

1 **Carbon sequestration in the deep Atlantic enhanced by Saharan dust**

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11 **Enhanced atmospheric input of dust-borne nutrients and minerals to the remote surface**  
12 **ocean can potentially increase carbon uptake and sequestration at depth. Nutrients can**  
13 **enhance primary productivity, and mineral particles act as ballast, increasing sinking**  
14 **rates of particulate organic matter. Here we present a unique 2-year time-series of**  
15 **sediment-trap observations of particulate organic carbon flux to 3000 m depth, measured**  
16 **directly in two locations: the dust-rich central North Atlantic gyre and the dust-poor**  
17 **South Atlantic gyre. We find that carbon fluxes are twice as high and a higher proportion**  
18 **of primary production is exported to depth in the dust-rich North Atlantic gyre. Low**  
19 **stable nitrogen isotope ratios suggest that high fluxes result from the stimulation of**  
20 **nitrogen fixation and productivity following the deposition of dust-borne nutrients.**  
21 **Sediment traps in the northern gyre also collected intact colonies of nitrogen-fixing**  
22 ***Trichodesmium* species. Whereas ballast in the southern gyre is predominantly biogenic,**  
23 **dust-derived mineral particles constitute the dominant ballast element during the**  
24 **enhanced carbon fluxes in the northern gyre. We conclude that dust deposition increases**

25 **carbon sequestration in the North Atlantic gyre through the fertilisation of the nitrogen-**  
26 **fixing community in surface waters and mineral ballasting of sinking particles.**

27 Flux of airborne desert dust into the surface ocean can increase the amount of  
28 photosynthetically fixed carbon dioxide (CO<sub>2</sub>) by reducing nutrient limitation of primary  
29 production and thus increase the flux of particulate organic carbon (POC) to the deep ocean<sup>1</sup>.  
30 Dense dust-derived lithogenic particles can also increase particle size through aggregation and  
31 enhance sinking velocity and preservation of POC through ballasting, allowing more carbon to  
32 penetrate deeper into the ocean's interior<sup>2</sup>. The impact of dust input on downward POC flux  
33 can be especially important in the subtropical low-nutrient low-chlorophyll (oligotrophic)  
34 gyres which occupy 60% of the global ocean surface<sup>3</sup> and thus are likely large sinks for  
35 atmospheric CO<sub>2</sub>. Even relatively small changes in downward POC flux in these immense areas  
36 would significantly affect the global carbon budget. However, the transport of organic carbon  
37 (*i.e.* Biological Carbon Pump) in oligotrophic regions is very poorly understood, and large  
38 uncertainties remain over the impact of enhanced dust deposition on the magnitude of POC  
39 flux below the depth of winter mixing (sequestration).

40 We tested the hypothesis that enhanced dust deposition increases POC sequestration in  
41 remote low-nutrient low-chlorophyll provinces by directly measuring downward deep POC  
42 flux in the centres of the subtropical North and South Atlantic gyres. The study regions  
43 represent permanently stratified systems characterised by restricted nutrient advection, and  
44 hence extremely low surface concentrations of macronutrients (nitrate and phosphate) and  
45 chlorophyll. Here, picoplankton dominate community structure<sup>4</sup>, while heterotrophic bacteria  
46 and cyanobacteria govern ecosystem metabolism, channelling a large proportion of POC into  
47 the microbial loop<sup>5</sup>, thus diminishing its export out of the euphotic zone. The subtropical North  
48 Atlantic, however, receives large depositional fluxes of Saharan dust with associated essential  
49 nutrients (e.g. nitrogen, phosphorus, iron)<sup>1</sup> blocked from the South Atlantic region by the Inter-

50 Tropical Convergence Zone<sup>6</sup>. The Fe-rich surface waters of the northern gyre are favoured by  
51 N<sub>2</sub> fixing microbes (diazotrophs) that generate bioavailable nitrogen for other phytoplankton<sup>7</sup>,  
52 thereby allowing for a higher proportion of primary production to be converted into sinking  
53 POC than would otherwise occur. Phosphate co-limits the Fe-induced N<sub>2</sub> fixation and  
54 production fuelled by atmospheric nitrogen<sup>8, 9</sup>. This can exert an important control over the  
55 amount of POC ultimately produced from diazotrophic and atmospheric sources and  
56 subsequently available for export. Along with this fertilisation effect, increased lithogenic  
57 particle concentration following dust input can also facilitate POC flux to depth through  
58 additional incorporation of dense dust particles<sup>10</sup>. Biomineral ballasting is otherwise regulated  
59 by calcite which is typically found in both gyres<sup>11</sup>. However, the degree to which lithogenic  
60 ballasting can drive the increased POC sedimentation would itself be limited by the amount of  
61 POC present<sup>12</sup>.

### 62 **Field observations in the central Atlantic gyres**

63 We directly captured POC flux in the central Atlantic gyres from 2007 to 2010 using  
64 sediment traps moored at 3000 m depth at sites NOG (23°N 41°W) and SOG (18°S 25°W)  
65 (Fig.1). During this period, NOG was subjected to, on average, ten-fold higher dust deposition  
66 compared to SOG (Fig. 2a), as inferred from dust concentration measurements over Barbados<sup>13</sup>  
67 for NOG and modelled data<sup>14, 15</sup> for SOG (Methods). At both sites, the average surface  
68 production rates derived from a Vertically Generalised Production Model (VGPM)<sup>16</sup> were  
69 lower than much of the global ocean<sup>17</sup>, and on average 23% higher at NOG than at SOG (Fig.  
70 2b). The observed POC fluxes to the trap at NOG (0.40-2.7 mg C m<sup>-2</sup> d<sup>-1</sup>; mean = 1.06 mg C m<sup>-2</sup>  
71 d<sup>-1</sup>) were always at least two-fold higher than at SOG (0.21-0.95 mg C m<sup>-2</sup> d<sup>-1</sup>; mean = 0.49  
72 mg C m<sup>-2</sup> d<sup>-1</sup>) (Fig. 2d, 3). The POC fluxes at NOG and SOG were significantly lower than the  
73 depth-normalised values reported for the oligotrophic sites in the western North Atlantic gyre  
74 (station OFP (BATS))<sup>18</sup> and subtropical North Pacific gyre (station ALOHA)<sup>19</sup>, and hence they

75 are the lowest measured in the global ocean. From the ratios of POC flux to VGPM primary  
76 production (both variables were averaged over the trap deployment period) we calculate almost  
77 double the fraction of surface production reaching 3000 m depth at NOG (0.60%) compared to  
78 SOG (0.37%). These very low values are similar to the records at BATS (0.59%)<sup>18</sup> and imply  
79 an overall more efficient downward POC transport in the dusty northern gyre. Lithogenic flux  
80 determined from aluminium concentrations in trap material was significantly lower at SOG  
81 than at NOG (Fig. 2c) and elsewhere in the subtropical North Atlantic<sup>18,20</sup>, indicating that the  
82 inter-basin differences in dust deposition propagated to depth. The NOG data bridge the  
83 previous observations of deep lithogenic flux in the eastern and western parts of the northern  
84 gyre<sup>18,20</sup> showing the westward gradient of decreasing deep lithogenic fluxes driven by the  
85 weakening of the Saharan dust transport towards the northwest-Atlantic<sup>13</sup>.

86         Although higher at NOG, at both sites, POC flux was enhanced during late summer-  
87 autumn (>120% of the annual mean value; Fig. 3), a period of warm sea-surface temperature  
88 (25.0 - 28.2 °C), relatively shallow mixed layer (<50 m), and low surface chlorophyll  
89 concentrations (<0.04 mg m<sup>-3</sup>; Supplementary Fig. 1). At NOG, the average dust input during  
90 summer-autumn (14.4±8.9 mg m<sup>-2</sup> d<sup>-1</sup>) exceeded the wintertime values (8.87±11.6 mg m<sup>-2</sup> d<sup>-1</sup>)  
91 (Fig. 3a). An enhanced input of dust-borne nitrogen, phosphorus and iron has likely occurred  
92 during this period. After nitrogen is exhausted by the dust-stimulated primary producers, the  
93 warm and strongly stratified water column would offer optimal conditions for enhanced N<sub>2</sub>  
94 fixation provided there is enough iron and phosphate present to satisfy cellular demands of  
95 diazotrophs<sup>7,8,9</sup>. Bloom-forming *Trichodesmium* spp. dominate diazotrophic biomass in the  
96 region of NOG<sup>21</sup>. Elevated N<sub>2</sub> fixation rates by these diazotrophs were reported during  
97 summer-autumn (median 34.9 μmol N m<sup>-2</sup> d<sup>-1</sup>) compared to winter-spring (median 12.2 μmol  
98 N m<sup>-2</sup> d<sup>-1</sup>) (refs<sup>21,22,23,24,25</sup>). This coincides with higher fluxes of aerosol iron in autumn than  
99 in spring<sup>26</sup> and higher surface concentrations of dissolved iron in the early autumn<sup>27</sup> (1.0-1.3

100 nmol L<sup>-1</sup>) than in winter<sup>28</sup> (0.18-0.54 nmol L<sup>-1</sup>). Lower phosphate concentrations measured in  
101 the central northern gyre during summer have also been attributed to the enhanced diazotrophic  
102 activity exhausting the phosphate pool<sup>8</sup>. Remarkably, we find a strikingly high POC flux of up  
103 to 2.7 mg C m<sup>-2</sup> d<sup>-1</sup> in August-September 2009 at NOG (Fig. 3a). This relatively short POC  
104 export pulse, never seen at SOG, accounted for 29% of total POC sequestered at NOG during  
105 2007-2009 and greatly exceeded the mean wintertime POC flux at NOG (0.88±0.13 mg C m<sup>-2</sup>  
106 d<sup>-1</sup>) and the daily flux at SOG. A notable presence of some intact *Trichodesmium* “tufts” (Figs.  
107 3a, 4) within this pulse suggests a potential involvement of these diazotrophs in driving the  
108 extreme POC sequestration event at NOG. Similarly short and efficient POC export pulses to  
109 > 2800 m depth have been regularly observed at ALOHA following a summertime increase in  
110 productivity and biomass of diatom-diazotroph symbiotic phytoplankton<sup>19</sup>.

#### 111 **Fertilisation effect of dust**

112 We measured markedly low stable nitrogen isotope ratios in the trap material ( $\delta^{15}\text{N}_{\text{PN}}$ ,  
113 in ‰ relative to air) from the dust-rich NOG (range 0.40-1.32‰; mass-weighted mean 0.77‰),  
114 indicating that isotopically light nitrogen introduced by enhanced N<sub>2</sub> fixation and potentially  
115 atmospheric deposition<sup>29</sup> significantly contributed to sinking particles. Some of this low  $\delta^{15}\text{N}$   
116 signal might have originated south of NOG (10°-16°N), before being transported to and  
117 accumulated at the NOG thermocline as low  $\delta^{15}\text{N}_{\text{nitrate}}$  during northward water mass transit<sup>7, 30</sup>.  
118 However, a strong inverse correlation between  $\delta^{15}\text{N}_{\text{PN}}$  and POC flux ( $R^2=0.67$ ,  $p=0.001$ ) with  
119 *Trichodesmium* “tufts” present at the lowest  $\delta^{15}\text{N}_{\text{PN}}$  values (Figs. 4, 5), is suggestive of a direct  
120 link between elevated POC flux at NOG and a local supply of newly fixed nitrogen by  
121 diazotrophs whose activity was likely stimulated by substantial inputs of dust-borne iron and  
122 phosphorus. Observations at NOG are qualitatively similar to those at ALOHA<sup>19</sup>, where  $\delta^{15}\text{N}_{\text{PN}}$   
123 minima and diazotroph-driven particulate POC flux maxima are closely associated. Dust

124 deposition, which is a substantial source of isotopically light nitrogen in the region ( $8.5 \mu\text{mol}$   
125  $\text{m}^{-2} \text{d}^{-1}$ ; ref<sup>31</sup>) could augment the deep POC flux lowering its  $\delta^{15}\text{N}$  signature.

126 In contrast to NOG, sinking particles from the dust-poor SOG carried significantly  
127 heavier  $\delta^{15}\text{N}_{\text{PN}}$  of 3.70‰ to 4.41‰ (mass-weighted mean 4.07‰). This is similar to the oceanic  
128 average  $\delta^{15}\text{N}$  of deep-water nitrate (4.8‰; ref<sup>29</sup>), and hence this source was probably fuelling  
129 primary production at SOG.

130 The deep  $\delta^{15}\text{N}_{\text{PN}}$  at NOG and SOG fit a broad range of  $\delta^{15}\text{N}$  values reported for  
131 particulate nitrogen in the upper waters of the central North and South Atlantic gyres<sup>32, 33</sup>  
132 (Supplementary Fig. 2). At both sites, trap material was  $^{15}\text{N}$ -enriched compared to the particles  
133 suspended in the euphotic zone (top 130 m) likely due to fractionation resulting from  
134 remineralisation processes in both the surface and mesopelagic (Ref 34). Similar  $\delta^{15}\text{N}$  values  
135 for trap material and particles from 150-160 m depth may also point to a potentially important  
136 contribution of heavier  $\delta^{15}\text{N}$  signal formed at the deep chlorophyll maximum to  $\delta^{15}\text{N}_{\text{PN}}$ .

137 We estimated the contribution of different nitrogen sources to  $\delta^{15}\text{N}_{\text{PN}}$  at NOG and SOG  
138 using a two-end member nitrogen mass-balance model<sup>29</sup> (see Methods and references therein).  
139 We assumed that the isotope budget of the mixed layer in the permanently oligotrophic gyres  
140 incorporates nitrogen supplied by diazotrophs, by vertical diffusion across the nitrate  
141 concentration gradient, and from dust (NOG only). We also assumed negligible isotopic  
142 fractionation following complete nitrogen assimilation by phytoplankton. The average isotopic  
143 signature of diazotrophic biomass ( $-1 \pm 1$ ‰) was used as the  $\text{N}_2$  fixation endmember. The upper  
144 thermocline nitrate endmember was represented by  $\delta^{15}\text{N}$ -nitrate averaged over the depth of the  
145 nitrate gradient spanning the euphotic layer at NOG ( $2.73 \pm 0.36$ ‰) and SOG ( $6.22 \pm 0.35$ ‰).  
146 The dust-derived nitrogen endmember was assigned  $\delta^{15}\text{N}$  of  $-3.1$ ‰ based on the average  
147 isotopic composition of bulk aerosols influenced by Saharan dust. Using these endmember  
148 values, we find that local  $\text{N}_2$  fixation could contribute on average  $50.4 \pm 8.4\%$  to the isotopic

149 signal of nitrogen sequestration at NOG, while aerosol nitrogen alone (if all bioavailable) could  
150 account for  $32.4 \pm 5.4\%$  (Supplementary Table 1). The relative contribution of diazotrophs to  
151  $\delta^{15}\text{N}_{\text{PN}}$  at NOG was higher than that at BATS (33%; at average  $\delta^{15}\text{N}_{\text{PN}} = +1\%$  (ref<sup>34</sup>) and nitrate  
152  $\delta^{15}\text{N} = +2.6\%$  (ref<sup>35</sup>)) and at ALOHA (range 21-48%; refs<sup>19, 36</sup>), where eddy transfer and lateral  
153 advection are important mechanisms of nitrogen supply<sup>36, 37</sup>. At SOG, newly fixed nitrogen  
154 contributed a smaller, yet considerable portion of  $\delta^{15}\text{N}_{\text{PN}}$  ( $29.7 \pm 3.1\%$ ), possibly owing to the  
155 activity of unicellular cyanobacteria, major  $\text{N}_2$  fixers in the South Atlantic<sup>7, 21</sup>. We, however,  
156 acknowledge a significant uncertainty of these results due to an overall lack of time-resolved  
157  $\delta^{15}\text{N}$  data for the surface nitrate and dust at the trap sites. Moreover, our budgets did not account  
158 for a possible origin of particles from a specific trophic level (*e.g.* faecal pellets) and alteration  
159 of  $\delta^{15}\text{N}_{\text{PN}}$  due to isotopic fractionation during particle remineralisation and transformation in  
160 the mesopelagic. However, regardless of these uncertainties, the isotope budgets suggest a large  
161 systematic difference in the contribution of newly fixed local nitrogen inputs between the North  
162 and South Atlantic gyres which likely contributes to the two-fold inter-basin difference in POC  
163 sequestration. Our observations thus set an important quantitative constraint on the downward  
164 flux of low  $\delta^{15}\text{N}$  material sinking to the subtropical North Atlantic. They provide compelling  
165 evidence for the origin of an isotopically light nitrate reservoir in the subtropical North Atlantic  
166 supporting previous observations (*e.g.* ref<sup>30</sup>).

167         The unique presence of intact *Trichodesmium* colonies in the deep particles at NOG  
168 (Fig. 4) indicates that *Trichodesmium* biomass is not always lost in the surface waters as  
169 previously assumed<sup>38, 39</sup>, but can leave the euphotic zone and contribute to POC export. It is  
170 possible that the “tufts” reached the abyssal depth at NOG in a rapidly sinking ( $>200 \text{ m d}^{-1}$ )  
171 *Trichodesmium* bloom, collapsed through viral lysis or programmed cell death<sup>39</sup>. Since Fe  
172 starvation at NOG is unlikely, exhaustion of bioavailable phosphorus<sup>8</sup> during the summer  
173 might be major triggers of the bloom collapse. Alternatively, the “tufts” might represent

174 *Trichodesmium* populations that migrated towards the phosphocline to “mine” phosphate but  
175 were unable to return to the light<sup>40</sup>. Finally, *Trichodesmium* can retain dust particles within  
176 their morphologically intricate colonies to accelerate Fe dissolution from dust<sup>41</sup>. Trapped dust  
177 particles may therefore “ballast” *Trichodesmium* colonies, increasing their density and  
178 allowing them to sink rapidly to depth and avoid remineralisation or grazing. This could partly  
179 explain the temporal coherence between low  $\delta^{15}\text{N}$ , elevated dust, POC, and lithogenic fluxes  
180 during late summer at NOG (Fig. 3a).

### 181 **Ballasting effect of dust**

182 Higher dust input significantly altered the composition of particles at NOG compared  
183 to SOG (Fig. 2e). Dust-derived lithogenic material was the second largest contributor  
184 ( $34.3 \pm 11.6\%$ ) to the total mass at NOG after calcite, whereas at SOG this value was  $4.7 \pm 2.3\%$ ,  
185 consistent with the difference in the amount of dust being deposited at each site (Fig. 2a).  
186 Although the seasonal signal of elevated dust flux at both sites was largely lost at 3000 m depth,  
187 we still observed elevated lithogenic flux at NOG ( $>120\%$  of the annual average) in winter  
188 2008 and summer-autumn 2008 and 2009 concurrently with the increased POC flux and  
189 following high dust input (Fig. 3a). Assuming that this temporal coherence was not accidental,  
190 we investigated the relative involvement of lithogenic and biogenic (opal + calcite) ballast  
191 phases in enhanced POC sequestration. Based on the outputs of the mineral-associated POC  
192 flux model and multiple linear regression analysis<sup>2, 42</sup> (Methods), 41.0% of POC flux at SOG  
193 was ballasted by lithogenic material. This, however, might be an overestimation driven by a  
194 relatively large carrying coefficient for lithogenic ballast (0.371) which resulted from a nearly  
195 1:1 ratio of POC to lithogenic flux and their strong positive correlation (Spearman’s  $p=0.91$ ).  
196 At NOG the percentage of POC ballasted by lithogenic particles increased from 45.7% during  
197 low POC flux to 70.1% during high flux in the summer-autumn (Supplementary Table 2).  
198 Overall, lithogenic material appears to be a more important ballast for POC in the central



199 northern gyre compared to its western boundary (25%), where lithogenic fluxes are lower and  
200 opal fluxes are ten times higher<sup>18</sup>. We suggest that at NOG elevated dust inputs may shift the  
201 dominant ballasting phase from biogenic to lithogenic, increasing POC flux to the deep ocean.  
202 This is likely achieved through a sudden increase in mineral particle concentration following  
203 dust deposition and subsequent stimulation of aggregation of organic matter, including that of  
204 diazotrophs, in the surface waters<sup>10</sup>. Moreover, clay particles, constituting >60% of the aerosol  
205 dust over the central North Atlantic<sup>43</sup>, are denser (2.79 g cm<sup>-3</sup>) than biomineral calcite (2.65 g  
206 cm<sup>-3</sup>) and opal (2.1 g cm<sup>-3</sup>), and thus would likely increase sinking velocity of POC upon  
207 aggregation. Although currently debated in the literature (e.g. refs.<sup>44, 45</sup>), lithogenic ballast  
208 might have also exert an enhanced protective effect on POC compared to calcite. Recent  
209 laboratory experiments<sup>45, 46</sup> demonstrated slower degradation rates for clay-ballasted POC  
210 relative to calcite-ballasted POC. The existence of such protective effect of lithogenic material  
211 is yet to be shown in the field.

### 212 **Mechanism of dust-induced enhancement of carbon sequestration**

213 Lithogenic particles did not represent the main ballasting phase for POC during periods  
214 of high and low lithogenic fluxes and were not associated with the biomineral fluxes at NOG  
215 (Supplementary Table 2). The ballasting ability of lithogenic particles at NOG appears to be  
216 confined to the summer-autumn period (Fig. 3a) when the surface fertilisation by dust was  
217 potentially the strongest. This tight temporal coupling suggests that the presence of additional  
218 fresh organic (i.e. fertilisation effect) matter might be required to activate effective lithogenic  
219 ballasting while lithogenic particles are critical to transport the fertilisation effect to the deep  
220 ocean. The variability in mineralogy and morphology of dust arriving at NOG from different  
221 locations in the Sahara during winter<sup>47</sup> and summer may have also impacted both fertilisation  
222 and ballasting properties of dust.

223 Overall, enhanced POC sequestration in the dust-rich NOG suggests that in the vast  
224 nutrient-limited Atlantic, the strength of the biological carbon pump could be significantly  
225 lower without concurrent dust-induced fertilisation and ballasting. The observed two-fold  
226 enhancement of POC sequestration under a ten-fold higher dust (iron) input at NOG further  
227 points to a potentially important role of phosphate in setting the upper bound for the Fe-driven  
228 enhancement of POC export. However, fertilisation could also stimulate the activity of  
229 heterotrophic bacteria, increasing remineralisation and a corresponding reduction of carbon  
230 export<sup>5</sup>.

231 Under the current climatic trends, the subtropical oligotrophic gyres are predicted to  
232 expand over the coming centuries<sup>48</sup>. Multi-decadal observations of dust concentrations over  
233 Barbados have already revealed a weakening of dust transport from North Africa to the North  
234 Atlantic as a function of increasing sea-surface temperature<sup>13</sup>. Predicted changes in wind  
235 patterns are expected to continue altering dust deposition into the ocean and hence input of  
236 nutrients and mineral ballast<sup>49</sup>. In parallel, ongoing ocean acidification might affect  
237 bioavailability of essential nutrients, including iron<sup>50</sup>. All these perturbations will certainly  
238 alter POC sequestration in the oligotrophic gyres, and hence global climate, in the coming  
239 centuries. Therefore, our study urges for a better understanding of the present Biological  
240 Carbon Pump functioning in the nutrient-limited oceans.

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458

459

460 **Figure captions**

461 **Figure 1| Chlorophyll and dust deposition flux in the Atlantic Ocean. a**, annual composite  
462 Moderate-resolution Imaging Spectroradiometer chlorophyll-*a* concentration ( $\text{mg m}^{-3}$ ) in 2009.  
463 Oligotrophic gyres are represented by dark blue areas of low chlorophyll concentrations ( $<0.1$   
464  $\text{mg m}^{-3}$ ). **b**, basin-wide annually averaged (1974-2004) modelled dust deposition flux re-plotted  
465 from ref<sup>14</sup>. Yellow triangles indicate the locations of the NOG and SOG sediment trap  
466 moorings, which are also on the annually repeated Atlantic Meridional Transect (AMT) line  
467 ([www.amt-uk.org](http://www.amt-uk.org)). The black solid line shows the AMT-19 cruise track (Oct-Nov 2009)  
468 passing through the NOG and SOG sites. Dashed lines indicate an approximate north-south  
469 boundary of the Inter-Tropical Convergence Zone.

470 **Figure 2| Surface ocean and deep particle flux data for the study sites. a-d**, mean  $\pm$  standard  
471 deviation values over the respective trap deployment periods. **a**, dust deposition flux (n=25 for  
472 NOG and n=26 for SOG). **b**, depth-integrated primary production derived from the  
473 chlorophyll-based Vertically Generalised Production Model<sup>16</sup> (Methods) **c**, lithogenic flux  
474 (n=40 for both sites). **d**, POC flux (n=40 for both sites). **e**, composition of sediment trap  
475 material. The height of the stacked bars represents total particle mass flux.

476 **Figure 3| Time-series fluxes at NOG (a) and SOG (b)**. The dust deposition (monthly values)  
477 and aluminium-derived lithogenic fluxes are presented on a logarithmic scale. The uncertainty  
478 of the dust flux to the South Atlantic is estimated to be at least a factor of 10 (ref<sup>14</sup>). For POC  
479 and lithogenic fluxes, the width of each bar corresponds to 14- or 21-day collection interval.  
480 Red circles depict stable nitrogen isotopic composition of particles ( $\delta^{15}\text{N}_{\text{PN}}$ ) from the selected  
481 cups. Arrows and a letter “T” indicate the cups where *Trichodesmium* spp. “tufts” were found.  
482 Summer-autumn periods are highlighted in yellow.

483

484 **Figure 4| *Trichodesmium* spp. “tufts” from the summer POC flux pulse at NOG.** Tufted  
485 colonies of *Trichodesmium* spp. cells were identified in the cups collecting in August and  
486 September 2009. This is the first record of *Trichodesmium* being exported to bathypelagic  
487 depth (>1500 m).

488 **Figure 5| POC flux vs. isotopic composition of the trap material ( $\delta^{15}\text{N}_{\text{PN}}$ ) from NOG (blue**  
489 **circles) and SOG (red circles).** The black line is the best fit line of the linear model. Arrows  
490 with letter “T” mark the cups where *Trichodesmium* tufts were found. The strong inverse  
491 relationship between the magnitude of POC flux and  $\delta^{15}\text{N}_{\text{PN}}$  at NOG signifies a potentially  
492 important role of local input of isotopically light N from  $\text{N}_2$  fixation (and dust deposition) in  
493 enhancing carbon sequestration at this site.

494

495 **Methods**

496 **Particle collection and processing.** Sinking particles were collected using 21-cup time-series  
497 Parflux Mark 78H–21 sediment traps (McLane Research Laboratories, USA) deployed on a  
498 bottom-tethered mooring at a depth of 3000 m in a water depth of >4200 m. At NOG, the traps  
499 were deployed from November 4, 2007 to October 5, 2008 and from November 23, 2008 to  
500 October 25, 2010, collecting particles over a total of 672 days. At SOG, the traps operated from  
501 May 11, 2008 to May 20, 2009 and from May 24, 2009 to June 20, 2010, collecting particles  
502 over a total of 766 days. Each trap cup collected for 14 or 21 days. Sample preservative  
503 consisted of a solution of sodium chloride (5 g L<sup>-1</sup>), di-sodium tetra-borate (0.25 g L<sup>-1</sup>), and  
504 formalin (5% vol/vol) made up with deep seawater. Upon recovery, pH was measured and  
505 found to be between 8.0 and 8.3. One mL of concentrated formalin solution was then added to  
506 the cups to supplement the existing formalin. Sample processing was carried out under dust-  
507 and metal-free conditions in a laminar flow cabinet using plastic- or glass-ware only. Prior to  
508 all analyses, zooplankton “swimmers” were identified under stereo-microscope (Meiji Techno,  
509 Japan) fitted with a photo-camera (Canon EOS-1000, Japan) and handpicked using PTFE-  
510 coated tweezers (Dumont, Switzerland) and a plastic pipette (Fisher Scientific, UK). The  
511 preservative/particle mixture in each cup was then split into 8 sub-samples using a custom-  
512 built rotary PVC splitter. Individual sub-samples from each cup were filtered, dried at 40°C  
513 and analysed for particulate organic carbon, opal, calcite, and trace metals including  
514 aluminium. Selected sub-samples were also analysed for stable nitrogen isotope composition.

515 **Chemical analyses of the trap material.** Particulate organic carbon (POC) was measured in  
516 tin capsules (HEKAtech GmbH) after removing carbonate by *in situ* acidification<sup>51</sup> with  
517 concentrated hydrochloric acid and using a high-temperature combustion technique on a CHN  
518 analyser (HEKAtech GmbH EURO EA CHNS-O Elemental Analyser) with analytical  
519 precision of <0.1%). The median filter blank contribution to POC signal was 2.7%. The



520 calculated limit of detection (LoD; based on three times standard deviations of the filter blanks)  
521 was 8.26  $\mu\text{g}$  (n=20). Particulate Organic Matter (POM) was calculated as  $2.2 \times \text{POC}$  (ref<sup>2</sup>). Splits  
522 for calcite were prepared by leaching in  $0.4 \text{ mol L}^{-1}$  nitric acid with calcium content measured  
523 by inductively coupled plasma optical emission spectrometry<sup>52</sup> (Perkin-Elmer Optima 4300DV  
524 ICP-OES; analytical precision of <1%). Procedural blanks consisting of unused polycarbonate  
525 membranes treated with nitric acid contributed <1% to Ca signal. The LoD of the blank-  
526 corrected Ca measurements was wavelength-dependent, ranging from 0.012 to 0.015  $\mu\text{g}$   
527 (n=10). Calcite mass flux was calculated by multiplying calcium-derived flux of particulate  
528 inorganic carbon by a factor of 8.3. Samples for opal were digested in  $0.2 \text{ mol L}^{-1}$  sodium  
529 hydroxide, neutralized with  $0.1 \text{ mol L}^{-1}$  hydrochloric acid and analysed as dissolved silicate on  
530 a SEAL QuAATro auto-analyser<sup>52, 53</sup>. The detection limit of the instrument was 0.3  $\mu\text{g}$ . The  
531 median contribution of procedural blanks was 3.1%. The LoD of the filter-blank corrected  
532 samples was run-dependent ranging from 1.19 to 11.5  $\mu\text{g}$  (n=9). Opal was calculated to be  $2.4$   
533  $\times$  biogenic silica flux assuming 10% water content<sup>52, 53</sup>. Labile and refractory fractions of  
534 aluminium in trap material were determined<sup>54</sup>. The labile fraction was extracted with 25%  
535 (vol/vol) acetic acid at room temperature, and then the more refractory fraction was fully  
536 digested in a mixture of concentrated nitric and hydrofluoric acids at  $150^\circ\text{C}$ . The residues of  
537 both fractions were redissolved in  $0.5 \text{ mol L}^{-1}$  nitric acid and analysed by inductively coupled  
538 plasma-mass spectrometry (Thermo Fisher Scientific Element 2 XR HR-ICPMS). The LoD of  
539 blank corrected aluminium measurements was  $0.12 \text{ ng g}^{-1}$  (n=10); the concentrations in acid  
540 mix and blank filters were  $0.764 \pm 0.8 \text{ ng g}^{-1}$ ; (n=10) and  $0.843 \pm 0.917 \text{ ng g}^{-1}$  (n=8), respectively.  
541 The accuracy of the measurements was established using a range of Certified Reference  
542 Materials, including HISS-1, NIST-1648a and NIST-1573a. The recoveries in these reference  
543 materials were 97.3-104.1% for aluminium. Total trace metal concentration was determined by  
544 adding leach and digest metal fractions. Total aluminium mass flux was used to calculate

545 lithogenic mass flux based on aluminium content of 7.1% in Saharan dust<sup>55</sup> and 7.7% in  
546 Patagonian dust<sup>56</sup> for NOG and SOG samples respectively. Stable nitrogen isotopic  
547 composition of sinking particulate nitrogen pool ( $\delta^{15}\text{N}_{\text{PN}}$ ) was determined from  $^{14}\text{N}/^{15}\text{N}$  mass  
548 ratio measured using Micro Cube elemental analyser (Elementar Analysensysteme GmbH,  
549 Hanau, Germany) interfaced to a PDZ Europa 20-20 isotope ratio mass spectrometer (Sercon  
550 Ltd., Cheshire, UK). The accuracy of the measurements was established using a set of  
551 laboratory standards calibrated against NIST Standard Reference Materials (IAEA-N1, IAEA-  
552 N2, IAEA-N3, USGS-40, and USGS-41). The analytical precision of the  $\delta^{15}\text{N}_{\text{PN}}$  measurements  
553 was <0.1‰, while the difference between duplicates ranged between 3.1 and 11% (n=4).  
554 Measurements were performed at the UC Davies Stable Isotope Facility, USA.

555

556 **Dust deposition flux.** Direct and time-resolved measurements of dust deposition at NOG and  
557 SOG are not available. At SOG we obtained monthly estimates of dust deposition using an  
558 atmospheric model<sup>14, 15</sup>, which utilizes reanalysis data (a combination of model and  
559 observations) to drive a dust chemical transport model, and was compared to long-term  
560 measurements of aerosol concentration. Dust deposition flux was modelled in four bins with  
561 the size distribution range of 0.1-0.5, 0.5-1.0, 1.0-2.5, and 2.5-10  $\mu\text{m}$ . Dust deposition velocities  
562 were calculated within the model as a function of meteorological conditions and resulted in  
563 averages of 0.01, 0.029, 0.115, and 0.674  $\text{cm s}^{-1}$  over our region. The modelled dust deposition  
564 fluxes were averaged for  $3^\circ \times 3^\circ$  area centred at the SOG location. The uncertainty of the model  
565 output for the South Atlantic Ocean is hypothesised to be at least a factor of 10 due to scarcity  
566 and uncertainties in observational data and uncertainties in model source, transport and  
567 deposition processes<sup>14</sup>. Dust deposition flux at NOG was inferred from time-series dust  
568 concentrations measured over Barbados which is heavily influenced by air-masses from Sahara  
569 and Sahel deserts<sup>13</sup>. The details of dust sampling and processing are described in ref<sup>13</sup>. Dust

570 deposition flux was calculated by multiplying dust concentrations by a range of deposition  
571 velocities ( $0.01\text{-}1.2\text{ cm s}^{-1}$ ) characteristic of relatively fine mineral dust aerosols of  $<5\text{ }\mu\text{m}$  in  
572 size typically arriving to the remote open ocean<sup>57</sup>. The resulting average dust deposition flux  
573 at NOG ranged from  $0.085\text{ to }10.2\text{ mg m}^{-2}\text{ s}^{-1}$ . Assuming a deposition velocity of  $1\text{ cm s}^{-1}$ , dust  
574 deposition flux is similar in magnitude to deep lithogenic flux at NOG. Thus, we considered  
575 this deposition velocity to be the most appropriate for calculations of daily dust deposition flux  
576 at NOG.

577

578 **Upper ocean hydrography.** Eight-day composite sea-surface temperature (SST) data were  
579 recorded by the Moderate Resolution Imaging Spectroradiometer (MODIS) sensor of NASA's  
580 Aqua satellite at  $9\text{ km}$  resolution and averaged for  $3^{\circ}\times 3^{\circ}$  box centred at each trap location. The  
581 annual cycle of mixed layer depth at the trap sites was derived from the ARGO-based  
582 climatology<sup>58</sup> averaged for  $3^{\circ}\times 3^{\circ}$  area over the trap sites. The base of the mixed layer was  
583 defined as the depth at which the density was  $0.03\text{ kg m}^{-3}$  less than that at  $10\text{ m}$ .

584

585 **Ancillary biogeochemical datasets** were provided by the British Oceanographic Data Centre  
586 (BODC) and include vertical profiles of chlorophyll (archived data under accession numbers  
587 SOC050136 and SOC110235), nitrate concentrations (refs<sup>59, 60</sup> and archived dataset under  
588 accession number MIT130172), isotopic composition of total nitrate (refs<sup>61, 62, 63</sup>), nitrogen  
589 fixation rates (refs<sup>22, 23, 24, 25</sup>),  $^{14}\text{C}$ -based primary production rates (ref<sup>64</sup> and archived data with  
590 accession numbers PP-PML090162, PP-PML110236 and PP-PML120146),

591

592 **Primary production.** Depth-integrated daily rates of primary production for the relevant time  
593 period were estimated from the chlorophyll-based eight-day resolved Vertically Generalized  
594 Production Model (VGPM)<sup>16</sup> and averaged for the  $3^{\circ}\times 3^{\circ}$  area centred at the trap sites. The

595 VGPM data were downloaded from the Ocean Productivity website  
596 (<http://www.science.oregonstate.edu/ocean.productivity/>). Within relevant time periods, the  
597 VGPM-based productivity rates at NOG ( $160\pm 14$  mg C m<sup>-2</sup> d<sup>-1</sup>) and SOG ( $139\pm 18$  mg C m<sup>-2</sup>  
598 d<sup>-1</sup>) were comparable to the values measured directly at the trap sites in October-November  
599 2008-2011 ( $240\pm 96$  mg C m<sup>-2</sup> d<sup>-1</sup> at NOG and  $204\pm 84$  mg C m<sup>-2</sup> d<sup>-1</sup> at SOG (see ancillary  
600 biogeochemical datasets above).

601

602 **Surface chlorophyll-*a* concentration.** Eight-day composite surface chlorophyll-*a* data were  
603 recorded by MODIS Aqua at 9 km resolution and averaged for 3°×3° box centred at each trap  
604 location. MODIS Aqua calculates near-surface chlorophyll concentrations from a model of  
605 ocean colour using an empirical relationship.

606

607 **Contribution of newly fixed nitrogen to the stable nitrogen isotope signal in trap material.**

608 The  $\delta^{15}\text{N}$  of the trap material reflects both the autotrophic particle formation and the subsequent  
609 heterotrophic transformations. In the latter, the diagenetic fractionation can potentially alter  
610  $\delta^{15}\text{N}$  of the bulk nitrogen export and sequestration. No significant relationship was observed  
611 between C/N ratios and  $\delta^{15}\text{N}$  of nitrogen export at NOG ( $r^2=0.02$ ,  $n=12$ ) while at SOG, this  
612 relationship was positive but weak and insignificant ( $r^2=0.25$ ,  $n=12$ ). This suggests that the  
613 observed variations in  $\delta^{15}\text{N}$  of the trap material were determined predominantly during algal  
614 production, with no significant influence from detrital material and/or non-phytoplankton  
615 organisms<sup>65</sup>. At both sites isotopic fractionation following nitrogen assimilation is expected to  
616 be negligible due to constant nitrogen limitation in the surface waters<sup>29</sup>. Therefore,  $\delta^{15}\text{N}$  of the  
617 produced organic matter should reflect the composition of dominating nitrogen sources to the  
618 euphotic zone, namely, upward diffusive flux of deep-water nitrate and N<sub>2</sub> fixation both having  
619 distinct isotopic signals. In addition, in the northern gyre, atmospheric dust deposition can

620 significantly contribute to the total pool of new nitrogen<sup>31, 66</sup>. Using equation (1) we describe  
621 isotopic composition of nitrogen export as mixing between diffused nitrogen from the upper  
622 thermocline and nitrogen from external source, represented by either diazotrophy or  
623 atmospheric deposition at NOG, and diazotrophy only at SOG:

$$\delta^{15}\text{N}_{\text{PN}} = (f_1 \times \delta^{15}\text{N}_{f1}) + (f_2 \times \delta^{15}\text{N}_{f2}) \quad (1)$$

624 where  $f_1$  and  $f_2$  and  $\delta^{15}\text{N}$  denote fractions and isotopic signatures of dominant nitrogen sources.  
625 We estimate the percent contribution of these sources from a single choice of their respective  
626 endmember  $\delta^{15}\text{N}$  values: +2.73‰ (NOG) and +6.22‰ (SOG) for nitrate diffusing from the  
627 shallow thermocline across the concentration gradient; -1.0‰ for  $\text{N}_2$  fixation (both sites), -  
628 3.1‰ for bulk aerosol input (NOG only). Due to sensitivity of the two-endmember mixing  
629 model to the values of the chosen endmembers, we performed sensitivity analyses to account  
630 for uncertainty of the changing  $\delta^{15}\text{N}$  endmembers on the fraction of  $\delta^{15}\text{N}_{\text{PN}}$  (in %) originating  
631 from this source at each site, similar to isotopic assessment in ref<sup>28</sup>. The choices of  $\delta^{15}\text{N}$   
632 endmembers for each nitrogen source and those used in the sensitivity tests are described in  
633 the section below and the results are summarised in Supplementary Table 1.

#### 634 **Sensitivity analyses and $\delta^{15}\text{N}$ endmember choice.**

635 **Nitrate endmember:** The choice of nitrate  $\delta^{15}\text{N}$  endmember was based on the biogeochemical  
636 data (nitrate  $\delta^{15}\text{N}$ , nitrate and chlorophyll concentrations, PAR) obtained at the NOG and SOG  
637 sites during AMT cruises in May-June 2005 and October 2005, and US-GEOTRACES cruise  
638 GA03 in December 2011 (see ancillary biogeochemical datasets above).

639 At the permanently oligotrophic NOG and SOG sites, winter mixing is weak, and thermocline  
640 nitrate is supplied into the euphotic zone largely by turbulence-driven upward diffusion<sup>67</sup>. The  
641 magnitude of diffusive nitrate flux is governed by nitrate concentration gradients as the changes  
642 in turbulent diffusivity are relatively small<sup>67</sup>. At both sites, nitrate concentrations remain at

643 nanomolar levels ( $< 0.01 \mu\text{mol L}^{-1}$ ) throughout the top 130-150 m and increase below,  
644 signifying the position of the nitracline (defined by a nitrate concentration of  $0.1 \mu\text{mol L}^{-1}$  (e.g.  
645 ref<sup>68</sup>). The largest nitrate flux with a characteristic  $\delta^{15}\text{N}$  signature is therefore expected at the  
646 depth of the maximum nitrate concentration gradient typically found at depths near the base of  
647 the euphotic zone (0.1 % surface PAR; includes the deep chlorophyll maximum).

648 Referring to vertical profiles of nitrate and chlorophyll concentrations, we calculate  
649 concentration-weighted average nitrate  $\delta^{15}\text{N}$  (ref<sup>33</sup>) from the top of the nitracline, where nitrate  
650 concentrations begin to consistently increase, to the base of the euphotic zone. At NOG this  
651 yields nitrate  $\delta^{15}\text{N}$  of  $+2.73 \pm 0.36 \text{ ‰}$  (n=5) for the depth range of 137-191 m. In our isotopic  
652 budgets this value represents an isotopic signal of nitrogen pool influenced by  $\text{N}_2$  fixation and  
653 atmospheric deposition, and sustained over time in the shallow thermocline. This is achieved  
654 through both the internal cycle of low-  $\delta^{15}\text{N}$  nitrate assimilation and subsequent  
655 remineralisation and 2) accumulation of low-  $\delta^{15}\text{N}$  nitrate imported during the northward water  
656 mass transit<sup>7, 30</sup>. This nitrogen pool has not yet been homogenised with the large global ocean  
657 nitrate reservoir ( $\sim 4.8 \text{ ‰}$ ) or  $^{15}\text{N}$ -enriched through denitrification<sup>29, 30</sup>. At SOG the most  
658 relevant depth range for measured nitrate  $\delta^{15}\text{N}$  spanned 226-230 m, substantially deeper than  
659 the base of the euphotic zone. The corresponding mean  $\delta^{15}\text{N}$  of  $+6.22 \pm 0.35 \text{ ‰}$  may thus  
660 overestimate the value for the shallower waters, where preferential remineralisation of  $^{14}\text{N}$  may  
661 introduce a  $^{15}\text{N}$ -depleted signal to the nitrogen pool<sup>29</sup>.

662 For the primary sensitivity test (Supplementary Table 1) we used the minimal nitrate  $\delta^{15}\text{N}$   
663 observed in the upper thermocline at NOG ( $+0.96 \text{ ‰}$  at 137 m depth) to estimate the least  
664 contribution of local  $\text{N}_2$  fixation to  $\delta^{15}\text{N}_{\text{PN}}$ . We also tested nitrate  $\delta^{15}\text{N}$  averaged from the top  
665 of nitracline down to  $26.8 \text{ kg m}^{-3}$  isopycnal surface, which marks the main thermocline depth  
666 at the study sites<sup>7, 69</sup>. The corresponding value at NOG was  $+3.53 \pm 0.40 \text{ ‰}$  (n=13) for the 136-

667 421 m depth range; the SOG value was  $+6.35\pm 0.32$  ‰ (n=3) for 226-306 m depth range.  
668 Finally, we included the oceanic global mean  $\delta^{15}\text{N}$  ( $+4.8$ ‰; ref<sup>29</sup>) to compare our isotope  
669 budgets with published data.

670 **Nitrogen fixation endmember:** We chose the mean  $\delta^{15}\text{N}$  for diazotrophic biomass ( $-1\pm 1$ ‰)  
671 to represent the  $\text{N}_2$  fixation endmember at both sites <sup>29, 70, 71</sup>. Assuming the mean nitrate  $\delta^{15}\text{N}$   
672 signal in the shallow thermocline, the range of the isotopic signal for diazotrophic nitrogen ( $-$   
673  $2$ ‰ to  $0$ ‰) generates average contributions of 39.8-68.5% and 26.1-34.4% to  $\delta^{15}\text{N}_{\text{PN}}$  at NOG  
674 and SOG, respectively (Supplementary Table 1).

675 **Atmospheric deposition endmember:** Atmospheric fluxes supply approximately  $\sim 9.9\times 10^9$   
676 mol N yr<sup>-1</sup> to the central North Atlantic gyre and  $5.8\times 10^9$  mol N yr<sup>-1</sup> to the South Atlantic  
677 gyre<sup>72</sup>. Although these values are notably smaller than regional estimates of new nitrogen  
678 inputs from diazotrophy ( $20\times 10^{11}$  mol N yr<sup>-1</sup>; ref<sup>73</sup>), recent studies<sup>30, 31, 35</sup> suggest that  
679 deposition fluxes can significantly lower the  $\delta^{15}\text{N}$  of the nitrogen pool. The published data on  
680  $\delta^{15}\text{N}$  in atmospheric fluxes in the open Atlantic Ocean is extremely scarce. Previous studies<sup>31,</sup>  
681 <sup>35, 74, 75, 76</sup> report a wide range of  $\delta^{15}\text{N}$  in bulk aerosol and rainfall samples ( $-6.8$ ‰ to  $+1.7$ ‰).  
682 Given that dry deposition dominates atmospheric input at NOG, a value close to an average  
683 isotopic signal of bulk aerosols seems the most appropriate to represent the dust endmember at  
684 NOG. We thus choose  $\delta^{15}\text{N}$  of  $-3.1$ ‰, based on the mean  $\delta^{15}\text{N}$  values measured in the Sahara-  
685 influenced aerosol samples collected the sub-tropical North Atlantic<sup>31, 74</sup> and Crete<sup>76</sup>. For the  
686 sensitivity test, we varied  $\delta^{15}\text{N}$  of aerosol N across the full range, also including annual ( $-4.5$ ‰)  
687 and seasonal cold ( $-6.8$ ‰; October-March) and warm ( $-1.9$ ‰; April-September) averages  
688 measured in the Bermuda rainfall<sup>35, 74, 75</sup> (Supplementary Table 1). We find that with the nitrate  
689  $\delta^{15}\text{N}$  of  $2.73$ ‰, aerosol nitrogen can account for a sizable fraction of  $\delta^{15}\text{N}$  of nitrogen export  
690 at NOG (21.7-176%). Therefore, with nitrogen input equal or greater to magnitude of  $\text{N}_2$   
691 fixation, dust deposition can have a similar or greater effect on the isotopic budget of trap

692 material from NOG. Hence, future studies should include the measurements of both magnitude  
693 and  $\delta^{15}\text{N}$  of dust deposition and  $\text{N}_2$  fixation to avoid under- or over-estimation of the  
694 importance of each of source.

695 **Assessment of ballast effect of lithogenic flux.** We examined the relationship between POC  
696 and (bio)mineral at NOG and SOG using POC flux model by ref<sup>2</sup>. The model divides POC  
697 flux into fractions ballasted by biomineral (opal + calcite;  $\text{POC}_{\text{bio}}$ ) and lithogenic ( $\text{POC}_{\text{lith}}$ )  
698 particles, and freely sinking POC ( $\text{POC}_{\text{free}}$ ). We use multiple linear regression to fit the particle  
699 flux data into equation (2) and determine correlation coefficients  $a$ ,  $b$  and  $c$  (hereafter, carrying  
700 coefficients) for each fraction, following approach in refs<sup>2, 42, 77</sup>.

$$\text{POC flux} = a \times \text{POC}_{\text{bio}} + b \times \text{POC}_{\text{lith}} + c \times \text{POC}_{\text{free}} \quad (2)$$

701 Carrying coefficients only reflect the size of the ballast-normalized fraction of POC flux, but  
702 not their absolute magnitudes, and are used to calculate the relative fraction (in %) of POC  
703 associated with each ballast type<sup>42, 77</sup>. We further assume that the  $\text{POC}_{\text{free}}$  fraction is negligible  
704 at 3000 m depth and force multiple linear regression to pass through zero<sup>2</sup>. The strong temporal  
705 variability of dust deposition limits the relevance of the annual-scale approach for estimating  
706 the role of lithogenic ballast to POC flux. Hence, we first assess the effect of lithogenic ballast  
707 based on different POC sequestration scenarios, namely, (1) elevated POC flux ( $\geq 120\%$  of  
708 annual mean) at NOG, (2) POC flux at NOG outside scenario (1), (3) POC flux at SOG. We  
709 evaluated the sensitivity of these results by performing multiple linear regression on the NOG  
710 flux dataset separated according to the high and low lithogenic fluxes (Supplementary Table  
711 2). Our approach differs from that applied previously by refs<sup>2, 42, 77</sup> in which carrying  
712 coefficients for both calcite and opal were determined. This is due to strong collinearity  
713 observed between calcite and opal in all POC-based groups, violating the independence  
714 assumption of multiple linear regression, as further determined by ridge regression analysis.  
715 The resulting carrying coefficients and calculated proportion of ballast-associated POC flux in



716 each surveyed group are summarised in Supplementary Table 2. The carrying coefficients for  
717 lithogenic material compared well with the global and the north Atlantic means (0.052, and  
718 0.058, respectively)<sup>2</sup> during low POC flux, but exceeded these values during high fluxes and  
719 overall at SOG. We acknowledge that the relatively large carrying coefficient for lithogenic  
720 ballast in the SOG group compared to the NOG groups and other time-series might be an  
721 overestimation introduced by a nearly 1:1 ratio between POC and lithogenic fluxes and a their  
722 strong positive correlation (Spearman's  $p = 0.91$ ). As a result, the proportion of POC flux  
723 ballasted by lithogenic material appears to be comparable between SOG and scenario (2) at  
724 NOG, despite the significant difference in their lithogenic fluxes (Supplementary Table 2).

725 **Data availability:** The data analysed during this study are available from the corresponding  
726 author upon reasonable request. The supporting data for this study are available from the  
727 repository of the British Oceanographic Data Centre upon request.

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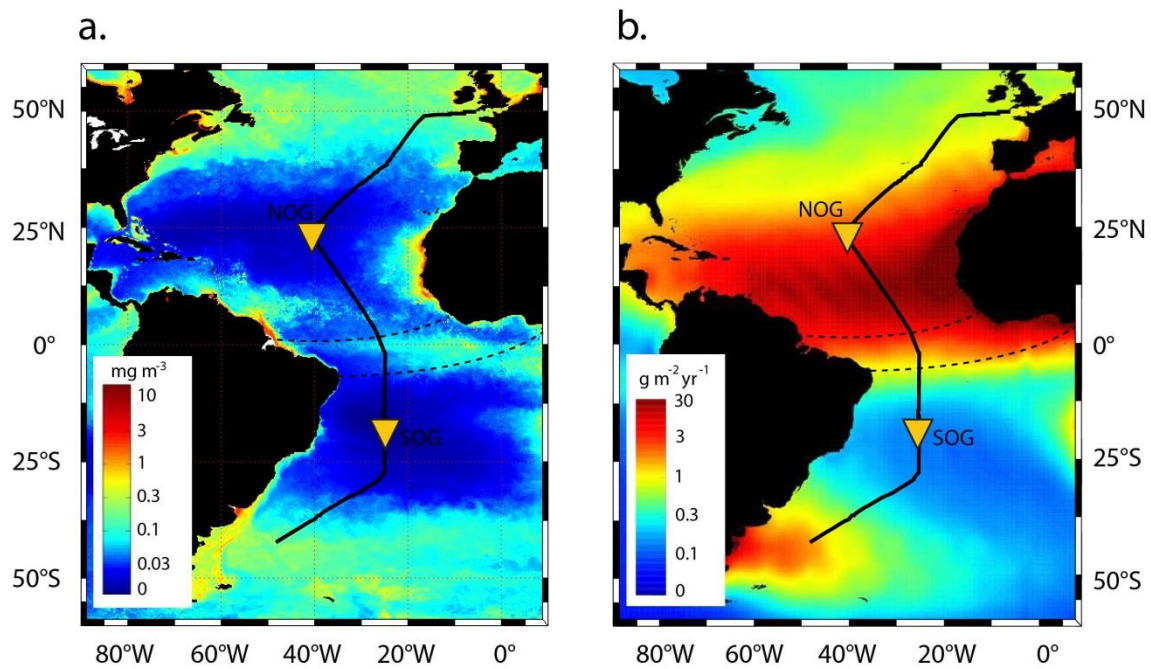
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841 **FIGURES**

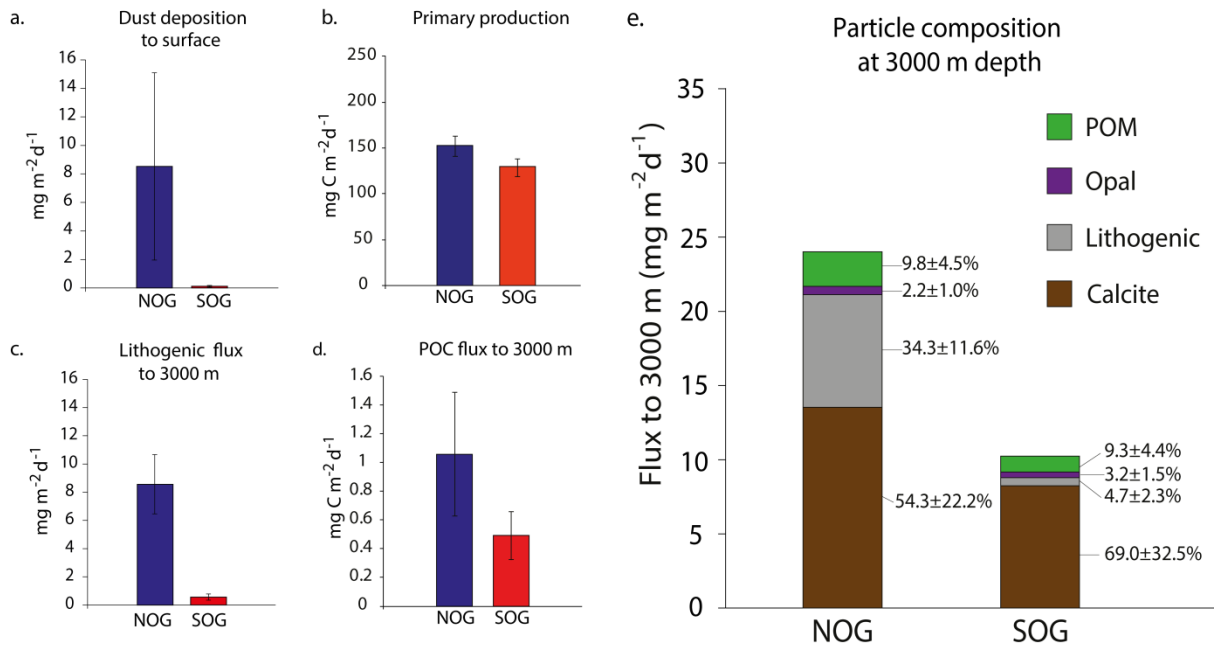
842 **Figure 1**



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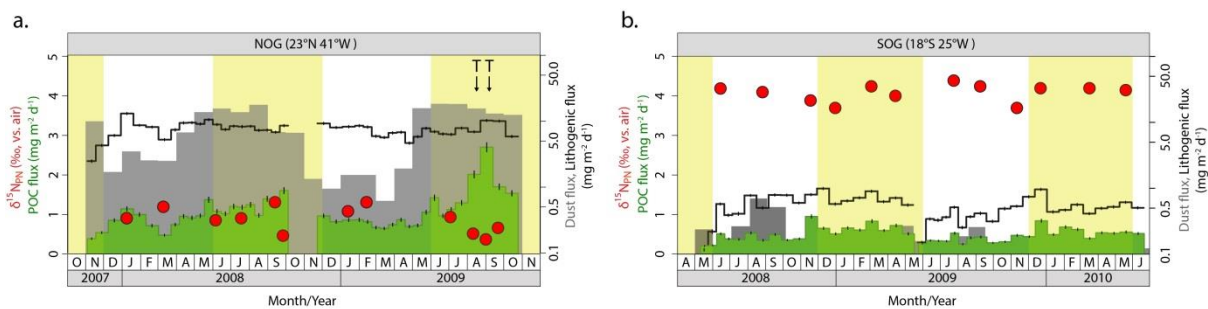
845 **Figure 2**



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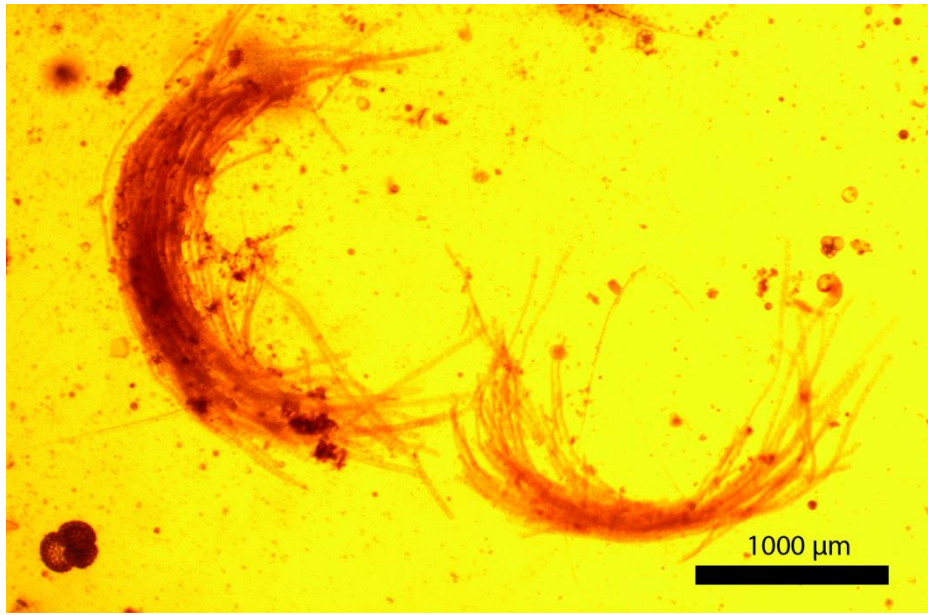




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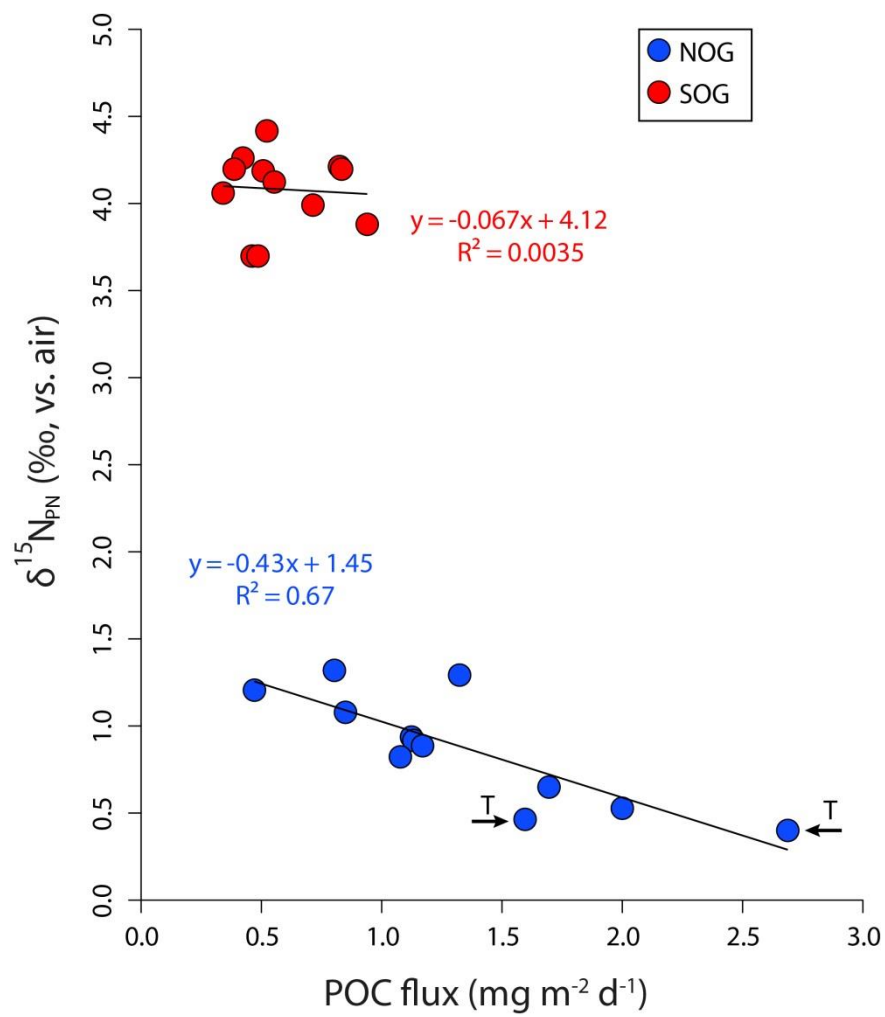
851 **Figure 4**



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854 **Figure 5**



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## 858 SUPPLEMENTARY INFORMATION

859 **Supplementary Table 1| Sensitivity analysis of the choice endmember  $\delta^{15}\text{N}$  value on**  
860 **calculated source contribution (%) to isotopic signal of nitrogen flux at 3000 m depth**

Nitrate endmember $\delta^{15}\text{N}$ (‰)	$\text{N}_2$ fixation (%)			Dry deposition (%)			Rainfall (%)		
	Min	Choice	Max	Min	Choice	Max	Min	Choice	Max
<i>NOG</i>	-2	-1	0	-6	-3.1	+1.7	-6.8	-4.5	-1.9
+0.96 (min <i>NOG</i> )	2.82	4.25	8.68	1.20	2.1	-11.3	1.07	1.53	2.91
<b>+2.73 (137- 191 m)</b>	<b>39.8</b>	<b>50.4</b>	<b>68.5</b>	<b>21.7</b>	<b>32.4</b>	<b>176.2</b>	<b>19.8</b>	<b>26.2</b>	<b>40.7</b>
+3.53 (137-421 m)	48.0	58.6	75.2	27.4	40.0	145.0	25.7	33.0	48.9
+4.8*	57.7	67.7	81.7	36.3	49.7	126.6	33.8	42.2	58.6
<i>SOG</i>	-2	-1	0						
<b>+6.22 (226-230 m)</b>	<b>26.1</b>	<b>29.7</b>	<b>34.4</b>						
+6.35 (226-306 m)	27.2	30.9	35.8						
+4.8*	10.6	12.5	15.1						

The source and choice of  $\delta^{15}\text{N}$  endmember values are described in Methods. Depth-range over which measured nitrate  $\delta^{15}\text{N}$  values were averaged (nitrate concentration weighted) is given in parentheses. Bold values show the percentage contribution values calculated with the preferred endmember  $\delta^{15}\text{N}$  values.

\*global average  $\delta^{15}\text{N}$  of deep-water nitrate (ref<sup>29</sup>)

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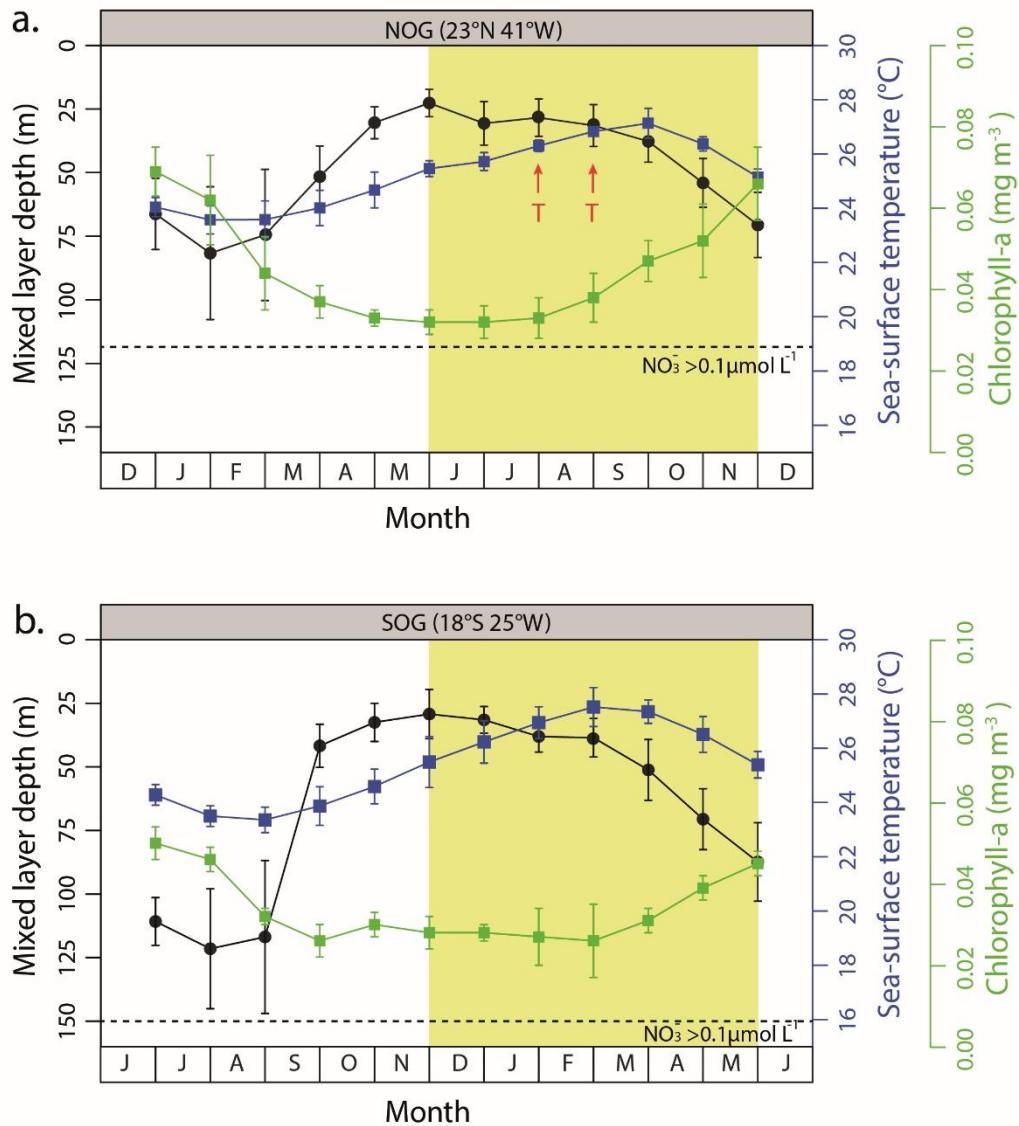
863 **Supplementary Table 2| Results of multiple linear regression model**

Group	Mean flux (mg m <sup>-2</sup> d <sup>-1</sup> )			Carrying coefficient*		Ballasted POC flux (%)		R <sup>2</sup> ‡
	POC	Opal + Calcite	Lithogenic	Opal + Calcite	Lithogenic	Opal + Calcite	Lithogenic	
<i>NOG groups</i>								
High POC (10)	1.64±0.43	17.4±6.7	9.14±1.9	0.027 <sup>ns</sup>	0.126	29.1	70.1	0.94
Low POC (31)	0.88±0.21	13.0±3.5	8.38±2.2	0.036	0.048	53.7	45.7	0.98
<i>SOG group (40)</i>								
High Lith (9)	1.31±0.58	15.6±1.81	11.3±1.44	0.067 <sup>ns</sup>	0.025 <sup>ns</sup>	79.7	21.2	0.87
Low Lith (32)	0.99±0.36	13.7±5.27	7.80±1.62	0.045	0.048	62.3	37.6	0.95
<i>SOG group (40)</i>	0.49±0.17	8.62±2.96	0.54±0.21	0.032	0.371	57.2	41.0	0.96

Values in parentheses indicate number of data points (=collection cups) pulled into each group. Statistically not significant carrying coefficients (p>0.01) are marked with 'ns'.

\* Correlation coefficients of multiple linear regression determined from equation (2) (see Methods).

‡ Overall model fit



865

866 **Supplementary Figure 1| Annual cycle of sea-surface temperature, mixed layer depth and**

867 **surface chlorophyll-a at NOG (a) and SOG (b).** Monthly-averaged sea-surface temperature

868 and chlorophyll-a concentrations are recorded by the MODIS-A satellite at 9km resolution

869 during 2007-2010. Error bars show one standard deviation of the temporal mean. The mixed

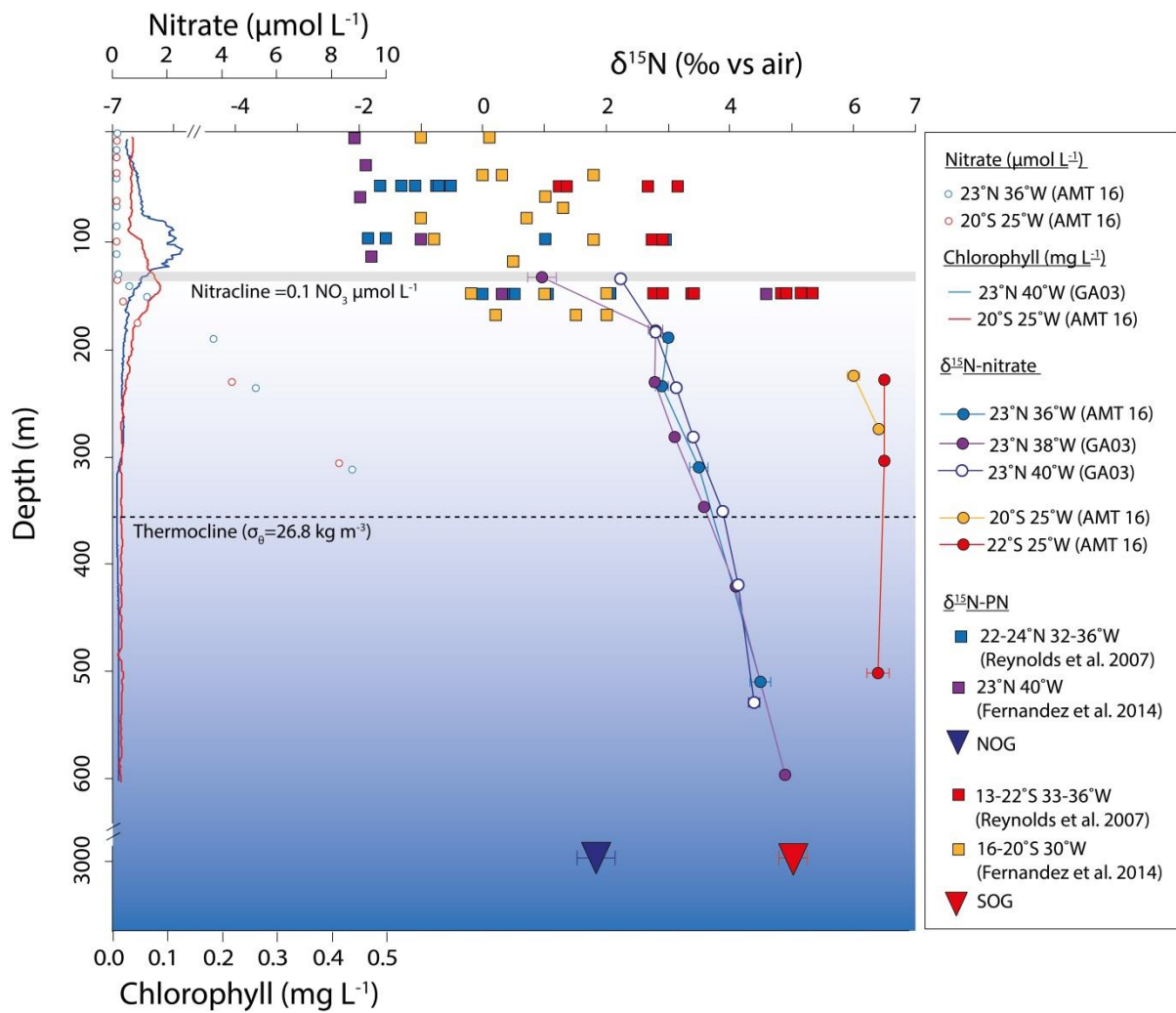
870 layer depth is derived from the ARGO-based climatology. Error bars show one standard

871 deviation of the areal mean. In **a** and **b**, dashed line indicates the approximate depth of

872 nitricline ( $\text{NO}_3^- > 0.1 \mu\text{mol L}^{-1}$ ), based on *in situ* nitrate measurements during AMT cruises 18-

873 21 in October-November 2008-2010 (see Methods for data sources). In **a**, red letter “T” point

874 to the months, when *Trichodesmium* “tufts” were recovered in the NOG traps.



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877 **Supplementary Figure 2| Nitrogen isotope data for water column total nitrate, suspended**  
 878 **particles and trap material near the study.** Concentrations of nitrate and chlorophyll  
 879 measured at the study sites are also shown. Data sources are described in the figure legend with  
 880 complete references provided in the Methods section.

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