th export flux.

218 In a similar way, Equation 7 is simplified to:

$$219 \quad J_z = \lambda \int_0^z A_{Thp} dz + P_z \quad (9)$$

220 Under these conditions, the net flux of scavenging J (source term) is defined by two output terms, the export of
 221 particulate ^{234}Th (P_z) due to sinking particles and the sorption of dissolved ^{234}Th onto non-sinking suspended particles.

222 2.6. *In-situ* primary production

223 In order to determine the *in-situ* daily PP, stable isotope incubation techniques ($\text{H}^{13}\text{CO}_3^-$) were conducted using
 224 seawater collected in the euphotic zone based on Photosynthetically Active Radiation (PAR) profiles as described
 225 elsewhere (Fonseca-Batista et al., 2018). Briefly, at each station, seawater was sampled from 3 to 6 depths (from 54 to
 226 0.2% of surface PAR) and incubated on deck with a $\text{H}^{13}\text{CO}_3^-$ enriched substrate. All on-board incubations were
 227 sampled at the initial state and after 24h of experiment, seawater was then filtered through microglass fiber filters
 228 (MGF, 0.7 μm porosity, Sartorius). At the home-laboratory, POC concentrations and isotopic composition were
 229 analyzed by EA-IRMS and uptake rates were deduced following the Hama et al. (1983) method. Daily PP was then



230 estimated for each station by integrating the uptake rates as a function of depth from the surface down to 0.2% of
231 surface PAR. Note that the sampling to determine the *in-situ* PP at Station 51 occurred 24h after the sampling of the
232 total ^{234}Th , particulate ^{234}Th and POC.

233 2.7. Satellite primary production

234 In addition to the daily *in-situ* PP described above, PP was obtained from satellite data products available from the
235 Ocean Productivity website at Oregon State University (<http://www.science.oregonstate.edu/ocean.productivity/>) with
236 a 9 km spatial resolution and 8-day temporal resolution obtained from MODIS and SeaWiFS satellites. Three different
237 models can be used to obtain satellite-derived PP: the standard Vertically Generalized Production Model (VGPM;
238 Behrenfeld and Falkowski, 1997), the Eppley-VGPM (Eppley, 1972) and the Carbon-Based Production Model
239 (CbPM; (Behrenfeld et al., 2005; Westberry et al., 2008). In this study, we present the PP data derived with the VGPM
240 model, since its output fitted best the *in-situ* PP. Furthermore, Puigcorbé et al. (2017) have shown that the PP data from
241 the CbPM deviated from *in-situ*, especially in subpolar areas, probably due to the presence of coccolithophorids and
242 large diatoms that increase disproportionately the backscattering due to their shells and frustules.
243 PP data were averaged over a 5×5 pixel box centered on the different sampling sites, corresponding to a surface area
244 of 2025 km² (45 km \times 45 km). The PP was averaged for the week (8 days), the month (32 days) and the whole
245 productive period prior to the sampling date.

246 3. Results

247 3.1. Depth distribution and spatial variability of the $^{234}\text{Th}/^{238}\text{U}$ disequilibria

248 The complete dataset of total ^{234}Th , ^{238}U activities and the corresponding $^{234}\text{Th}/^{238}\text{U}$ ratios are presented in Table S1
249 and Figure 2 shows the depth profiles of total ^{234}Th and ^{238}U activities. A deficit of ^{234}Th relative to ^{238}U ratio indicates
250 a loss of ^{234}Th due to the export by particles (Buesseler et al., 1992; Cochran and Masqué, 2003). Conversely, excess
251 of ^{234}Th relative to ^{238}U implies ^{234}Th accumulation, which can be related to particle degradation, releasing ^{234}Th in the
252 dissolved phase (Waples et al., 2006). Along the transect, total ^{234}Th activities ranged between 1.23 and 2.90 dpm L⁻¹,
253 while ^{238}U activities ranged from 2.19 to 2.53 dpm L⁻¹.

254 At all stations, deficits of ^{234}Th relative to ^{238}U were observed in the upper 100 m. The lowest $^{234}\text{Th}/^{238}\text{U}$ ratios were
255 measured in the upper 40 m, ranging from 0.50 (Station 38) to 0.90 (Station 44). Generally, the lowest ratios were
256 observed in the west European and Icelandic basins (median 0.61 ± 0.12 , $n=4$), while in the other basins, the median
257 surface $^{234}\text{Th}/^{238}\text{U}$ ratios was 0.74 ± 0.06 ($n=8$). Moreover, at Stations 21, 26 and 32 located within the west European
258 and Icelandic basins, ratios below 0.8 deepened further than the other stations (91 ± 14 m, $n=3$, compared to 33 ± 16
259 m, $n=8$). Total ^{234}Th activities increased progressively with depth and were back to equilibrium with ^{238}U at different
260 depths between stations: Eq reached 100 ± 10 m ($n=2$) in the Iberian basin, increased to 128 ± 51 m ($n=4$) in the west
261 European and Icelandic basins and finally decreased to 68 ± 27 m in the Irminger and Labrador basins ($n=5$).

262 This Eq depth matched relatively well with the base of the PPZ depth, as only 16 m of difference was observed between
263 both depths in average along the transect and with the biggest difference (~ 60 m) at Stations 1, 32 and 51 (Fig. 2). The



264 PPZ depth, defined as the depth at which the fluorescence reaches 10% of its maximum value (Marra et al., 2014), was
265 used in different studies as the integration depth for ^{234}Th deficits (Owens et al., 2014; Roca-Martí et al., 2016b) but
266 the good correspondence between Eq and PPZ confirms that the Eq depth is appropriate to calculate export fluxes.
267 Below Eq, significant excesses of ^{234}Th relative to ^{238}U (i.e., $^{234}\text{Th}/^{238}\text{U}$ ratio >1.1) were only observed at Stations 1
268 (800 m), 13 (140 m), 21 (300 m) and 77 (400 and 700 m). Occasionally a small but significant ^{234}Th deficit was also
269 observed at depths deeper than Eq. This is the case for Station 44 at 140 m and 800 m and at Station 51 between 400
270 and 700 m (Fig. 2).

271 3.2. Export and scavenging fluxes of ^{234}Th

272 Steady state (SS) ^{234}Th export fluxes, integrated at the Eq and PPZ depths ranged respectively from 321 to 2282 dpm
273 $\text{m}^{-2} \text{d}^{-1}$; and from 321 to 1723 dpm $\text{m}^{-2} \text{d}^{-1}$ (Table 1). Similar fluxes were found at both integration depths with
274 differences smaller than 12%, except at Stations 32 and 51 where fluxes at Eq were 36 and 46% greater than those at
275 the base of the PPZ. Considering that there can be export (or remineralisation) below or above the PPZ depth, in the
276 following, only the export fluxes at the Eq depth are discussed as they represent the fully-integrated depletion of ^{234}Th
277 in the upper waters and thus the maximal export.

278 The highest ^{234}Th export fluxes at Eq using the SS model were observed in the west European and Icelandic basins,
279 reaching $2282 \pm 119 \text{ dpm m}^{-2} \text{d}^{-1}$ at Station 32, while the lowest flux was determined in the Irminger basin (321 ± 66
280 $\text{dpm m}^{-2} \text{d}^{-1}$ at Station 44; Table 1).

281 Using the SS model, we also determined the ^{234}Th scavenging fluxes at the Eq depth (Equations 8 and 9), along the
282 transect. These fluxes ranged from 1495 to 3917 dpm $\text{m}^{-2} \text{d}^{-1}$ at Stations 38 and 21, respectively (Table 1). In general,
283 the highest ^{234}Th scavenging fluxes were observed in the west European basin and at Stations 13 and 32 in the Iberian
284 and Icelandic basins, respectively (median value: $3294 \pm 548 \text{ dpm m}^{-2} \text{d}^{-1}$, $n=4$). The lowest fluxes were determined in
285 the Labrador basin and at Stations 1 and 38 in the Iberian and Icelandic basins respectively ($1495 \pm 176 \text{ dpm m}^{-2} \text{d}^{-1}$,
286 $n=5$).

287 3.3. Particulate ^{234}Th and POC distributions

288 Particulate ^{234}Th activities and POC concentrations for the small size fraction (SSF; 1-53 μm) and the large size fraction
289 (LSF; $>53 \mu\text{m}$) are presented in Table S2.

290 LSF particles were collected on silver and GF/F filters (see Section 2.2), and POC concentrations and ^{234}Th activities
291 were determined on both filter types. The POC concentrations and ^{234}Th activities compared well between both filter
292 types, indicating a relatively good repeatability in sampling ($^{234}\text{Th}_{\text{GFF}} = 0.63 \times ^{234}\text{Th}_{\text{silver}} + 0.01$ with $r^2=0.88$, p-
293 value <0.01 and $n=58$; and $\text{POC}_{\text{GFF}} = 0.86 \text{ POC}_{\text{silver}} + 0.08$ with $r^2=0.90$, p-value <0.01 and $n=58$; Fig. S1). Yet,
294 concentrations from GF/F filters are systematically lower than the ones from Ag filters, most likely because of the
295 different pore size filter (0.7 μm for GF/F filter vs 0.45 μm for Ag filter).

296 High POC concentrations and ^{234}Th activities were observed in the upper 100 m, ranging respectively from 0.42 to 17
297 $\mu\text{mol C L}^{-1}$ and from 0.02 to 1.2 dpm $^{234}\text{Th L}^{-1}$ at Stations 64 and 44 for the SSF and from 0.16 to 4.0 $\mu\text{mol C L}^{-1}$ and
298 from 0.01 to 0.61 dpm $^{234}\text{Th L}^{-1}$ at Stations 38 and 44 for the LSF. At all stations, both POC concentrations and ^{234}Th



299 activities decreased rapidly in the subsurface to remain essentially constant below 200 m. In general, the lowest POC
300 concentrations and particulate ^{234}Th activities were determined in the Iberian basin. Moderate concentrations and
301 activities were measured in the west European, Icelandic and Labrador basins, except at Station 77 where POC
302 concentrations and ^{234}Th activities were higher in surface reaching $11 \mu\text{mol C L}^{-1}$ and $0.45 \text{ dpm } ^{234}\text{Th L}^{-1}$ for the SSF,
303 and, $3.0 \mu\text{mol C L}^{-1}$ and $0.20 \text{ dpm } ^{234}\text{Th L}^{-1}$ for the LSF. The highest concentrations and activities were measured in
304 the Irminger basin.

305 Along the transect, POC contents and ^{234}Th activities were predominantly carried by the SSF, with POC and ^{234}Th in
306 the LSF accounting only for 13 and 12% (median values; $n=56$) of total POC and ^{234}Th , respectively. This feature was
307 accentuated in the Iberian basin, especially at Station 13, where ^{234}Th in the LSF averaged $6.6 \pm 1.3\%$ (median ± 1
308 s.d.; $n=5$) of the total particle associated activity and in the Icelandic basin where the LSF represented only 9.2 ± 8.1
309 % ($n=10$) of the total POC. The highest proportion of POC and ^{234}Th in the LSF were observed in surface waters of
310 the west European basin (reaching respectively 51 and 47%), the Irminger basin (reaching respectively 39 and 56%)
311 and the Labrador basin (reaching respectively 42 and 51%).

312 Large variations were also observed along the transect when comparing the fractions of the whole particulate ^{234}Th
313 (sum of the LSF and SSF), accounting from 9% (Station 1) to 94% (Station 44) of the total ^{234}Th activity. The median
314 value was 26% ($n=11$) but four stations were characterized by different partitioning compared to the general trend.
315 Stations 1, 64 and 69 were characterized by a low particulate ^{234}Th activity accounting for 9, 10 and 15% of the total
316 ^{234}Th activity in agreement with the low POC concentrations observed at these stations. Conversely, Station 44 was
317 characterized by the highest fraction of particulate ^{234}Th (94%), reflecting an important particle concentration in surface
318 waters.

319 **3.4. POC: ^{234}Th ratios in particles**

320 Profiles of POC: ^{234}Th ratios for the SSF and LSF are shown in Figure 3. POC: ^{234}Th ratios spanned two orders of
321 magnitude, ranging between 0.51 (Station 1, 800 m) to 53.7 (Station 32, 30 m) $\mu\text{mol dpm}^{-1}$ in the SSF and from 1.05
322 (Station 21, 400 m) to 30.6 (Station 1, 30 m) $\mu\text{mol dpm}^{-1}$ in the LSF. The highest and most variable ratios were
323 determined in the upper water column (~ 30 m) with values ranging from $4.73 \mu\text{mol dpm}^{-1}$ at Station 13 to $53.7 \mu\text{mol}$
324 dpm^{-1} at Station 32 for the SSF, and from $5.6 \mu\text{mol dpm}^{-1}$ at Station 38 to $30.6 \mu\text{mol dpm}^{-1}$ at Station 1 for the LSF.
325 Then, the ratios decreased with depth, in both size fractions, down to 100 m. Below 100 m, the ratios remained
326 relatively constant in both size fractions with median values of $1.8 \pm 1.1 \mu\text{mol dpm}^{-1}$ in the Iberian basin ($n=13$; Stations
327 1 and 13), $3.0 \pm 1.3 \mu\text{mol dpm}^{-1}$ in the west European and Icelandic basins ($n=24$; Stations 21, 26, 32 and 38), $3.7 \pm$
328 $0.9 \mu\text{mol dpm}^{-1}$ in the Irminger basin ($n=10$; Stations 44 and 51) and $7.0 \pm 3.8 \mu\text{mol dpm}^{-1}$ in the Labrador basin ($n=16$;
329 Stations 64, 69 and 77). The decrease of the POC: ^{234}Th ratio with depth illustrated the preferential degradation of
330 carbon relative to ^{234}Th .

331 Because the POC to ^{234}Th ratio has to be determined at the export depth for the conversion of ^{234}Th flux into POC
332 export flux (Buesseler et al., 2006), the POC: ^{234}Th ratios in the LSF and SSF were estimated at this specific depth (Eq
333 in the present study) using the power law interpolation of the measured ratios. The highest POC: ^{234}Th ratios at Eq in
334 the SSF and the LSF were determined in the Labrador basin reaching 16.8 and $13.7 \mu\text{mol dpm}^{-1}$, respectively, at Station



335 69. At most stations, the POC:²³⁴Th ratios at Eq were comparable for both size fractions with differences between the
336 LSF and SSF smaller than a factor 1.7 (Table S2). The highest differences were determined at Station 1 with the
337 POC:²³⁴Th ratios for the LSF being 1.7 fold higher than the one of the SSF, and at Stations 13 and 44 where the
338 POC:²³⁴Th ratios for the SSF being 1.5 and 1.7 fold higher than those of the LSF.

339 3.5. POC export fluxes

340 We estimated the POC export fluxes by multiplying the ²³⁴Th export flux with the POC:²³⁴Th ratio, both determined
341 at the Eq depth. POC fluxes were determined by using the POC:²³⁴Th ratios of the LSF (>53 μm) and the SSF (1-53
342 μm; Table 2) in order to compare both estimations.

343 Except at Stations 1, 26 and 64, the POC fluxes were between 1.1 to 1.5 folds higher when using the SSF ratio.
344 However, when considering the uncertainties, POC fluxes determined with the POC:²³⁴Th ratios in SSF and LSF were
345 not significantly different. Moreover, as we did not have the possibility to compare the ratios with those from sediment
346 traps, we cannot affirm that the small particles participated to the export. As large and rapidly sinking particles usually
347 drive most of the export (Lampitt et al., 2001; Villa-Alfageme et al., 2016), most of the studies dedicated to POC
348 export fluxes in the North Atlantic used the POC:²³⁴Th ratios from the LSF (Ceballos-romero et al., 2016; Le Moigne
349 et al., 2013; Moran et al., 2003; Owens et al., 2014; Sanders et al., 2010; Thomalla et al., 2008). Therefore, in the
350 following, we only discuss the POC fluxes determined with the POC:²³⁴Th ratios from the LSF (Table 3).

351 The POC export fluxes at Eq using the LSF ranged from 1.4 ± 0.5 mmol m⁻² d⁻¹ at Station 44 to 12 ± 22 mmol.m⁻².d⁻¹
352 at Station 1 and the median was 6.1 ± 3.3 mmol m⁻² d⁻¹ (n=11). Besides Station 1 where the POC flux reached 12 ± 22
353 mmol m⁻² d⁻¹, two main open-ocean areas were characterized by high POC export fluxes: 1) the west European and
354 Icelandic basins, in particular Stations 26 and 32 where POC export fluxes reached 7.9 and 8.3 mmol m⁻² d⁻¹
355 respectively and 2) the Labrador Sea basin and in particular Station 69 where POC export flux reached 10 ± 1 mmol
356 m⁻² d⁻¹.

357 3.6. *In-situ* and satellite primary production

358 *In-situ* PP obtained along the GEOVIDE cruise and discussed in this study are presented and argued in more details
359 for the subtropical area (Stations 1, 13 and 21) in Fonseca-Batista et al. (2018; this issue). Across the North Atlantic,
360 the PP varied by a factor of 6, ranging from 27 ± 5 at Station 69 to 174 ± 19 mmol C m⁻² d⁻¹ at Station 26 (Table 3).
361 Low PP were determined in the Iberian basin, with one of the lowest values measured at Station 1 (33 mmol C m⁻² d⁻¹)
362 and a moderate PP at Station 13 (79 mmol C m⁻² d⁻¹; Fonseca-Batista et al., 2018; this issue). The west European
363 basin was highly productive with PP reaching 135 and 174 mmol C m⁻² d⁻¹ at Stations 21 and 26, respectively.
364 Similarly, the Station 32, within the Icelandic basin was highly productive with a PP reaching 105 ± 11 mmol C m⁻² d⁻¹
365 but Station 38 was characterized by a lower PP (68 ± 7 mmol C m⁻² d⁻¹). Within the subpolar area, the PP was high
366 in the Irminger basin, ranging from 137 ± 2 to 166 ± 32 mmol C m⁻² d⁻¹ at Stations 44 and 51, respectively, but the PP
367 was lower in the Labrador basin, ranging from 27 ± 5 to 80 ± 21 mmol C m⁻² d⁻¹ at Stations 69 and 77, respectively.
368 In addition to incubation data, we looked at the annual record of satellite-derived PP in order to document the recent
369 trend in the biological production before the cruise. 8-days averaged PP data for the year 2014 are shown in Figure 4.



370 The intensity and duration of the productive period was highly variable between basins. Most of the stations were
371 sampled during the spring bloom period yet at different stages, except Stations 1 and 13 within the Iberian basin, which
372 were sampled 10 to 12 weeks after the start of the bloom. At these stations, PP increased very early in the year (early
373 to mid-March) and collapsed rapidly (end of March to mid-April) probably due to the setup of oligotrophic conditions
374 (Fonseca-Batista et al., 2018). Northward, the stations in the west European basin were the most productive with the
375 highest PP peak observed at Station 21 ($403 \text{ mmol m}^{-2} \text{ d}^{-1}$), 13 days before the sampling. At Station 26 close to the
376 SAF, the sampling took place during a secondary PP increase. Further north, in the Icelandic and the Irminger basins
377 the spring bloom period started in May. At sampling time, PP was still increasing at Station 32 while the Stations 44
378 and 51 as well as the stations of the Labrador basin (64, 69 and 77) were sampled two to three weeks after the bloom
379 peak.

380 Using the 8-day average data, PP was estimated for the preceding month (32 days) and the whole productive period
381 prior to the sampling date in order to account for different timescales in PP and to compare with export fluxes estimates
382 (Table 3 and Fig. 5). Comparable values (differences smaller than a factor 1.5) were obtained at Stations 13, 21, 32,
383 38, 44, 64 and 77 between *in-situ* and satellite-derived PP data (8-day average). At the other stations, the *in-situ* data
384 were larger, up to 2.3 folds (Station 26), or lower, down to 4 folds (Station 69), compared to the 8-day average satellite
385 data.

386 4. Discussion

387 In the following section, we first discuss the potential impact of the physics and non-steady state conditions on the
388 ^{234}Th export flux estimations, prior to defining different ^{234}Th export and attenuation regimes along the GEOVIDE
389 transect. In line with the ^{234}Th regional variability, the POC export fluxes are discussed with regards to the
390 biogeochemical characteristics of the different basins and compared with published studies in the North Atlantic.
391 Finally, we examine carbon export and transfer efficiencies along the transect.

392 4.1. ^{234}Th export fluxes under the potential influence of physical conditions

393 The GEOVIDE section sampled a diversity of dynamic regimes (Zunino et al., 2017; see Section 2.1), including
394 continental margins affected by strong zonal surface currents (LC, WGC and EGC; Mercier et al., 2015; Reverdin et
395 al., 2003), a local and seasonal upwelling (close to the Iberian Margin), as well as a deep convection zone in the
396 Labrador Sea. In such conditions, the Equation 3, which assumes the physical components (lateral and vertical
397 advective and diffusive fluxes) as negligible, may not always be appropriate (Savoye et al., 2006). Whenever possible,
398 we explore quantitatively or qualitatively, the potential errors introduced in our calculation.

399 Lateral processes associated to high velocities currents and intense mesoscale activity are known to affect the ^{234}Th
400 distribution (Benitez-Nelson et al., 2000; Resplandy et al., 2012; Roca-Marti et al., 2016b; Savoye et al., 2006). In our
401 case, this may concern several stations located at or close to margins such as Stations 51 and 64 subject to the powerful
402 East and West Greenland Currents on the Greenland Margin, Station 77 with the LC on the Newfoundland Margin and
403 Station 1 with the Portugal Current on the Iberian Margin (Fig. 1). However, the impact of the lateral advection cannot
404 be quantified from our dataset, as the required horizontal gradients of ^{234}Th cannot be resolved at a sufficient resolution.



405 As an alternative, we can compare the stations close to each other, as for instance Stations 44 and 51, both located in
406 the Irminger Basin where surface currents are strong. The relatively high variability of the ^{234}Th fluxes (321 and 922
407 $\text{dpm m}^{-2} \text{d}^{-1}$, respectively) found at these two stations may indicate a potential influence of lateral advection. The higher
408 export flux at Station 51 may suggest an input of ^{234}Th depleted waters originating from the Arctic and/or the Greenland
409 shelf. However, Arctic (Cai et al., 2010; Roca-Martí et al., 2016a) and Greenland shelf waters (Station 53, see Table
410 S1) reveal very limited depletions of ^{234}Th relative to ^{238}U . Thus, the ^{234}Th deficit at Station 51 reasonably seems to be
411 essentially driven by vertical rather than horizontal processes.

412 The impact of physics concerns also the open ocean sites, such as stations within the west European and Icelandic
413 basins (Stations 26 and 32) that are subjected to mesoscale activity. An inverse modeling study carried out in that
414 region, at the Porcupine Abyssal Plain site, suggests that the vertical transport of ^{234}Th associated with small-scale
415 structures could represent up to 20% of the estimated vertical export flux (Resplandy et al., 2012). This error is larger
416 than our analytical uncertainty and should be kept in mind when considering the export flux data in this area.

417 The vertical advection can also impact the distribution of ^{234}Th . In upwelling systems, this contribution has been shown
418 to be important (Buesseler, 1998; Buesseler et al., 1995). Near the Portuguese coasts in the Iberian margin, the intensity
419 of the upwelling is seasonally dependent (Costa Goela et al., 2016; Zúñiga et al., 2016) and was rather inactive at the
420 time of the GEOVIDE cruise (<http://marine.copernicus.eu/>). Therefore, the input of ^{234}Th -rich deep waters to the
421 surface is likely to be limited, as already observed in the northern Iberian margin in early summer (Hall et al., 2000).
422 Downwelling systems, such as the intense convection that occurred in the Labrador basin during the winter prior to
423 our sampling (Kieke and Yashayaev, 2015), are also prone to impact the ^{234}Th distribution. However, a strong vertical
424 advection would homogenize the ^{234}Th activities in the water column, which is not the case during our study (Fig. 2).
425 Indeed, the greatest mixed layer depth along the GEOVIDE transect reached 40 m and significant ^{234}Th deficits relative
426 to ^{238}U were observed in surface waters compared to deeper depths. Therefore, the influence of vertical advection on
427 ^{234}Th export fluxes was neglected.

428 Finally, the contribution of the vertical molecular diffusion was estimated using the vertical gradients of total ^{234}Th
429 activity in upper waters and a K_z value ranging between 10^{-4} and $10^{-5} \text{m}^2 \text{s}^{-1}$, as observed in the upper 1000 m between
430 Portugal and Greenland along the OVIDE transect (Ferron et al., 2014). The highest vertical diffusive flux was
431 determined at Station 69 and reached $181 \text{dpm m}^{-2} \text{d}^{-1}$, which is in the range of the ^{234}Th flux uncertainties. Therefore,
432 the impact of the vertical diffusion has not been further considered.

433 In conclusion, although considered to have limited or no impact on the measured ^{234}Th export fluxes, physical
434 processes should be kept in mind when interpreting these export fluxes.

435 4.2. Accounting for non-steady state conditions

436 As the cruise sampling scheme did not allow to collect samples through a time series, it was necessary to assume steady
437 state conditions (i.e., no variation of ^{234}Th activity with time). However, as documented in previous studies in the west
438 European and Icelandic basins (Buesseler et al., 1992; Martin et al., 2011), this assumption was invalid and large
439 variations of ^{234}Th activity were observed at a time scale of one to three weeks along with the onset of the seasonal
440 biological productivity. As a consequence, the SS model was shown to poorly describe the magnitude of the ^{234}Th



441 export flux as it underestimated fluxes by up to a factor 3 compared to the non-steady state (NSS) model (Buesseler et
442 al., 1992; Martin et al., 2011). Indeed, large changes in satellite-derived PP have been observed during the weeks
443 preceding the sampling (Fig. 4, Section 3.5). Most of the stations were sampled during the most productive period of
444 the year except Stations 1 and 13 sampled in post bloom conditions.

445 In order to evaluate the potential error introduced by the SS approach, we have attempted to apply a NSS model.
446 Without time series data, the calculation should not be performed *sensu stricto* (Buesseler et al., 1992; Savoye et al.,
447 2006) but we chose to set the initial conditions for each station, as done by Rutgers van der Loeff et al. (2011) in the
448 South Atlantic (Eq. 4).

449 Satellite-derived PP data were used to estimate the starting date of the bloom, defined by a PP increase of 30 % above
450 the winter value. ^{234}Th was assumed to be in equilibrium with ^{238}U at this time point and the time interval (Δt) for the
451 calculations stretched from the bloom start until the sampling date. All physical terms were considered negligible. The
452 highest NSS ^{234}Th fluxes were determined in the west European and Icelandic basins, reaching $3540 \pm 113 \text{ dpm m}^{-2} \text{ d}^{-1}$
453 at Station 32, while the lowest flux was determined in the Irminger basin ($516 \pm 90 \text{ dpm m}^{-2} \text{ d}^{-1}$ at Station 44; Table
454 1). The NSS ^{234}Th fluxes were all exceeding or equalling those deduced using the SS model because only a decreasing
455 trend of surface ^{234}Th activity (i.e., an increasing deficit) was considered in the NSS model. Also, because the initial
456 ^{234}Th activity is the same for all stations, the differences between NSS and SS fluxes are essentially driven by Δt . For
457 stations sampled shortly after the start of the bloom such as in the Irminger, Icelandic and Labrador basins (Δt ranges
458 from 23 to 43 days), the fluxes predicted by the NSS model are from 1.4 to 2.1 folds higher relative to the SS ones. In
459 the west European and Iberian basins, this difference is reduced (NSS fluxes are from 1.1 to 1.3 folds higher) due to
460 the greater Δt (from 48 to 78 days).

461 As a conclusion, the SS export fluxes may have underestimated ^{234}Th export fluxes at some stations by a maximum
462 factor of 2 in the Icelandic basin. Yet, we need to keep in mind that this NSS approach has limitations by assuming the
463 equilibrium between ^{234}Th and ^{238}U at the bloom start and by considering only an increasing deficit during Δt .

464 4.3. Surface export regimes of ^{234}Th

465 In addition to the export flux (P), we also used the measured partitioning between the particulate and the dissolved
466 phase to estimate the scavenging flux of ^{234}Th (J). As described in Section 2.4, the scavenging flux accounts for the
467 total removal of ^{234}Th from the dissolved phase and thus represents the sum of two contributions: the ^{234}Th sorption
468 flux onto suspended non-sinking particles and the ^{234}Th export flux via sinking particles. The comparison between the
469 export flux (P) and scavenging flux (J) in terms of P/J ratio (export ratio) offers a valuable metric for estimating the
470 export efficiency of ^{234}Th . A low P/J ratio indicates that the removal of dissolved ^{234}Th is controlled by sorption onto
471 suspended particles rather than export. Conversely, a high P/J ratio indicates that the ^{234}Th is preferentially exported
472 rather than adsorbed and is thus efficiently removed from the upper waters.

473 Along the GEOVIDE section, the ^{234}Th export ratios (P/J) varied strongly (Fig. 6), highlighting variable export regimes
474 of ^{234}Th across the North Atlantic. The most striking feature is the very low value determined in the Irminger basin (as
475 low as 0.2 at Station 44) suggesting that export of ^{234}Th is particularly inefficient at this location. The retention of ^{234}Th
476 in the surface layer probably reflects an accumulation phase of biomass in this area. For the other stations, the export



477 ratio is much higher, ranging from ~0.5 (corresponding to a balanced situation between P and J fluxes) to up to >0.75
478 indicating an efficient export of ^{234}Th by sinking particles. Variations can also be observed within the same basin. For
479 instance, in the Iberian and Labrador basins, Stations 1 and 64 close to the Iberian and Greenland margins respectively,
480 display much higher export ratios compared to off-shore stations (Stations 13 and 69, respectively). This reflects
481 different particle dynamics and the more efficient export of ^{234}Th at these margin stations is possibly related to the
482 presence of numerous lithogenic particles (Gourain et al., 2018; Lemaitre et al., in prep.), scavenging the ^{234}Th . The
483 scavenging of ^{234}Th onto non-organic particles has already been observed in the North Atlantic, in particular in benthic
484 nepheloid layers (Owens et al., 2014).

485 4.4. ^{234}Th export flux attenuation in the upper mesopelagic zone

486 Excess of ^{234}Th relative to ^{238}U is indicative of particle break-up and remineralisation by heterotrophic bacteria and/or
487 zooplankton (Benitez-Nelson et al., 2001; Black et al., 2017; Buesseler et al., 2008; Cai et al., 2010; Maiti et al., 2010;
488 Owens et al., 2014; Planchon et al., 2013; Savoye et al., 2004; Usbeck et al., 2002). To estimate the intensity of this
489 remineralisation just below the upper waters, export flux calculations were extended 100 m below the Eq depth. Note
490 that conditions of ^{234}Th excess below Eq yield fluxes integrated until Eq+100 m which are smaller than fluxes
491 integrated over Eq. As reported in Table 1, the reduction of ^{234}Th flux (R100, see Section 2.3) was observed only at a
492 limited number of stations. Evidence for shallow remineralisation (R100 values above uncertainties) can be found in
493 the Iberian basin (Station 13, $R100=410 \pm 218 \text{ dpm m}^{-2} \text{ d}^{-1}$), the west European basin (Station 21, $R100=360 \pm 255$
494 $\text{dpm m}^{-2} \text{ d}^{-1}$) and the Labrador basin (Station 69, $R100=401 \pm 159 \text{ dpm m}^{-2} \text{ d}^{-1}$ and Station 77, $R100=252 \pm 165 \text{ dpm}$
495 $\text{m}^{-2} \text{ d}^{-1}$). Recently, Black et al. (2017) determined R100 values in the southeastern tropical Pacific which are of the
496 same order of magnitude, averaging $400 \pm 200 \text{ dpm m}^{-2} \text{ d}^{-1}$ but with values up to $1200 \text{ dpm m}^{-2} \text{ d}^{-1}$. Relative to the
497 surface export flux, the flux reductions represented a decrease of 30, 20, 50 and 40% at Station 13, 21, 69 and 77,
498 respectively. The significant flux attenuation at Stations 13 and 21 likely reflects an important bacterial activity,
499 reinforced in warm waters (>13°C in the upper 100 m; Iversen and Ploug, 2013; Marsay et al., 2015; Rivkin and
500 Legendre, 2001). In the Labrador Sea, the particulate biogenic Ba_{ex} proxy also revealed evidence of enhanced
501 mesopelagic remineralisation, especially at Station 69 (Lemaitre et al., 2018).
502 For other stations located in the Irminger or Icelandic basins, no apparent decrease in flux was detected suggesting that
503 shallow remineralisation was probably less intense.

504 4.5. Spatial trends in POC export fluxes

505 In the North Atlantic, the intensity and the stage of the bloom, as well as the phytoplankton composition significantly
506 vary spatially. As the ^{234}Th proxy integrates the activity deficits over a timescale of several weeks preceding the
507 sampling and as the phytoplankton size structure and composition are known to exert a control on the magnitude of
508 the POC export flux (Allredge and Jackson, 1995; Boyd et al., 1999; Guidi et al., 2009), it appears essential to compare
509 the spatial variations of these parameters in order to better understand the spatial variability of POC export. In the
510 following, the POC export fluxes are discussed according to the different biogeochemical regions sampled during the
511 survey.



512 *The Iberian basin*

513 One of the lowest POC export flux was determined within the Iberian basin, at Station 13 ($2.2 \pm 0.3 \text{ mmol m}^{-2} \text{ d}^{-1}$)
514 where the PP remained rather low along the season (Fig. 4) and where the highest abundance of pico-phytoplankton
515 was observed (Tonnard et al., in prep.). These conditions are typical of the subtropical and oligotrophic waters (Dortch
516 and Packard, 1989). Villa-Alfageme et al. (2016) highlighted that small cells are usually slow-sinking particles that
517 can be easily remineralised in shallow waters. Their small sinking velocity ($<100 \text{ m d}^{-1}$) allows time for bacteria and
518 zooplankton to degrade these particles, thus reducing the export flux. In the same area, a low flux was also measured
519 later in the season (October) by Owens et al. (2014), confirming the lower carbon export fluxes in this oligotrophic
520 area. Still in the Iberian basin, the highest POC export flux (albeit the strong associated error) was determined at Station
521 1 ($12 \pm 22 \text{ mmol m}^{-2} \text{ d}^{-1}$). As Station 13, Station 1 was sampled after the bloom period and was characterized by low
522 nutrient concentrations but, conversely to Station 13, Station 1 was characterized by a mixed proportion of micro-,
523 nano- and pico-phytoplankton. Moreover, this station was also influenced by lithogenic inputs from the Iberian margin
524 (Gourain et al., 2018; Lemaitre et al., in prep.). The greater proportion of larger size cells, such as diatoms and
525 haptophytes, just like the presence of lithogenic particles, can have increased the sinking speed of the organic matter,
526 leading to a greater POC export flux.

527 *The west European basin*

528 Relatively high POC export fluxes were observed at Stations 21 and 26, reaching respectively 4.8 ± 0.8 and 7.9 ± 5.0
529 $\text{mmol m}^{-2} \text{ d}^{-1}$. The sampling was performed during the bloom, and the highest PP peak along the section was observed
530 at Station 21 (Fig. 4 and 5) just before the sampling. Station 26 was also sampled after a first bloom peak, and these
531 prior-sampling and high PP peaks could have promoted these high exports. These stations were also characterized by
532 an important proportion of micro-phytoplankton communities which could also explain the high POC exports.
533 Diatoms, known for strongly ballasting the POC exports with their dense frustules (Klaas and Archer, 2002), were
534 dominating. The resulting fast-sinking particles could have promoted the relatively high POC export fluxes (Lemaitre
535 et al., in prep.). For the same area, other studies reported similar POC export fluxes during (May; Thomalla et al.,
536 2008), or just after (July-August; Lampitt et al., 2008; Le Moigne et al., 2013) our sampling period. However, Buesseler
537 et al. (1992) have determined much higher POC fluxes (up to $41 \text{ mmol m}^{-2} \text{ d}^{-1}$) in April-May, during the North Atlantic
538 Bloom Experiment, highlighting an important temporal variability in this basin.

539 *The Icelandic basin*

540 One of the highest POC export flux along the transect was determined at Station 32, reaching $8.3 \pm 0.5 \text{ mmol m}^{-2} \text{ d}^{-1}$
541 while the POC flux at Station 38 was $4.8 \pm 0.4 \text{ mmol m}^{-2} \text{ d}^{-1}$. Both stations were sampled during the productive period,
542 although the peak of the bloom was not yet reached (Fig. 4). Nevertheless, the PP at Station 38 remained rather low
543 along the season (Fig. 4) possibly explaining the lower POC export there compared to Station 32. Both stations were
544 dominated by haptophytes, including coccolithophorids (Tonnard et al., in prep.). Their calcium carbonate shells have
545 been shown to promote the export of POC (Francois et al., 2002; Lam et al., 2011; Lemaitre et al., in prep.). In this
546 basin, studies reported higher POC export fluxes, up to $52 \text{ mmol m}^{-2} \text{ d}^{-1}$ (Ceballos-romero et al., 2016; Giering et al.,
547 2016; Martin et al., 2011; Sanders et al., 2010).

548 *The Irminger basin*



549 At higher latitudes, diatoms were dominating the phytoplankton communities. The Irminger basin (Stations 44 and 51)
550 was sampled close to the bloom maximum but, unlike the west European and Icelandic basins, the Irminger basin was
551 characterized by low POC export fluxes (1.4 ± 0.5 and 2.7 ± 0.3 mmol m⁻² d⁻¹, respectively), probably reflecting an
552 accumulation phase of biomass rather than an export phase. Indeed, this area was also characterized by a high
553 proportion of particulate ²³⁴Th in surface waters (reaching 94% of the total ²³⁴Th activity at Station 44) and by a very
554 low P/J ratio indicating that ²³⁴Th was retained in the upper waters rather than exported (Fig. 6; Table 1). In the
555 literature, a relatively large range of POC export fluxes has been observed in this basin. Puigcorbé et al. (2017)
556 observed POC export fluxes ranging from 1.5 to 43 mmol m⁻² d⁻¹. Ceballos-Romero et al. (2016) also determined much
557 higher POC fluxes compared to those observed in the present study, with differences reaching factors of 27 and 19 the
558 month before and after our sampling, respectively.

559 The Labrador basin

560 High POC exports were observed within the Labrador basin and in particular at Station 69 where POC export flux
561 reached 10 ± 1 mmol m⁻² d⁻¹. This basin was dominated by micro-phytoplankton species, such as diatoms, and was
562 sampled just shortly after the peak of PP, indicating the beginning of the decline of the bloom. The combination of the
563 important PP a few weeks before our sampling and the decline of the diatom bloom likely triggered the high POC
564 export fluxes, as observed elsewhere (Martin et al., 2011; Roca-Marti et al., 2016b; Stange et al., 2016). As for the
565 Irminger basin, Puigcorbé et al. (2017) determined a low POC export (0.7 mmol m⁻² d⁻¹) the month before our sampling
566 period, while Moran et al. (2003) observed higher fluxes reaching 47 mmol m⁻² d⁻¹ in July, one month after our
567 sampling period.

568 Overall, POC exports varied strongly along the transect with a factor of 8.6 between the highest and the lowest POC
569 export flux. The magnitude of these fluxes seems to be dependent on the phytoplankton community structure and thus
570 on the particle composition and density (Buesseler, 1998; Francois et al., 2002; Honjo, 1996; Lam et al., 2011). The
571 influence of the ballast effect on POC export is discussed in more details in a companion paper (Lemaitre et al., in
572 prep.). We have also shown that the magnitude of the POC exports were dependent on the evolution of the bloom, with
573 high exports during post bloom periods. Studies using deep sediment traps showed these high export events were
574 driven by large and rapidly sinking aggregates (Lampitt et al., 2001, 2010; Turner, 2002). The comparison with the
575 literature also suggests that POC export fluxes are strongly variable temporally. Indeed, in only 1 month time lag, POC
576 fluxes can vary up to a factor of 27 (Ceballos-Romero et al., 2016; Fig. 7) confirming the fast changes of the
577 biogeochemical parameters controlling the sinking particles and thus the export fluxes.

578 **4.6. Export and transfer efficiencies of POC**

579 In order to study the biological carbon pump in the North Atlantic, we used two parameters: the export efficiency
580 (ThE), which is calculated by dividing the POC export flux at Eq by the PP (Buesseler, 1998) and the transfer efficiency
581 (T100) which is calculated by the POC export flux at 100 m below Eq divided by the POC export flux at Eq (Fig. 8).
582 Note that the POC export flux at Eq+100 (Table 3) has been calculated by multiplying the ²³⁴Th flux at Eq+100 by the
583 POC to ²³⁴Th ratio of large particles at this same depth.



584 Considering the *in-situ* PP (Table 3), ThE ranged from 1 (Station 44) to 38% (Station 69) with a median value of 7%
585 along the transect. The highest export efficiencies were determined at Stations 1 and 69 where ThE reached 35 and
586 38%, respectively. Other stations were characterized by $\text{ThE} \leq 14\%$ with a higher range (7 – 14%) at Stations 32, 38,
587 64 and 77. Export efficiencies around 10% are common in the open ocean (Buesseler, 1998). Lower export efficiency
588 can be related to important microbial and zooplankton grazing activities as to biomass accumulation in surface waters
589 (Planchon et al., 2013, 2015). On the contrary, ratios greater than 10% highlight an efficient export of the PP out of
590 the surface layer. High ThE can be caused by many processes such as the presence of large and/or dense and fast
591 sinking particles, low surface remineralisation, active zooplankton migration or nutrient stress (Ceballos-romero et al.,
592 2016; Le Moigne et al., 2016; Planchon et al., 2013). Interestingly, stations with the highest ThE were also
593 characterized by the lowest PP (Stations 1 and 69) while stations with the lowest ThE were characterized by the highest
594 PP (Stations 44 and 51). This inverse relationship between PP and ThE was significant for all stations of the GEOVIDE
595 cruise (regression slope: -0.20; $r^2=0.58$; $p<0.01$; $n=11$; Fig. S2) and has been explained in the Southern Ocean by the
596 temporal decoupling between PP and export (Henson et al., 2015), biomass accumulation in surface waters (Planchon
597 et al., 2013), and other processes such as zooplankton grazing and bacterial activities (Maiti et al., 2013; Le Moigne et
598 al., 2016; Roca-Marti et al., 2016a). Indeed, efficient recycling of particles in the upper waters has been observed in
599 the North Atlantic due to high microbial or grazing activities (Collins et al., 2015; Giering et al., 2014; Marsay et al.,
600 2015) limiting the POC export to the deep ocean. A recent study in the Icelandic and Irminger basins highlighted the
601 importance of the bloom dynamics on the particle export efficiency suggesting a strong seasonal variability of the ThE
602 (Ceballos-Romero et al., 2016). Our estimates are generally in the lower range of export efficiencies reported for the
603 North Atlantic with values ranging from 1 to 42% in the western European basin (Buesseler et al., 1992; Lampitt et
604 al., 2008; Thomalla et al., 2008), from 5 to 8% in the Icelandic basin (Ceballos-romero et al., 2016), from 4 to 16% in
605 the Irminger basin (Ceballos-romero et al., 2016) and from 4 to >100% in the Labrador basin (Moran et al., 2003).
606 This large range confirms that export efficiencies are highly variable with time and that the North Atlantic during the
607 period of our study seemed to behave like most of the highly productive areas of the world's ocean, with a low export
608 efficiency.

609 However, the ThE calculation is based on two parameters that are integrating processes over different time scales: 24
610 h for *in-situ* PP and several weeks for export. As a result of this temporal mismatch and due to the strong variability in
611 PP, ThE ratios were also estimated using the satellite-derived 8-day, 32-day and seasonal PP (Table 3). As seen in
612 Section 3.6, there are no significant differences between the satellite PP estimates regardless of the integrations times,
613 and thus no significant differences between the ThE values, except at Stations 1 and 69. Indeed, the ThE values
614 decrease from 35 to 12% and from 38 to 8% respectively at Stations 1 and 69, suggesting an *in-situ* PP unusually low
615 during our study, leading to an over-estimated ThE. At the other stations along the transect, no significant ThE changes
616 were observed regardless of the temporal PP integration.

617 Carbon transfer efficiencies (T100) ranged from 30 (Station 69) to 78% (Station 32). They were characterized by
618 greater error bars (see Fig. 8) due to the greater incertitude on the deep POC export flux. The highest T100 were
619 observed within the Icelandic basin at Stations 32 and 38 with values reaching 78 and 74%, respectively. On the
620 contrary, the lowest T100 values were determined at Stations 1, 13, 21 and 69 (46, 33, 49 and 30%, respectively) and
621 highlight firstly, an important regional variability, also reported elsewhere (Lam et al., 2011; Lutz et al., 2002), and



622 secondly, a greater carbon remineralisation between Eq and Eq+100 m at these latter stations. As already discussed for
623 the R100 values (see Section 4.4), the low T100 values observed in the eastern part of the transect may be explained
624 by an important bacterial activity, reinforced in warm waters. This efficient recycling is characteristic of tightly coupled
625 regeneration-based microbial food webs of oligotrophic regimes (Karl, 1999; Thomalla et al., 2006), such as at Stations
626 1, 13 and, to a lesser extent, at Station 21. In the Icelandic basin (Stations 32 and 38), the high T100 may be related to
627 the important abundance of coccolithophorids (Tonnard et al., in prep.) known to enhance the POC export flux to the
628 deep ocean by ballast effect (Francois et al., 2002; Lam et al., 2011). Indeed, Bach et al. (2016) highlighted that a
629 bloom of coccolithophorids can increase the transfer efficiency through the mesopelagic layer by 14-24%. Finally, the
630 Labrador and Irminger basins exhibit relatively similar T100 (between 50 and 69%), except at Station 69 where we
631 determined the lowest T100. This is also in agreement with the highest R100 (Section 4.4) and carbon remineralisation
632 flux determined with the $B_{a_{xs}}$ proxy (Lemaitre et al., 2018).

633 5. Conclusion

634 As part of the GEOTRACES program, the GEOVIDE GA01 section allowed us to investigate the ^{234}Th and POC
635 export fluxes across the North Atlantic during spring 2014.

636 The export of ^{234}Th through sinking particles was particularly efficient at stations close to the margins and in the
637 Icelandic basin (Stations 1, 64 and 38), while the Irminger basin was characterized by an important retention of ^{234}Th
638 in surface waters. This could be due the development of the bloom leading to an accumulation phase of biomass rather
639 than an export phase. Close to the margins, the abundance of lithogenic particles may have enhanced the ^{234}Th
640 scavenging and its subsequent removal to deeper levels in the water column. ^{234}Th fluxes were also calculated 100 m
641 below Eq in order to investigate remineralisation processes. ^{234}Th attenuation appeared more intense at Stations 13 and
642 21 characterized by warm waters, reinforcing the bacterial activity, and in the Labrador Sea. In the latter area, the
643 particulate biogenic $B_{a_{xs}}$ proxy also revealed evidence of enhanced mesopelagic remineralisation, especially at Station
644 69 (Lemaitre et al., 2018).

645 The carbon export fluxes varied by a factor ~ 9 along the transect highlighting an important spatial variation. In the
646 North Atlantic, some studies reported similar POC export estimates but some others determined much higher POC
647 export fluxes, up to a factor of 27 in only 1 month lag, confirming the high temporal variation of the POC export fluxes
648 in this ocean, as shown by studies using fixed sediment traps (Antia et al., 2001; Billet et al., 1983; Lampitt et al.,
649 2010; Peinert et al., 2001).

650 Different factors were identified for controlling the POC export fluxes regionally and temporally:

- 651 i) The magnitude of the POC export flux is directly related to the intensity and the stage of the bloom.
652 During the bloom, an accumulation of biomass in surface water may induce a limitation of the POC
653 export fluxes while exports can increase during the decline of the bloom, likely due to increasing numbers
654 of rapidly sinking particles.
- 655 ii) The phytoplankton size structure might have influenced the magnitude of the POC export fluxes. Indeed,
656 the only station characterized by pico-phytoplankton communities was characterized by one of the lowest
657 POC export flux. However, the areas composed by nano- or micro-phytoplankton were both



658 characterized by high POC export fluxes, indicating that the size structure was not the main factor
659 influencing the fluxes during GEOVIDE.

660 iii) The phytoplankton community seems to impact the particle composition and density, which play a crucial
661 role on the particulate sinking velocities and thus on the magnitude of the POC export fluxes. The highest
662 POC export fluxes were determined at stations where diatoms or coccolithophorids dominated,
663 suggesting the importance of the ballast effect in the North Atlantic (Lemaitre et al., in prep.).

664 For most stations, the fraction of primary production that is exported from the surface zone (export efficiency) was
665 $\leq 14\%$, which is in agreement with the global ocean export efficiency ($\sim 10\%$; Buesseler, 1998). Export efficiency was
666 also inversely related to primary production, highlighting that the North Atlantic during our study seems to behave like
667 most of the highly productive areas of the world's ocean, with a low export efficiency. Finally, the fraction of POC
668 that is not remineralised in the mesopelagic zone (transfer efficiency) fits within the range of measured transfer
669 efficiencies reported elsewhere (e.g., Black et al., 2017; Buesseler and Boyd, 2009). The highest transfer efficiencies
670 were determined at stations where coccolithophorids dominated.

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977 **Table 1:** Summary of the ^{234}Th export and scavenging fluxes using steady state (SS) and non-steady state (NSS) models. The
 978 ^{234}Th export fluxes using the SS model are calculated at the depths corresponding to the bottom of the primary production
 979 zone (PPZ), the equilibrium (Eq) depth and 100 m below Eq (Eq+100); the latter being used to estimate a remineralisation
 980 flux of ^{234}Th (R100). Negative R100 values indicate an increase of the export flux between Eq and Eq+100. Note that the
 981 depth was fixed to 100 m at Station 26 because of the lower sampling vertical resolution. Consequently, the export flux at
 982 Eq+100 and the R100 were not determined at Station 26.

Basin	Station		Export depth m	Th export (SS)		Th export (NSS)		Th scavenging (SS)	
				dpm m ⁻² d ⁻¹		dpm m ⁻² d ⁻¹		dpm m ⁻² d ⁻¹	
Iberian	1	PPZ	155	1327	± 137	1442	± 80	1509	± 189
		Eq	90	1264	± 104				
		Eq+100	190	1348	± 199				
		R100		-84	± 224				
	13	PPZ	82	1247	± 99	1588	± 86	2898	± 285
		Eq	110	1418	± 111				
		Eq+100	210	1008	± 187				
		R100		410	± 218				
West European	21	PPZ	82	1723	± 82	2352	± 70	3917	± 212
		Eq	110	1873	± 97				
		Eq+100	210	1513	± 235				
		R100		360	± 255				
	26	PPZ	95	1432	± 117	1968	± 98	2839	± 220
		Fixed	100	1486	± 117				
Icelandic	32	PPZ	75	1455	± 92	3540	± 113	3690	± 199
		Eq	130	2282	± 1519				
		Eq+100	230	2200	± 227				
		R100		81	± 256				
	38	PPZ	70	1136	± 80	2345	± 115	1495	± 160
		Eq	80	1134	± 95				
		Eq+100	180	949	± 151				
		R100		185	± 178				
Irminger	44	PPZ	37	321	± 66	516	± 90	1802	± 71
		Eq	40	321	± 66				
		Eq+100	140	454	± 114				
		R100		-132	± 132				
	51	PPZ	37	495	± 67	1625	± 108	2189	± 260
		Eq	100	922	± 103				
		Eq+100	200	873	± 114				
		R100		49	± 154				
Labrador	64	PPZ	83	853	± 129	1423	± 122	1142	± 192
		Eq	80	855	± 95				
		Eq+100	180	733	± 200				
		R100		123	± 221				
	69	PPZ	35	684	± 57	1068	± 53	1257	± 112
		Eq	40	758	± 57				
		Eq+100	140	357	± 148				
		R100		401	± 159				
77	PPZ	55	693	± 77	1169	± 75	1529	± 148	
	Eq	60	696	± 77					
	Eq+100	160	444	± 146					
	R100		252	± 165					



983 **Table 2:** Comparison of the steady state POC export fluxes at Eq as determined using the POC:²³⁴Th ratios in the large
984 (LSF; > 53 μm) and small size fraction (SSF; 1-53 μm).

Station #	LSF POC flux mmol m ⁻² d ⁻¹	SSF POC flux mmol m ⁻² d ⁻¹
1	12 ± 22	6.9 ± 2
13	2.2 ± 0.3	3.3 ± 0.6
21	4.8 ± 0.8	6.3 ± 1.4
26	7.9 ± 5.0	6.1 ± 3.7
32	8.3 ± 0.5	8.8 ± 0.5
38	4.8 ± 0.4	5.2 ± 0.7
44	1.4 ± 0.5	2.4 ± 0.5
51	2.7 ± 0.3	3.8 ± 0.5
64	7.8 ± 1.5	5.5 ± 4.9
69	10 ± 1	13 ± 1
77	6.1 ± 1.5	7.5 ± 0.9

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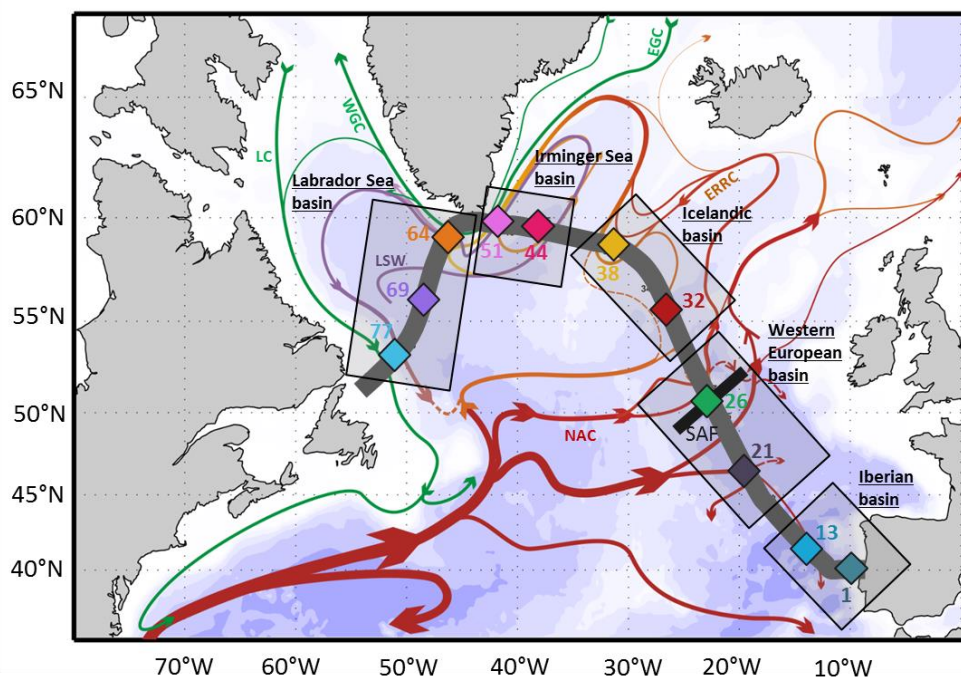
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987 **Table 3:** POC (particulate organic carbon) to ^{234}Th ratios (in $\mu\text{mol dpm}^{-1}$), POC export fluxes (in $\text{mmol m}^{-2} \text{d}^{-1}$) at the Eq depth, *in-situ* PP (Fonseca-Batista et al., 2018 and
 988 this study) and satellite-derived PP from the Vertically Generalized Production Model (VGPM) integrated over 8 days, 32 days and over the whole season (in $\text{mmol m}^{-2} \text{d}^{-1}$)
 989 and the POC fluxes at Eq+100 m (in $\text{mmol m}^{-2} \text{d}^{-1}$). Because of the lower vertical sampling resolution at Station 26, no POC export flux was determined at Eq+100. *The
 990 sampling to determine the *in-situ* PP at Station 51 occurred 24h after the sampling of the particulate ^{234}Th and POC.

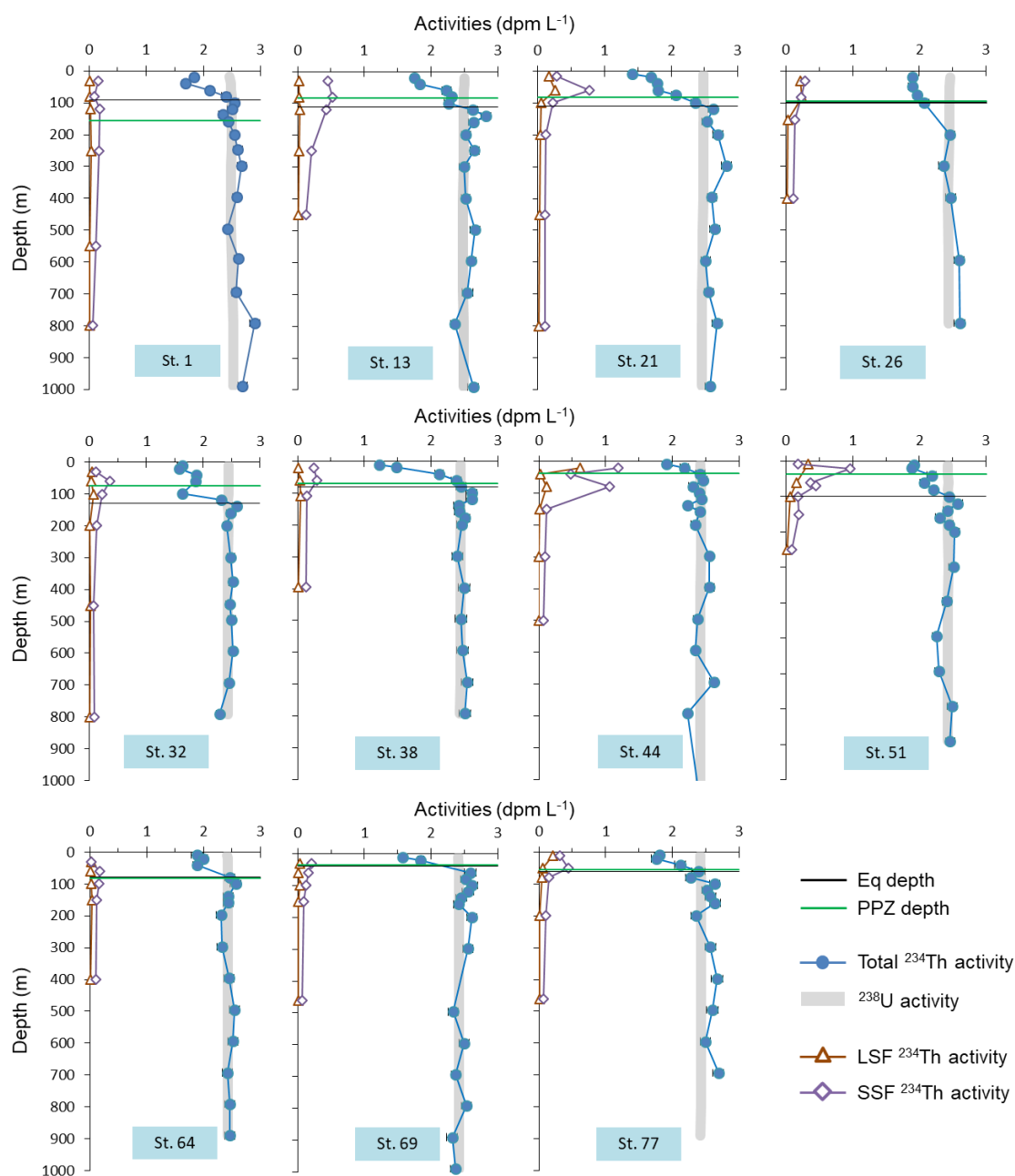
Station #	POC: ^{234}Th at Eq $\mu\text{mol dpm}^{-1}$	POC flux at Eq $\text{mmol m}^{-2} \text{d}^{-1}$	<i>in-situ</i> PP $\text{mmol m}^{-2} \text{d}^{-1}$	8-days VGPM based PP $\text{mmol m}^{-2} \text{d}^{-1}$	32-days VGPM based PP $\text{mmol m}^{-2} \text{d}^{-1}$	seasonal VGPM based PP $\text{mmol m}^{-2} \text{d}^{-1}$	POC flux at Eq+100 $\text{mmol m}^{-2} \text{d}^{-1}$
1	9 ± 17	12 ± 22	33 ± 2	76 ± 3	80 ± 11	96 ± 62	5.3 ± 23.2
13	1.6 ± 0.2	2.2 ± 0.3	79 ± 3	64 ± 7	72 ± 18	81 ± 63	0.7 ± 0.2
21	2.6 ± 0.4	4.8 ± 0.8	135 ± 2	161 ± 21	260 ± 97	201 ± 119	2.3 ± 0.4
26	5.3 ± 3.3	7.9 ± 5.0	174 ± 19	77 ± 14	74 ± 19	112 ± 59	
32	3.6 ± 0.1	8.3 ± 0.5	105 ± 11	105 ± 7	95 ± 13	87 ± 13	6.5 ± 0.7
38	4.2 ± 0.1	4.8 ± 0.4	68 ± 7	82 ± 5	94 ± 34	109 ± 32	3.5 ± 0.6
44	4.4 ± 1.3	1.4 ± 0.5	137 ± 2	89 ± 3	110 ± 65	101 ± 66	0.8 ± 0.4
51	2.9 ± 0.01	2.7 ± 0.3	*166 ± 32	95 ± 7	125 ± 118	125 ± 118	1.7 ± 0.2
64	9.2 ± 1.1	7.8 ± 1.5	54 ± 18	59 ± 18	109 ± 115	103 ± 122	4.9 ± 1.5
69	14 ± 0.04	10 ± 1	27 ± 5	108 ± 8	134 ± 80	134 ± 80	3.1 ± 1.3
77	8.8 ± 1.9	6.1 ± 1.5	80 ± 21	108 ± 8	134 ± 80	134 ± 80	3.1 ± 1.3

991



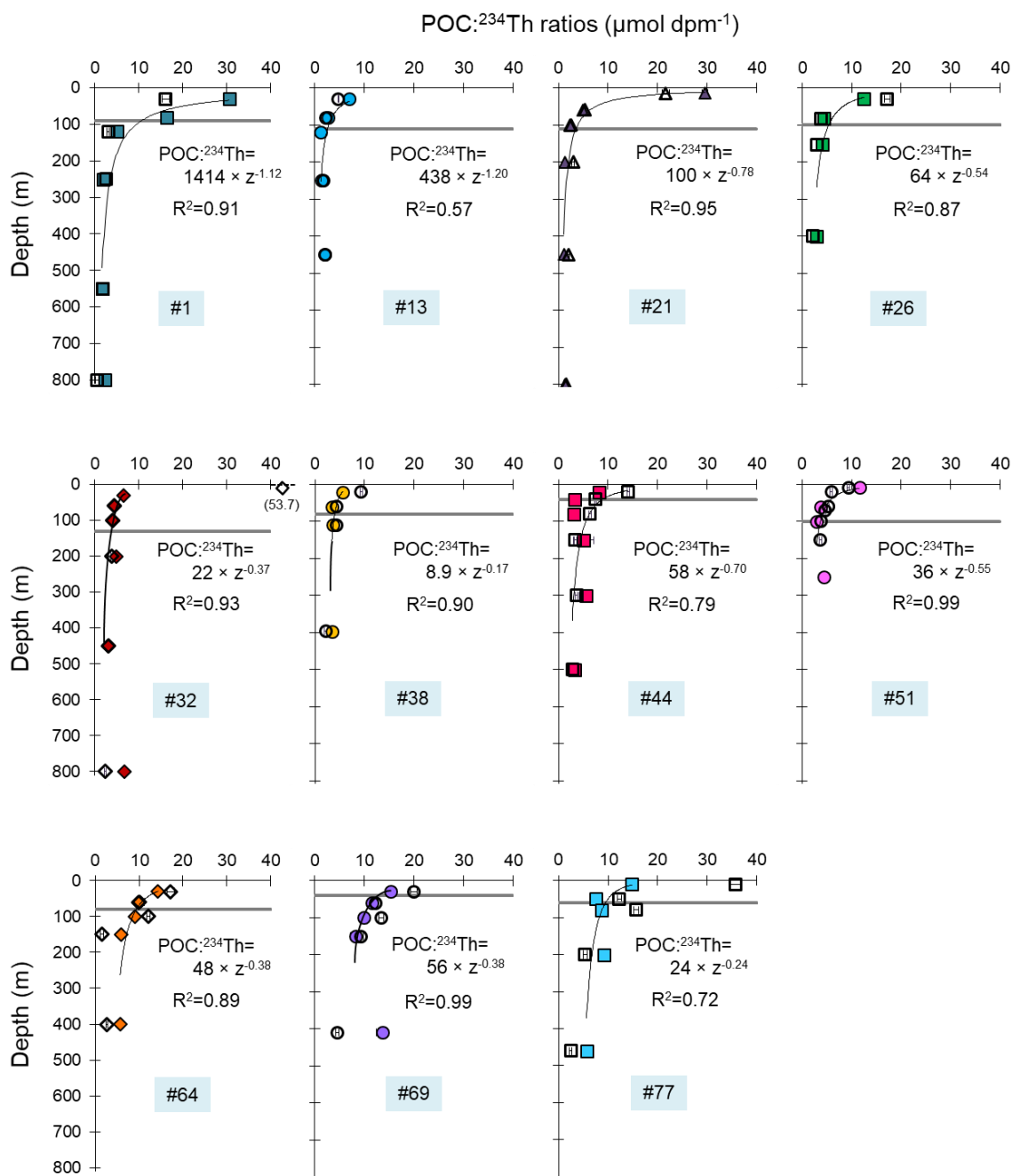
992

993 **Figure 1:** Simplified schematic of the surface circulation in the North Atlantic (adapted from Danialt et al., 2016)
994 superimposed with the GEOVIDE cruise track (thick grey line) and stations (colored diamonds). Main surface currents are
995 indicated: East Greenland Current (EGC), West Greenland Current (WGC), Labrador Current (LC), Eastern Reykjanes
996 Ridge Current (ERRC), North Atlantic Current (NAC). The Sub-Arctic Front (SAF) and the Labrador Seawater (LSW)
997 when in surface (i.e. within the Labrador basin) are also represented. Station colors are reused in the following figures.



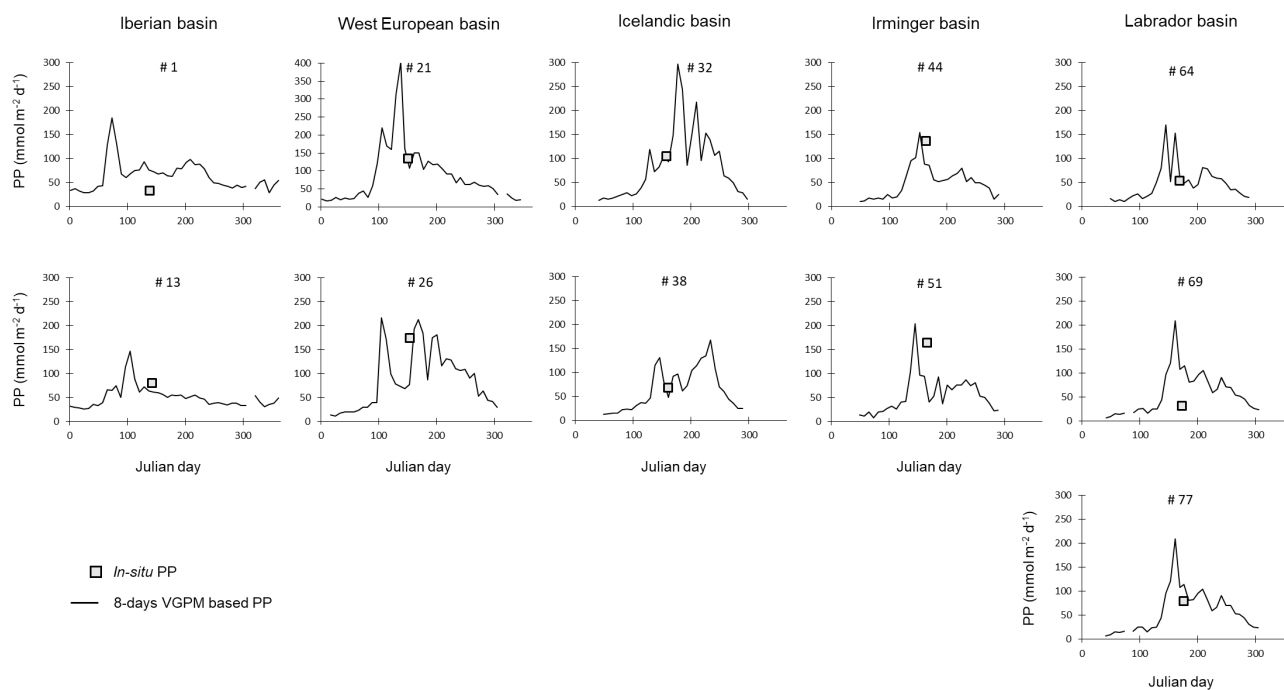
998

999 **Figure 2:** Profiles of the total ²³⁴Th (closed blue circles), total ²³⁸U (thick grey vertical line) and particulate ²³⁴Th activities
 1000 for the small size fraction (SSF; 1-53 μm; open diamonds) and for the large size fraction (LSF; >53 μm; open triangles). All
 1001 activities are expressed in dpm L⁻¹. The horizontal black line is the Eq depth (depth where ²³⁴Th returns to equilibrium with
 1002 ²³⁸U), and the horizontal green line is the depth of the PPZ (primary production zone). Error bars are smaller than the size
 1003 of the symbols



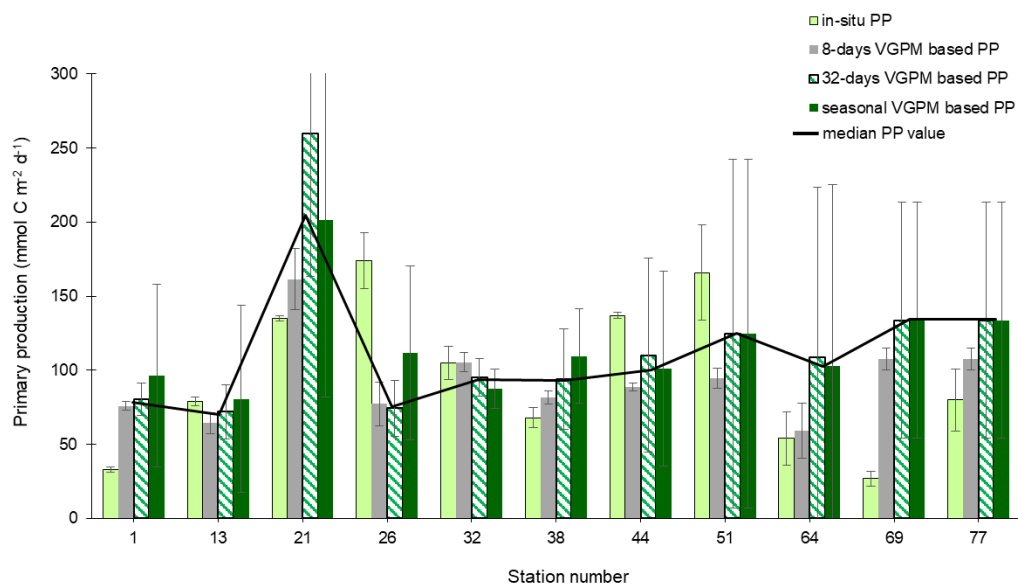
1004

1005 **Figure 3:** Profiles of the POC:²³⁴Th ratios (μmol dpm⁻¹) in the SSF (open symbols) and LSF (closed symbols). The Eq depth,
 1006 where ²³⁴Th is back to equilibrium with ²³⁸U, is indicated with the grey horizontal line. The thin black line represents the
 1007 power law fit (POC:²³⁴Th=a×Z^{-b}) of the LSF.



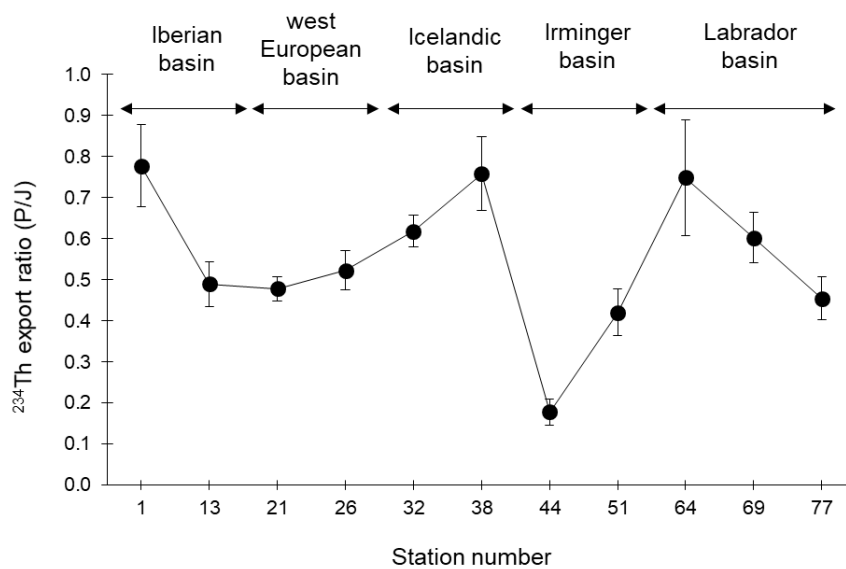
1008

1009 **Figure 4:** *In-situ* (squares) and satellite-derived (continuous lines) primary production (PP; in $\text{mmol m}^{-2} \text{d}^{-1}$) data at the time of our sampling and along the year 2014.



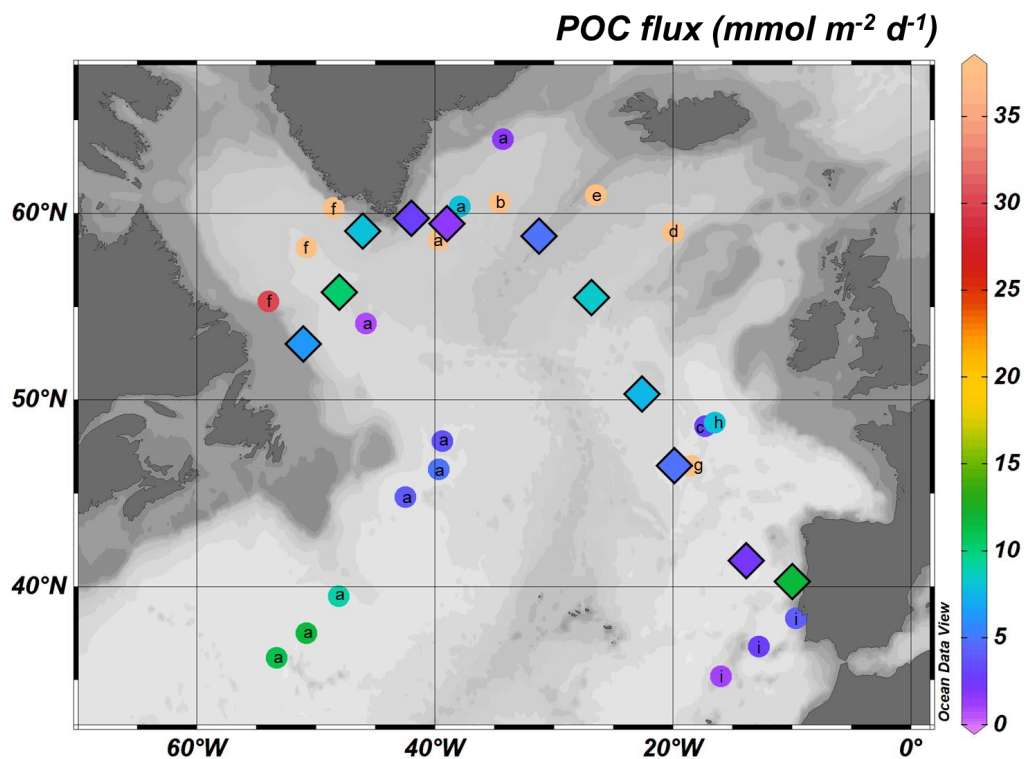
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Figure 5: Comparison of *in-situ* and satellite (8-days, 32-days and seasonal averages) primary productivities ($\text{mmol C m}^{-2} \text{d}^{-1}$) along the GEOVIDE transect. The median value is also indicated (black line).



1013

1014 **Figure 6:** Variability of the ^{234}Th export ratio (i.e., the ratio of the ^{234}Th export flux over the ^{234}Th scavenged flux; P/J ratio)
1015 along the GEOVIDE section.

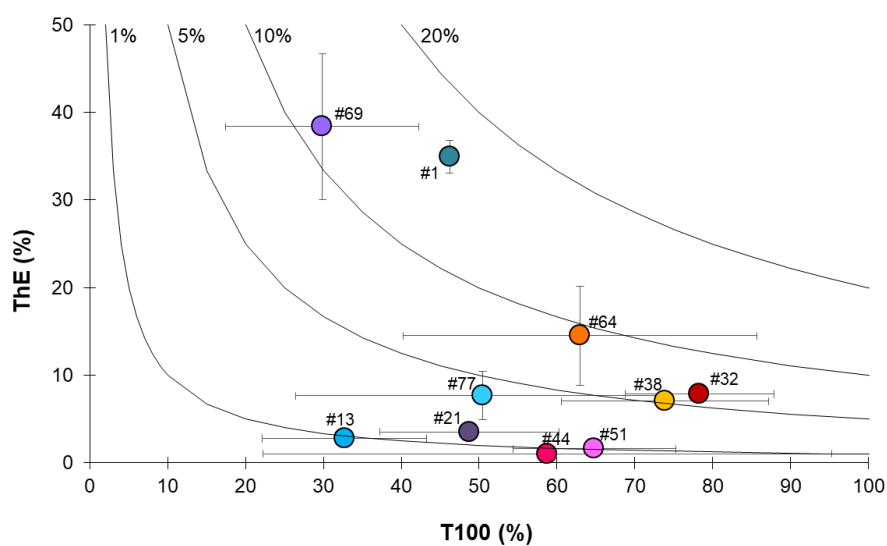


1016

1017 **Figure 7:** Comparison of the POC export fluxes from this study (diamonds with black borders) with other ²³⁴Th-derived
1018 estimates of POC exports in the North Atlantic (a: Puigcorb  et al., 2017; b: Ceballos-Romero et al., 2016; c: Thomalla et
1019 al., 2008; d: Sanders et al., 2010; e: Martin et al., 2011; f: Moran et al., 2003; g: Buesseler et al., 1992; h: Le Moigne et al.,
1020 2013; i: Owens et al., 2015).



1021



1022

1023 **Figure 8:** Export efficiency (ThE = Export at Eq / *in-situ* PP) versus transfer efficiency (T100 = Export flux at Eq+100 /
1024 Export flux at Eq). The black lines represent the modelled 1, 5, 10 and 20% of PP exported to depths > Eq+100 m.