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Observation-Based Estimates of Eulerian-Mean Boundary Downwelling in the Western Subpolar North Atlantic

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1 **Observation-based estimates of Eulerian-mean boundary downwelling in the western**
2 **subpolar North Atlantic**
3

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9 **Key Points:**

- 10 • The long-term mean full-depth density field of the subpolar North Atlantic's boundary is
11 reconstructed from hydrography data
- 12 • The along-boundary densification results in a 2.12 ± 0.43 Sv Eulerian-mean downwelling
13 between Denmark Strait and Flemish Cap
- 14 • A first observation-based regional and seasonal distribution of near-boundary Eulerian-
15 mean downwelling is provided
16

17 **Abstract**

18 A significant fraction of the Eulerian-mean downwelling feeding the lower limb of the Atlantic
19 Meridional Overturning Circulation (AMOC) occurs along the subpolar North Atlantic
20 continental slopes and is maintained by along-boundary densification and large-scale geostrophic
21 balance. We here use Argo and shipboard hydrography data to map the 2002-2015 long-term
22 mean density field along the boundary via a dedicated optimal interpolation tool. The overall
23 downstream densification implies an Eulerian-mean downwelling of 2.12 ± 0.43 Sv at 1100 m
24 depth between Denmark Strait and Flemish Cap. A clear regional pattern emerges with
25 downwelling in the Irminger Sea and western Labrador Sea and upwelling along Greenland
26 western continental slope. Comparisons with independent cross-basin estimates confirm that
27 vertical overturning transport across the marginal seas of the subpolar North Atlantic mainly
28 occurs along the continental slopes, and suggest the usefulness of hydrographic data in providing
29 quantitative information about the sinking branch of the AMOC.

30 **Plain Language Summary**

31 The Atlantic Meridional Overturning Circulation (AMOC), a critical component of the Earth's
32 climate system due to its role in redistributing heat and freshwater between low and high
33 latitudes, is anticipated to decline over the next century. The downwelling of surface waters in
34 the subpolar North Atlantic that feeds the lower limb of AMOC is a vital yet vulnerable process.
35 As revealed by previous theoretical and modelling work, the overall downstream densification
36 along the boundary results in a significant boundary downwelling. Here, the density along the
37 western boundary between Denmark Strait and Flemish Cap is reconstructed to provide a first
38 observation-based description of the regional and seasonal distribution of this boundary-focused
39 downwelling in the subpolar North Atlantic. This study not only provides valuable insights into
40 how to improve existing ocean circulation theories of overturning but also contributes to a solid
41 benchmark for evaluating how climate models simulate the sinking branch of the AMOC.

42 **1 Introduction**

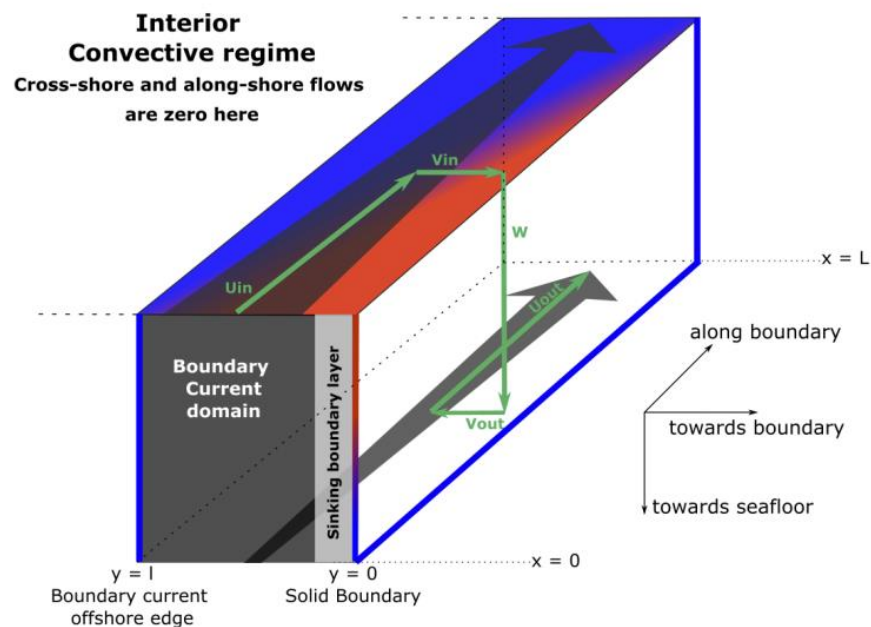
43 Through its role in redistributing heat, freshwater, and chemical properties between low and high
44 latitudes, the Atlantic Meridional Overturning Circulation (AMOC) is a critical component of
45 Earth's climate system. Warm and saline waters flow from low latitudes towards the poles within
46 the AMOC upper limb. Following significant mixing-driven and surface-forced water mass
47 transformation, colder and fresher waters are returned southward as intermediate and deep-water
48 masses within the AMOC lower limb. Climate model studies indicate that the AMOC could
49 weaken significantly over the next century (IPCC, 2021), and significant efforts have been
50 consequently made over the last decades to understand the processes maintaining the AMOC and
51 the drivers of its variability (e.g., Lozier et al., 2012; Buckley and Marshall, 2016; Johnson et al.,
52 2019).

53 The vertical connection between the upper and lower limb of the AMOC, as well as the
54 underlying mechanism, are still under investigation (e.g., Spall & Pickart, 2001; Pedlosky, 2003;
55 Straneo, 2006; Spall, 2010; Brüggemann & Katsman, 2019). The location of this downwelling
56 has long been associated with regions of intense open-ocean convection, such as the Labrador
57 and Irminger seas, where dense and deep waters form because of convective mixing. However,
58 deep convection regions have large vertical heat and salt transports (in density space) but a
59 negligible vertical mass transport (in depth space) (Marshall and Schott 1999; Send and Marshall

60 1995; Spall 2003, 2004). Instead, the Eulerian-mean downwelling connecting the AMOC upper
 61 and lower limbs should occur near continental boundaries, where geostrophy breaks down and a
 62 new balance between vertical stretching of planetary vorticity and dissipation of relative vorticity
 63 in a thin boundary layer emerges (Spall, 2010). Previous research has established that such
 64 downwelling through modification of the boundary current properties in a region is undergoing
 65 buoyancy loss (e.g., Spall & Pickart, 2001; Straneo, 2006; Spall, 2010; Cenedese, 2012; Katsman
 66 et al., 2018). Notably, the dynamics of Eulerian-mean downwelling are primarily governed by
 67 the large-scale geostrophic flow, so that the details of the thin boundary layer where vorticity is
 68 ultimately dissipated do not need to be resolved for capturing the overall impact of downwelling
 69 (Spall, 2008).

70 Several studies have focused on the location and underlying mechanism of boundary
 71 downwelling. Spall and Pickart (2001) investigated the boundary sinking using a thermodynamic
 72 balance. During winter, the boundary current loses heat and buoyancy. As shown schematically
 73 in Figure 1, the resulting along-boundary density increase requires a vertically sheared cross-
 74 shore baroclinic flow in thermal wind balance (V_{in}) fed by the boundary current itself (U_{in} ; there
 75 is no exchange with the basin interior). Therefore, the flow rotates counterclockwise with depth
 76 along the boundary within a so-called "cooling spiral". To maintain local mass conservation, a
 77 net downwelling (W) is required to balance the mass flux towards the boundary. This causes a
 78 barotropization of the boundary current, with the upper part decelerating and the lower part
 79 accelerating (Spall 2004, 2008; Straneo 2006). Without any mean mass flux between the
 80 boundary region and the basin interior, the sinking water joins the lower part of the boundary
 81 current (V_{out}) and is rapidly exported within the lower limb of AMOC (U_{out}). Thus, the along-
 82 boundary pressure (or density) gradient and associated cross-shore geostrophic flow are vital for
 83 sinking to occur.

84



85

86 **Figure 1.** The cooling spiral and associated boundary downwelling. Gray arrows represent the
87 background boundary current. The blue-red shading depicts the horizontal density fields, with
88 the distribution of light waters in red and dense waters in blue giving rise to a cross-shore
89 baroclinic flow in thermal wind balance (V_{in}) fed by the boundary current (U_{in}). Local mass
90 conservation lead to a downward flow at the wall (W), a deep entrainment in the boundary
91 current (V_{out}), and a rapid along-boundary export (U_{out}). The overall effect is a net downwelling
92 within and a barotropization of the boundary current.

93 Several processes, including surface buoyancy flux or mean and eddy-driven heat advection,
94 likely contribute to maintaining an along-boundary density gradient. While the respective
95 contribution of these processes is regionally unknown, lateral eddy-driven heat exchanges –
96 generated by the lateral density gradients and baroclinic instability – are widely acknowledged as
97 important in densifying the boundary region, as seen for instance in the Labrador Sea (Lilly et
98 al., 2003; Katsman et al., 2004). Lateral eddy-induced heat fluxes are required to balance the heat
99 loss to the atmosphere and restratify open-ocean water columns following deep convection
100 events (Katsman et al., 2004; Hátún et al., 2007; Chanut et al., 2008; Kawasaki and Hasumi,
101 2014; Tagklis et al., 2020). Additionally, eddies contribute to interior downwelling and boundary
102 current barotropization through along-isopycnal water masses stirring and exchange between the
103 boundary current and the interior (Khatiwala and Visbeck, 2000; Brüggemann and Katsman,
104 2019).

105 Recent studies have used realistic and high-resolution model simulations to investigate the net
106 downwelling in the entire Subpolar Gyre (SPG) (Katsman et al. 2018, Sayol et al. 2019) or in
107 marginal seas characterized by high convective activity, such as the Labrador Sea (Brüggemann
108 and Katsman 2019; Georgiou et al., 2019). Katsman et al. (2018) and Sayol et al. (2019)
109 confirmed that the bulk of the vertical volume flux occurs along the boundary and that its spatial
110 integral in the SPG is close to the magnitude of the AMOC – the zonally-integrated meridional
111 flow – at the southern exit of the SPG (i.e., near 45°N). Katsman et al. (2018) further
112 demonstrated that the amount of boundary sinking is largely determined by density change,
113 except the region where eddy-driven processes or hydraulic controls may become significant.
114 The role of eddy-driven processes was particularly studied by Brüggemann and Katsman (2019)
115 and Georgiou et al. (2019), who showed the importance of eddies in balancing heat loss over the
116 Labrador Sea and in triggering boundary downwelling along steepened isopycnal surfaces.

117 Estimates of boundary-focused sinking and its associated mechanisms from *in situ* observations
118 are still lacking. Here, we use hydrography data from global ocean monitoring programs (e.g.,
119 Argo, Go-Ship) to estimate for the first time the long-term Eulerian-mean along-boundary
120 downwelling and its spatial and seasonal distribution within the western SPG, from Denmark
121 Strait (DKS) to Flemish Cap (FC). While such observations cannot resolve the intricate and
122 small-scale dynamics of downwelling within the thin boundary layer where it is most likely to
123 occur, they can be used to infer the large-scale geostrophic balance governing it.

124 We begin by introducing a specific optimal interpolation method to map temperature and salinity
125 along the boundary, and then describe the method used for calculating cross-shore velocities and
126 resulting vertical transport (section 2). Section 3 describes the reconstruction of the along-
127 boundary density field, as well as the subsequent calculation of geostrophic velocity and
128 associated downwelling. A summary and a discussion conclude this study (section 4).

129 2 Data and Methodology

130 2.1 Data

131 We use three datasets of temperature and salinity profiles to map the long-term mean (2002–
132 2015) and seasonal cycle of hydrographic properties in the SPG (52–66°N; 30°W–66°W): the In
133 Situ Analysis System-ISAS15 (Kolodziejczyk et al., 2021; Gaillard et al., 2016), the Coriolis
134 data set for ReAnalysis-CORA (Szekely et al., 2019; Cabanes et al., 2013), and EN4.4.2.2 (Good
135 et al., 2013). Whereas ISAS15 only contains Argo profiles, EN4 and CORA include profiles
136 from fixed moorings and shipboard full-depth CTD profiles. These profiles have undergone
137 quality control checks in delayed mode, and we only use *good* profiles (i.e., fully acceptable).
138 Because most of Argo floats drift at 1000 m depth, ISAS15 exhibits an uneven distribution
139 across the domain with decreasing near the shelf (see Figure S1). The EN4 and CORA provide
140 adequate coverage of profiles near the shelf. The temperature and salinity fields are then used to
141 derive the in situ (ρ) and potential (σ_0) density fields. Bathymetry is derived from ETOPO2.

142 2.2 Optimal Interpolation

143 Optimal interpolation (OI) is a frequently used technique for generating gridded property fields
144 from sparse and irregular data (Bretherton et al., 1976, Kaplan et al., 1997). The conventional OI
145 algorithms for climatological mapping use circle-shaped Gaussian correlation functions (see
146 Gaillard et al., 2016 for additional details), but we herein employ an elliptic Gaussian correlation
147 function parallel to isobaths, with length scales of 67 km (cross-shore axis) and 145 km (along-
148 shore axis) to account for the larger (smaller) correlation length scales in the along-shore (cross-
149 shore) directions. In other words, this provides data along the boundary current's primary path
150 larger weights in the estimation. Based on convergence sensitivity tests, the horizontal resolution
151 is set as 4km, which enables us to refine the properties along boundaries and topographic
152 features (such as the sharpest portions of the continental slope). Interpolation is conducted at 152
153 levels independently between 0 and 2000 m; the vertical spacing is 5 m down to 100 m, 10 m
154 down to 800 m, and 20 m below. The configuration of the OI is explained in more detail in
155 Gaillard et al., (2016). The overall mapping methodology was eventually validated by good
156 performance in estimating properties along the OVIDE and AR7W hydrographic lines (see
157 Figure S2-S5). To estimate the vertical transport, the final multiproduct mappings of temperature
158 and salinity for the four seasons (Winter (Jan-Mar), Spring (Apr-Jun), Summer (Jul-Sep),
159 Autumn (Oct-Dec)) is the average of the fields derived independently from the three datasets.
160 The density below 2000 m is obtained by linear extrapolation from the above 2000 m. The three-
161 dimensional density field is eventually interpolated onto the locations of 100m-spaced isobaths
162 spanning 700 m to 3000m from DKS to FC.

163 2.3 Computing cross-shore geostrophic velocity and the vertical transport

164 To calculate the absolute geostrophic cross-shore velocity, a local mass balance assumes equal
165 inshore and offshore transport. Any horizontal recirculations between the boundary current and
166 the interior are assumed to be mostly barotropic, with no effect on the cross-shore baroclinic
167 density field. These assumptions are supported by laboratory experiments showing that water
168 downwelling along a vertical wall returns offshore in the opposite direction (Cenedese, 2012),
169 and by observations of the strong barotropic nature of inner gyres and lateral entrainments (Våge
170 et al., 2011). The baroclinic component of the cross-shore velocity relative to the sea surface

171 ($v_{\text{baroclinic}}$) is firstly derived from the along-boundary density gradient and the thermal wind
 172 balance:

173

$$\frac{\partial v_{\text{baroclinic}}(x, y, z)}{\partial z} = -\frac{g}{\rho_0 f} \frac{\partial \rho(x, y, z)}{\partial x} \quad (1)$$

174

175 where x , y , and z represent the along-shore, cross-shore, and vertical directions, respectively. ρ
 176 is the *in-situ* density field derived from OI, ρ_0 is 1024 kg/m^3 , and f is $1.26 \times 10^{-4} \text{ s}^{-1}$.

177 The reference velocity $v_{\text{reference}}$ is then obtained by applying the local *zero-mass transport*
 178 *constraint*:

$$v_{\text{reference}}(x, y) = \frac{-\int_{H(x,y)}^0 (v_{\text{baroclinic}}(x, y, z)) dz}{\Delta z} \quad (2)$$

179

180 where $H(x, y)$ denotes the bottom depth and Δz the water depth.

181

182 We calculate vertical transport along the SPG at isobaths ranging from 700 to 3000 m with a
 183 100-meter interval. The analysis focuses offshore of the 700 m isobath because of poorer
 184 sampling inshore. This has little effect on the maximum transport estimate, however, which is
 185 found across much deeper slope. For a given isobath (y_0), the vertical transport streamfunction
 186 $\psi(z)$ is estimated as follows:

$$\psi(y_0, z) = \int_{x_u}^{x_d} \int_z^0 (v_{\text{baroclinic}}(x, y_0, z') + v_{\text{reference}}(x, y_0)) dz' dx \quad (3)$$

187

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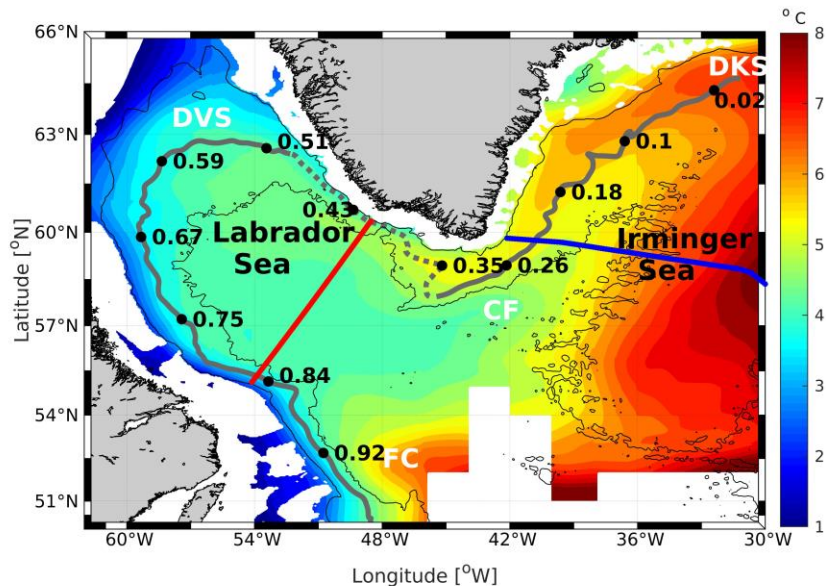
189 where x_u and x_d are the upstream and downstream locations along a given isobath y_0 . The
 190 maximum of $\psi(y_0, z)$ represents the net shoreward transport integrated along an isobath between
 191 x_u and x_d . The maximum vertical transport is thus defined as the maximum of the net shoreward
 192 transport ψ_{max} across the 100-m spaced isobaths between 700 m and 3000 m.

193

194 A Monte Carlo approach is used to add an uncertainty estimate to our mean vertical transport
 195 calculation. We use the errors estimated from the OI calculation, which depend on the
 196 distribution density and variability of adjacent profiles, to perturb the temperature and salinity
 197 fields randomly and re-compute overturning streamfunctions. 12000 bootstrap estimates of the
 198 streamfunctions in four seasons are calculated using 3000 iterations on the three datasets. The
 199 errors of the transport are considered as two times the standard deviation of the 12000 estimates
 200 (see Figure S6).

201 **3 Results**

202 Figure 2 shows the multiproduct long-term mean temperature field averaged over the upper 300
 203 m depth. It depicts the cooling of the warm boundary current in the upper layer as it flows from
 204 DKS to FC. The full-depth surface-referenced in situ (ρ) and potential (σ_0) fields are determined
 205 along 100 m-spaced isobaths ranging from 700 to 3000m. The 2400m isobath is shown here (for
 206 the reason that will become obvious later) and labeled using the distance relative to the isobath's
 207 total length (4032 km), starting at $s=0$ near the DKS at -30°W and ending at $s=1$ near FC at $-$
 208 48°W .

209
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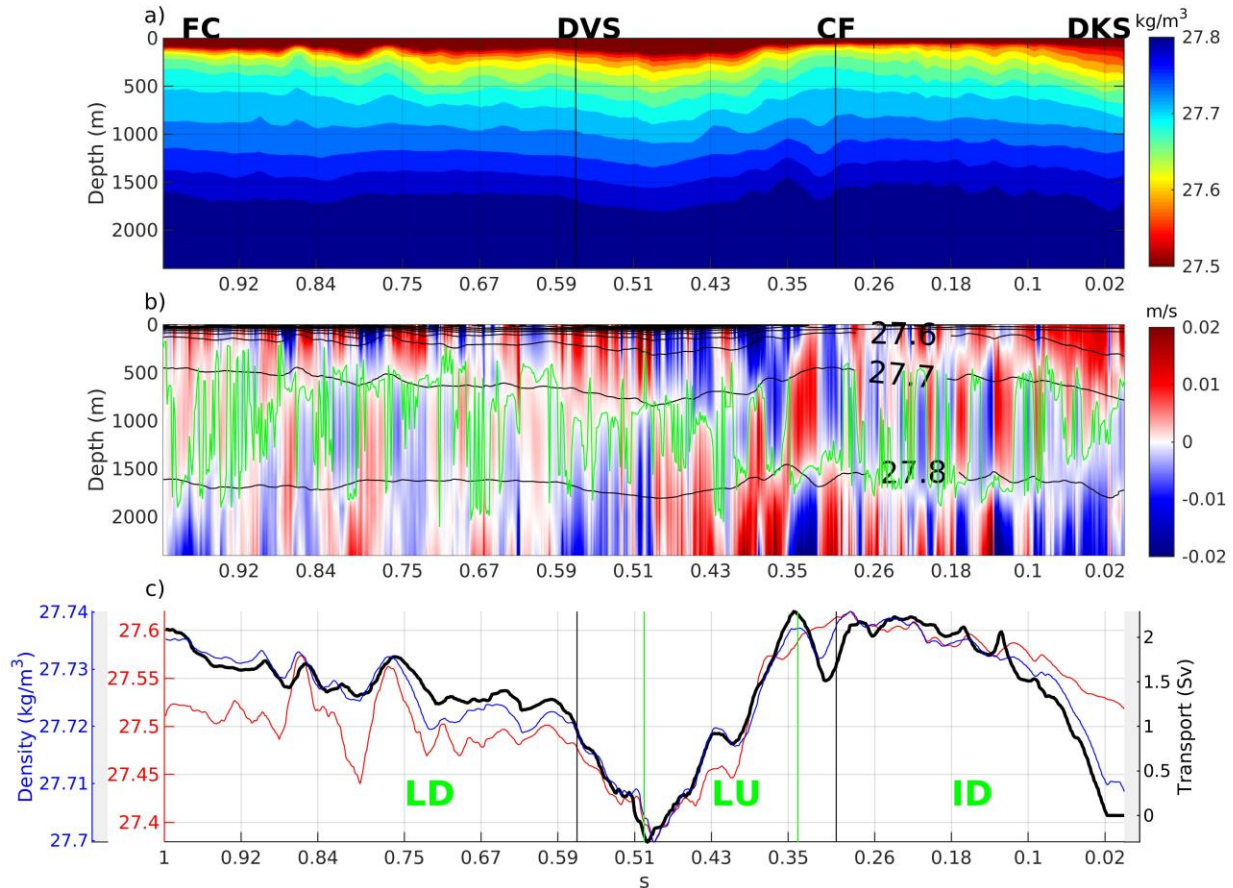
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212 **Figure 2.** Observed multiproduct time-mean (2002-2015) temperature field averaged in upper
 213 300 m as obtained from an OI tool specifically designed for the boundary region (see section
 214 2.2). The isobaths 700 m and 3000 m are shown as thin black contours, and the isobath 2400 m
 215 (thick black line) used in Figure 3 to show along-boundary density and velocity fields, from $s = 0$
 216 at Denmark Strait until $s = 1$ at Flemish Cap. Solid and dashed portions of this contour refer to
 217 (Eulerian) downwelling and upwelling regions. Key locations noted in the plot are Denmark
 218 Strait (DKS), Cape Farewell (CF), Davis Strait (DVS), and Flemish Cape (FC). Red line
 219 indicates the AR7W section, whereas the blue line indicates the OVIDE section.

220

221 Figure 3ab depicts the density and associated cross-shore geostrophic velocity sections following
 222 isobath 2400 m from DKS to FC (the along-isobath baroclinic and reference velocities in
 223 Equation 2 are shown in Figure S8). The density generally increases downstream (Figure 3a, c)
 224 and is primarily determined by the change in along-shore temperature (see Figure S7). The
 225 density of the upper (above 300m) and intermediate layer (300-1500 m) gradually increases from
 226 $s=0$ to $s=0.3$, while it decreases from $s = 0.3$ until the east of Davis Strait (DVS) ($s=0.5$)
 227 (Figure 3c). Downstream of DVS, the density in the intermediate layer increases again but
 228 remains constant in the upper layer. Overall, even though the density in the upper layer remains
 229 constant, the increase in density in the intermediate layer along isobath 2400 m from 27.71 kg m^{-3}
 230 to 27.74 kg m^{-3} should be expected to drive an overall downwelling. The resulting cross-shore
 231 absolute geostrophic velocity (Figure 3b) shows the water flows inshore (offshore) and offshore
 232 (inshore) in the upper and lower layers, driving downwelling (upwelling). There are two levels of
 233 no motion at some boundary locations due to more complex changes of stratification (induced by
 234 the freshwater inflow near DVS, for instance). The level of no motion is deeper in the Irminger
 235 Sea than in the Labrador Sea and ranges from 500 to 1500m depth.

236



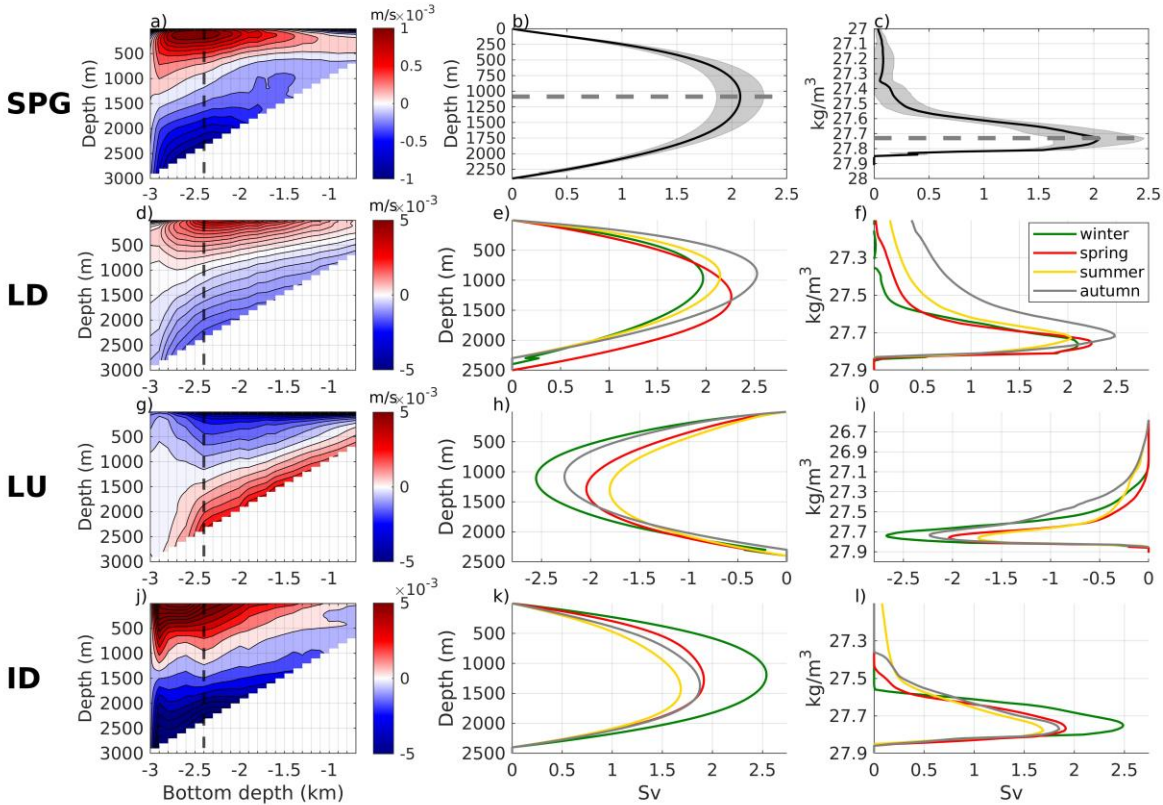
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238 **Figure 3.** (a) OI-derived multiproduct density (σ_0) field along isobath 2400 m. (b) The cross-
 239 shore geostrophic velocity field along isobath 2400 m. The positive (negative) values indicate
 240 inshore (offshore) flows. The green line indicates the depth at which the maximum (accumulated)
 241 vertical transport is found (i.e., levels of no motion). The $\sigma_0=27.6, 27.7, 27.8 \text{ kg m}^{-3}$ isopycnals
 242 are contoured in black. (c) The density along the boundary from $s = 0$ to $s = 1$ averaged in the
 243 upper layer (1-300m, red curve) and the intermediate layer (301-1500m, blue curve). The
 244 maximum vertical transport (black curve) accumulated along the boundary from $s = 0$ to $s = 1$.
 245 The upward slopes indicate the downwelling (LD-Labrador Downwelling, $s = 0.5\sim 1$, ID-
 246 Irminger Downwelling, $s = 0\sim 0.33$), and the downward slope indicates upwelling (LU-Labrador
 247 Upwelling, $s = 0.33\sim 0.5$).

248

249 A vertical section of the DKS-FC along-shore average of the cross-shore velocity crossing
 250 different isobaths is shown in Figure 4a. It reveals a cell-like structure in the upper and lower
 251 layers, with a cell center in the isobath 2400 m at 100 m depth. In line with theory and
 252 experimental works (Spall, 2008 and Cenedese, 2012), the cross-shore velocity increases from
 253 the coast offshore to a maximum near the 2400 m isobath and then decreases offshore until the
 254 along-shore density gradient becomes insignificant at circa the 3000 m isobath. The mean
 255 maximum vertical transport along the SPG boundary is 2.12 Sv at 1088 m, with an uncertainty of
 256 0.43 Sv (2 times STD) estimated from 12000 bootstrap estimates. The transport in density space
 257 is $2.11 \pm 0.28 \text{ Sv}$ by crossing the nearly-horizontal 27.73 kg/m^3 isopycnal surface. Note that this

258 does not fully account for water mass formation near the boundary since the diapycnal flux is
 259 also carried out by the along-stream densification in the upper layer and cross-stream eddy
 260 circulation in the deeper layer (Brüggemann & Katsman, 2019).
 261



262

263 **Figure 4.** (a) The along-shore average of the cross-shore velocity along the SPG averaged at the
 264 100 m-spaced isobaths between 700 m and 3000 m. The x-axis represents distinct isobaths rather
 265 than the cross-shore distance. The dashed line indicates the isobath (2400m) where the maximum
 266 vertical transport is found. (b)(c) The corresponding geostrophic streamfunctions obtained by
 267 accumulating the vertical transport from the surface in depth space and density space. The errors
 268 in gray patch are one standard deviation derived from the bootstrap simulations. The dashed lines
 269 indicate the depth (density) of maximum overturning. Panels (d)-(l) are the same as (a)-(c), but
 270 for the three regions defined in Figure 1 and Figure 2c and for the four seasons: Winter (Jan-
 271 Mar), Spring (Apr-Jun), Summer (Jul-Sep), Autumn (Oct-Dec). The uncertainty estimates and
 272 maximum overturning depths for each region are represented in Table 1.

273

274 The cumulative transport (Figure 3c) along the 2400 m isobath reveals that the entire DKS-FC
 275 boundary can be divided into three distinct areas: a downwelling region in the Irminger Sea (ID,
 276 $s=0$ to $s=0.33$), an upwelling region in the eastern Labrador Sea (LU, $s=0.33$ to $s=0.5$), and a
 277 downwelling region in the western Labrador Sea (LD, $s=0.5$ to $s=1$). These
 278 downwelling/upwelling patterns result from the downstream density increase/decrease along the
 279 boundary (Figure 3a). The cross-shore velocities averaged along the isobaths in these
 280 three regions are shown in Figure 4d-g-j. They also capture cell-like structures within the
 281 boundary current system, with waters flowing inshore (offshore) in the upper layer and offshore

282 (inshore) in the lower layer of the downwelling (upwelling) region. Maximum cross-
 283 shore transports, which are consistently observed at the 2400-m isobath, are of comparable
 284 magnitude in each of the three regions with upwelling (2.17 Sv) along the western slope of
 285 Greenland balancing half of the total downwelling (4.28 Sv) (Figure 4ehk).

286
 287 Water sinks (upwells) all year in the downwelling (upwelling) regions. The Eulerian-mean
 288 transport in the LD region is 2.23 ± 0.71 Sv (Figure 4e and Table 1) at a depth of 986 m. The
 289 transport in the LU region (Figure 4h and Table 1) is 2.17 ± 0.74 Sv at a depth of 1182 m. The ID
 290 has similar downwelling of 2.05 ± 0.79 Sv to the LD (Figure 4k and Table 1) at a depth of 1304
 291 m. The regional density-space transport follows a very similar pattern to that in depth-space, and
 292 the density level of maximum transport remains nearly constant at around 27.71 to 27.76 kg/m³.
 293 The uncertainty estimates in Table 1 show the standard deviations calculated from 3000
 294 iterations using the three datasets. Seasonal and regional errors are all between 0.3 and 0.8 Sv.
 295 The seasonal variations are of the same magnitude as the uncertainties derived from the bootstrap
 296 estimates based on the three datasets, and the present estimates of seasonal changes of the
 297 Eulerian-mean downwelling are not statistically significant for the whole SPG or individual
 298 basins. This is in line with Li et al., (2021) who did not detect a statistically significant signal
 299 cycle when estimating the composite monthly mean MOC transport using the OSNAP program's
 300 46-month records. Furthermore, Sayol et al. (2019) used a high-resolution model to find that
 301 seasonal variability in boundary sinking is rather insignificant in comparison to interior sinking,
 302 and that the variability is thought to be driven by ageostrophic dynamics that our large-scale
 303 geostrophic balance cannot capture. Nevertheless, based on our observations, a potential upper
 304 bound on the seasonal cycle amplitude has been established.

305
 306

Season	LD	LU	ID	SPG
Winter	1.97 ± 0.30 (970)	-2.56 ± 0.44 (1108)	2.57 ± 0.31 (1190)	1.99 ± 0.58 (975)
Spring	2.27 ± 0.31 (1274)	-2.04 ± 0.31 (1288)	1.92 ± 0.61 (1278)	2.14 ± 0.81 (1253)
Summer	2.16 ± 0.28 (923)	-1.81 ± 0.34 (1301)	1.84 ± 0.84 (1425)	2.19 ± 0.82 (903)
Autumn	2.54 ± 0.51 (891)	-2.27 ± 0.60 (1084)	1.88 ± 0.45 (1372)	2.14 ± 0.47 (1044)
Annual	2.23 ± 0.71 (986)	-2.17 ± 0.74 (1182)	2.05 ± 0.79 (1304)	2.12 ± 0.43 (1088)

307 Table 1. The maximum vertical transport (Sv) derived from the multiproduct mean in the LD,
 308 LU and ID regions (defined in Figure 2) and whole SPG. The errors are two times of standard
 309 deviation derived from the bootstrap simulations. The depths (m) of the maximum transport are
 310 displayed in parentheses.

311

312 4 Discussion and Conclusion

313 We have provided here a first observation-based estimate of the Eulerian-mean downwelling
 314 along the boundary of the SPG. Using *in situ* temperature and salinity profiles from sustained
 315 global ocean observational networks, we mapped the long-term mean density along the
 316 continental slope from the Denmark Strait to Flemish Cap. The cooling and resulting
 317 densification along the boundary region result in a geostrophic cross-shore flow and a
 318 downwelling above the continental slope, with a maximum value of 2.12 ± 0.43 Sv at 1100 m
 319 depth. The transport in density space, which is here representative of the cross-shore diapycnal
 320 flux, is very similar to the transport in depth space. Cross-shore velocity is found to increase

321 offshore and peak near the 2400 m isobath, before eventually disappearing near the 3000m
322 isobath, in good agreement with observation-based descriptions of the distinct circulation of the
323 interior and the boundary current (Pacini et al., 2020), and with theoretical or modelling works
324 suggesting limited mean advective exchanges between those two regimes (Cenedese, 2012;
325 Brüggemann and Katsman 2019). Our analysis further reveals no strong or statistically
326 significant seasonality in the magnitude of downwelling across the entire SPG.

327

328 Interestingly, the total Eulerian-mean downwelling from CF to FC is estimated as only 0.1 ± 0.03
329 Sv due to compensation between upwelling along the western slope of Greenland and
330 downwelling in the remaining portion of the Labrador Sea. Downwelling was estimated to be 1.4
331 Sv in that region by Katsman et al. (2018) in high-resolution simulations. Western Greenland is
332 in fact generally described as a region with strong along-shore densification due notably to eddy-
333 driven heat flux, as described in several modelling works (Katsman et al., 2004; Georgiou et al.,
334 2019). This apparent discrepancy could arise for several reasons. Warming in the intermediate
335 layer might be overly represented due to an uneven distribution of profiles along western
336 Greenland's narrowing boundary, which may be insufficient to represent the boundary current's
337 features. It is also possible that the subsurface warming along western Greenland (also observed
338 in other estimates, see, e.g., Palter et al., 2008) indicates a signal of restratification following
339 baroclinic instability. The tilted isopycnals on the offshore side of an unstable current should
340 deepen downstream and drive a warming signal in this portion of the water column.
341 Additionally, the cross-shore transport along the narrowing boundary might represent the
342 deflection of upper layer water into the interior. An offshore mass flux, resulting from either a
343 flux of less dense water off the shelf or a broadening of the boundary current due to baroclinic
344 instability, would be diagnosed as upwelling because of the assumption of no net flow across the
345 isobaths. Future work may focus on describing the properties in shallower depths further inshore
346 when sampling is denser. Nevertheless, the net downwelling in SPG is not sensitivity to the
347 distribution of boundary downwelling or upwelling, as demonstrated by the fact that the strong
348 downstream warming and subsequent upwelling in the eastern Labrador Sea are compensated by
349 the significant downstream cooling and subsequent downwelling in the western Labrador Sea.
350 Indeed, the full-basin integrated downwelling only depends on the density variation between the
351 DKS and FC and thus remains a robust estimate.

352

353 Further comparison with independent overturning estimates from cross-basin arrays or realistic
354 modelling provides insights and confidence in the values reported herein. The Irminger Sea
355 downwelling is estimated to be 2.05 ± 0.79 Sv at 1200 m depth, in line with the high-resolution
356 modelling estimates of 1.4 Sv from Katsman et al. (2018) and 1.13 Sv from Sayol et al. (2019).
357 Summer downwelling in the Labrador Sea is estimated to be 0.85 ± 0.15 Sv at depth 685 m by
358 integrating the transport between the two ends of the AR7W line (Figure 2), which agrees
359 quantitatively with Pickart and Spall's (2007) observation-based meridional transports of about 1
360 Sv at depth 800 m. The annual mean downwelling rate in the Labrador Sea is estimated here as
361 0.72 ± 0.07 Sv, similar to Holte and Straneo's (2017) and Lozier et al. (2019) estimates, which
362 amount to 0.9 Sv and 0.8 Sv, respectively. This also confirms that the majority of sinking occurs
363 near the Labrador Sea's boundary, with little vertical mass transport occurring in the convective
364 interior.

365

366 Furthermore, we quantify here only the western boundary downwelling downstream of DKS,
367 omitting potential upstream contributions around Reykjanes Ridge (RR) or the Rockall Plateau,
368 for instance. The zero-mass constraint used herein to estimate cross-shore overturning is likely
369 inadequate in such regions where the vertical integral of the cross-shore geostrophic flow is not
370 zero (consider for instance the net westward flow above the crest of RR). Additionally,
371 overflows across the DKS or the Iceland-Scotland Ridge contribute significantly to the overall
372 net sinking in the SPG – about 7 Sv (Hansen & Østerhus, 2000) – via hydraulically-controlled
373 dynamics not captured by our geostrophic estimate. A refinement of the boundary sinking
374 calculation is being investigated for these specific areas.

375
376 Ekman transport was estimated from the long-term mean (2002-2015) along-shore wind stress
377 averaged from the NCEP/NCAR Reanalysis 1 and ERA5 dataset. It carries 0.7 to 0.9 Sv of water
378 inshore of the DKS-FC boundary. Two-dimensional models require that this onshore flow
379 returns to the basin interior (e.g. Lentz and Chapman, 2004; Choboter et al., 2011), but three-
380 dimensional models with spatially variable winds (or coastlines) demonstrate that some of this
381 water flows along the shelf (e.g. Allen, 1976; Suginohara, 1982). It is difficult to determine a
382 priori the partition of this downwelled water between along-shelf and interior pathways as it will
383 depend on various factors such as stratification, topography, nonlinearity, wind pattern, etc.
384 However, any of the Ekman transport that does flow back into the interior will remain relatively
385 shallow and have a negligible effect on the deep overturning circulation, and so is not included in
386 our estimate.

387
388 We finally emphasize that novel descriptions of the Eulerian-mean downwelling rooted in
389 observations are critical for understanding AMOC variability in the current context of
390 anthropogenic climate change (IPCC, 2021). Both upper-ocean warming and increased ice
391 melting may significantly affect along-boundary density gradients and result in the reduced
392 sinking and a consequent weakening of the AMOC. A better observation-based understanding of
393 boundary-focused vertical transport may also help us to properly represent or parameterize the
394 details of the underlying boundary dynamics in climate models. It can assist in not only
395 describing the circulation and diagnosing its controls, but also in providing valuable insights into
396 how to improve existing ocean circulation theories of overturning.

397

398 **Data Availability Statement**

399 The In Situ Analysis System-ISAS15 dataset (2002-2015) is available from the SEANOE
400 repository (<https://www.seanoe.org/data/00412/52367/>, last access: July 2021) and described in
401 Gaillard et al. (2016). The Coriolis data set for ReAnalysis-CORA V5.2 (1950-to present) is
402 available from the SEANOE repository (<https://www.seanoe.org/data/00351/46219/>, last access:
403 September 2019) and described in Szekely et al. (2016). The EN4.4.2.2 dataset for years 1990 to
404 present is provided by Met Office Hadley Centre and available at
405 <https://www.metoffice.gov.uk/hadobs/en4/> (last access: November 2021) and described in Good
406 et al. (2013). The A25-Ovide and AR7W hydrography sections are available via the CLIVAR
407 and Carbon Hydrographic Data Office (CCHDO) platform (<https://cchdo.ucsd.edu/>). The
408 NCEP/NCAR Reanalysis 1 (1948/01/01 to present) is provided in National Centers for
409 Environmental Prediction (</data/gridded/data.ncep.reanalysis.html>). The ERA5 dataset (1950-to
410 present) is available in Copernicus Climate Change Service Climate Data Store (CDS),

411 <https://cds.climate.copernicus.eu/cdsapp#!/home>. The present analysis does not use new
412 unpublished data.

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