

1
2 North Atlantic extratropical and subpolar gyre variability during the last 120 years:
3 a gridded dataset of surface temperature, salinity, and density.

4 Part 1: Dataset validation and RMS variability

5
6 G. Reverdin¹, A. R. Friedman², L. Chafik³, N. P. Holliday⁴, T. Szekely⁵, H. Valdimarsson⁶, I.
7 Yashayaev⁷

8
9 ¹ Sorbonne-Université, CNRS/IRD/MNHN (LOCEAN), Paris, France. ORCID ID
10 <https://orcid.org/0000-0002-5583-8236>

11 ² School of Geosciences, University of Edinburgh, UK. ORCID ID [https://orcid.org/0000-](https://orcid.org/0000-0001-6994-2037)
12 [0001-6994-2037](https://orcid.org/0001-6994-2037)

13 ³ Department of Meteorology and Bolin Centre for Climate Research, Stockholm University,
14 Stockholm, Sweden. ORCID ID <https://orcid.org/0000-0002-5538-545X>

15 ⁴ National Oceanography Centre, Southampton, UK. ORCID ID [https://orcid.org/0000-0002-](https://orcid.org/0000-0002-9733-8002)
16 [9733-8002](https://orcid.org/0000-0002-9733-8002)

17 ⁵ IUEM, Brest, France

18 ⁶ Marine and Freshwater Research Institute, Reykjavik, Iceland

19 ⁷ Bedford Institute of Oceanography, Fisheries and Oceans Canada, Dartmouth, NS, Canada

20
21
22 Corresponding author: G. Reverdin, Laboratoire d'Océanographie et de climatologie par
23 expérimentation et analyse numérique, Institut Pierre Simon Laplace, Sorbonne-Université,
24 case 100, 4 pl. Jussieu, 75252 Paris Cedex 05, France (gilles.reverdin@locean-ipsl.upmc.fr ;
25 tel : 33-1-44272342 fax : 33-1-44273805 ; orcid.org/0000-0002-5583-8236)

26
27 Published in Ocean Dynamics, Springer. DOI: 10.1007/s10236-018-1240-y

28

29 Abstract

30 We present a binned annual product (BINS) of sea surface temperature (SST), sea surface
31 salinity (SSS) and sea surface density (SSD) observations for 1896–2015 of the subpolar
32 North Atlantic between 40°N and 70°N, mostly excluding the shelf areas. The product of bin
33 averages over spatial scales on the order of 200 to 500 km, reproduces most of the interannual
34 variability in different time series covering at least the last three decades or of the along-track
35 ship monitoring. Comparisons with other SSS and SST gridded products available since 1950
36 suggest that BINS captures the large decadal to multidecadal variability. Comparison with the
37 HadSST3 SST product since 1896 also indicates that the decadal and multidecadal variability
38 is usually well reproduced, with small differences in long-term trends or in areas with
39 marginal data coverage in either of the two products. Outside of the Labrador Sea and
40 Greenland margins, interannual variability is rather similar in different seasons. Variability at
41 periods longer than 15 years is a large part of the total interannual variability, both for SST
42 and SSS, except possibly in the south-western part of the domain. Variability in SST and SSS
43 increases towards the west, with the contribution of salinity variability to density dominating
44 that of temperature in the western Atlantic, except close to the Gulf Stream and North
45 Atlantic Current in the southwest area. Weaker variability and larger relative temperature
46 contributions to density changes are found in the eastern part of the gyre and south of Iceland.

47

48 Keywords: sea surface temperature, sea surface salinity, surface density, North Atlantic,
49 decadal variability

50

51
52
53
54
55
56
57
58
59
60
61
62
63
64
65
66
67
68
69
70
71
72
73
74
75

1. Introduction

The North Atlantic subpolar gyre (SPG) receives salty surface water from the subtropical North Atlantic as well as fresh water originating from the Arctic and the Nordic seas and circulating mostly on and along the shelves. It is a formation site of deep and intermediate waters and thereby a major contributor to the lower limb of the Atlantic meridional overturning circulation (AMOC) (Rhein et al. 2011). There is wintertime densification of the surface Labrador and Irminger Seas resulting in intermittent deep convection reaching and exceeding 2400 m in the mid-1990s and 2000 m in recent years (Yashayaev and Loder 2017). Deep dense overflows from the Nordic Seas and the associated mixing with and entrainment of local waters also add important hydrographic connections between the regions and the surface and deep layers. In both instances, one expects that changes in surface density will modulate these vertical exchanges and the properties of the deep and intermediate waters formed, and thus the AMOC. Many model studies (at various spatial resolution) also suggest an influence of surface salinity changes in this region on changes in the AMOC (Frankignoul et al. 2009; Rahmstorf et al. 2015, Böning et al. 2016). Whereas the connection between these surface changes and AMOC is still subject to discussion (Lozier 2012; Williams et al. 2015; Buckley and Marshall 2016), the relation between changes of surface density and upper or intermediate water formation is rather well established (Yashayaev and Loder 2016 2017; Piron et al. 2017). Seawater density is a function of temperature and salinity, with the relative influence of salinity increasing when temperature decreases. Thus, investigating surface density variability and attempting to establish its past patterns of change requires a concurrent analysis of temperature and salinity fields.

76
77
78
79
80
81
82
83
84

Hydrographic data in the North Atlantic and neighboring seas have been used to analyze salinity and temperature on decadal time scales with a low spatial resolution, but only since the mid-1960s and in more restricted areas since 1950 (Polyakov et al. 2005; Skliris et al. 2014; Holliday et al. 2015; Yashayaev 2007; Yashayaev and Loder 2016 2017). These data also rarely provide direct indications of winter surface density field (regional estimates can be found in Yashayaev and Loder (2016 2017)) that is the variable key to the connection between the surface ocean and the ocean interior. In the last twenty years, the advent of profiling floats (Lavender et al. 2000; Roemmich et al. 2009; Riser et al. 2016) has allowed

85 year-round three-dimensional monitoring of large-scale upper and intermediate ocean
86 variability (Tesdal et al. 2018), but these measurements do not yet help to resolve the decadal
87 to multidecadal frequencies. Multidecadal time series with at least annual or seasonal
88 resolution have also been produced at a few sites: Rockall Trough (Holliday et al., 2015), the
89 Faroe-Shetland Channel (Hughes et al., 2012), south of Iceland (Icelandic hydrographic
90 surveys), in the center of the Labrador Sea (Yashayaev and Loder 2016 2017), or at station
91 Mike and hydrographic sections in the Norwegian Sea (Yashayaev and Seidov 2015).
92 Otherwise, mapped monthly fields of T and S have been constructed by different objective
93 mapping methods based on the EN4 (Good et al. 2013), CORA (Cabanes et al. 2013) and
94 Ishii et al. (2006) datasets that could have some coverage in a large part of this domain for the
95 last 65 years. Monthly mapped sea surface temperature (SST) fields have also been produced
96 with sufficient coverage into large parts of the SPG since the 1890s using mostly ship log data
97 with a shift in the last two decades to drifter SST data. Carefully estimated bias correction has
98 been applied to subsets of these data, such as done in Hadley Centre SST (HadSST3;
99 Kennedy et al. 2011a and 2011b). However, these binned SST datasets present gaps in some
100 northern regions, in particular before 1922.

101
102 Long-term box averages in the SPG with near-annual resolution were presented in Reverdin
103 (2010) and Friedman et al. (2017). Reverdin (2010) described a record of SST and SSS in the
104 northeastern part of the SPG from 1895–2009. There, co-variability of temperature (T) and
105 salinity (S) was found on decadal or multidecadal time scales in different seasons, with a
106 slight dominance of temperature on density variability (partially compensated by the salinity
107 contribution). Friedman et al. (2017) examined interannual (1-2-1 smoothed over successive
108 years) salinity time series in larger boxes over the Atlantic from 20°S–70°N (the temperature
109 reports associated with the salinity data were not examined). The largest salinity root mean
110 square (RMS) variability was found in the northern tropics, from 5°–20°N. In the subpolar
111 North Atlantic, RMS variability was shown to decrease to the northeast from the Labrador
112 Sea to the Nordic seas (Friedman et al. 2017, Fig. 1), though the large grid box size did not
113 permit finer examination of the spatial structure.

114
115 Using similar box averaging as in these previous two studies, we present a higher-resolution
116 binned product of temperature and salinity since 1896 in the subpolar North Atlantic and
117 southern Nordic seas and draw first conclusions obtained with the dataset. Focusing on the
118 region north of 40°N allows us to describe co-located SSS and SST at higher spatial scales

119 (less than 500km) than presented previously, without having large periods with data gaps
120 (except during or just after WWI and WWII in some sub-regions).

121

122 In this paper we present and validate the binned product at interannual time scales, and also
123 discuss its RMS variability. A second part of this study (in preparation) examines the decadal
124 and multidecadal variability, and relationship with North Atlantic climate variations. The
125 dataset sources and construction are described in Section 2. In Section 3 we validate the
126 dataset and compare it with other products. In Section 4, we examine the spatial distribution
127 of temporal variability of SSS, SST, and surface density, followed by summary and
128 discussion in Section 5.

129

130

131 **2. Data and methods**

132 **2.1. Dataset construction**

133 The main data sources are described in Reverdin et al. (1994), Reverdin (2010), and Friedman
134 et al. (2017). A significant part of the temperature and salinity (reported as practical salinity)
135 data was collected on merchant and other selected vessels (including the weather ships), in
136 particular before the mid-1970s north of 50°N, and before WWII and since 1977 south of
137 50°N. This is also the case since 1993, mostly along two ship tracks between southern
138 Newfoundland and Reykjavik and between the North Sea and Greenland (Reverdin et al.
139 2018), as well as near France and between the English Channel and eastern North America.
140 For the first two tracks, intake temperature was measured part of the time, XBTs were
141 dropped and water samples collected, which allows for identification / estimation of possible
142 temperature and salinity biases in the records from thermosalinographs (TSG) (Alory et al.
143 2015; Reverdin et al. 2018). The identification of the temperature biases is less certain for the
144 other vessels since 1993, but water samples were also collected to correct salinity biases in
145 TSG records. This is complemented by near-surface hydrographic data during cruises, and
146 since 1996, by measurements from drifters, and upper level data (often near 5-8m depth) from
147 Argo floats (Roemmich et al. 2009; Riser et al. 2016), as well as from earlier PALACE floats
148 (Lavender et al. 2000; Davis et al. 2001).

149

150 The depth of sampling and the methods of collection and analysis have changed in time, as
151 well as data accuracy, and thus some subsets of the data require correction of identified
152 biases, both for temperature and salinity, as commented in the earlier papers. Bias correction

153 follows Reverdin et al. (1994) for ship data before the 1990s (based on comparison with
154 surface data from Nansen casts). There is nonetheless the possibility of seasonal stratification
155 between the data collected below the surface (for example, from TSG data, Argo floats and
156 CTD/Nansen casts) and the surface data (for example from drifters, or in earlier times from
157 bucket sampling). In particular, the change in depth of sampling in time can result in artificial
158 time variability, due to near-surface stratification. In other parts of the world, this has been
159 documented due to surface heating and evaporation or precipitation (Boutin et al., 2016). In
160 regions of sustained wind conditions, and away from shelves with surface freshwater sources
161 (melting sea ice or icebergs, outflows from fjords, river plumes), we expect stratification on
162 the order of or less than 0.01 in practical salinity and 0.1°C in temperature. Sustained wind
163 conditions are commonly melt in all seasons near 55-65°N, and probably a little less so in
164 particular in spring/summer further south and north.

165

166 We construct annual time series of deviations from the seasonal cycle in boxes for four
167 different seasons, using a similar methodology as in Friedman et al. (2017) and applied from
168 1896–2015 for both T and S, over smaller spatial bins. This new box-product will be referred
169 to as BINS. The boundaries of the 34 boxes were redesigned to better fit with oceanic fronts
170 and main currents than in Friedman et al. (2017). One box in the central Labrador Sea
171 corresponds to what is used by Yashayaev and Loder (2016) for creating a combined
172 Labrador Sea time series, and box boundaries in the eastern subpolar North Atlantic are
173 chosen as in Holliday et al. (2015). Except for one box off southwest Greenland and one box
174 on the southern part of the Grand Banks, the boxes do not overlap with the shelves and with
175 areas of large seasonal sea ice cover. In the northeastern SPG, the resulting boxes provide a
176 slightly lower spatial resolution than what was used in Reverdin (2010), but for which the
177 different time series were rather well correlated. To give an idea of the resolution achieved,
178 the SSS and SST late-winter climatologies are mapped onto the BINS grid boxes (**Fig. 1**),
179 which shows that the main characteristics of spatial variability are retained by this choice of
180 grid.

181

182 The construction of box time series and their error estimates is described in Friedman et al
183 (2017). Here, the main steps are summarized, with specifics and additions related to BINS
184 further described. The SSS and SST climatologies were constructed on a 1°×1° grid with little
185 spatial smoothing and gap-filling, with an adjustment to the mean and up to three sinusoidal
186 harmonics of the year, when data was sufficient. Deviations from this climatology of

187 individual data are estimated (T being in °C and S reported in the practical salinity scale 1978
188 (PSS-78; Fofonoff 1985; UNESCO 1981). For T, outliers were removed using a spatially
189 variable threshold (between 3°C and 6°C), inspired by the local standard deviation. We
190 compared this to a fixed 3°C error threshold, which yielded generally similar results albeit
191 with smaller interannual excursions. Largest differences were found in the central Labrador
192 Sea in the 1920s, a period with extremely low data coverage. S outliers were removed
193 similarly, with less influence on the variability. In each box, these individual deviations are
194 then combined for the individual seasons and years, by median-averaging the individual
195 deviations. The seasonal anomalies are 1-2-1 smoothed over successive years (not applied at
196 the start and end years) and then averaged annually from December–November, inversely
197 weighted by error. [The annual period was incorrectly stated as March–February in Friedman
198 et al. (2017)]. Before being combined, the seasonal anomalies were adjusted to a common
199 baseline over the 40-year period from 1956–1995, which was chosen due to data coverage.

200
201 The fall season (September–November) is not included in most areas north of 45°N in the
202 western Atlantic, including in the central Labrador and southwest Labrador boxes due to
203 poorly correlated T or S time series in the fall with the other seasons in these regions. In the
204 Norwegian Sea, central Labrador, southwest Labrador, and West Greenland, the winter
205 season (December–February) data are not included before 1947 due to insufficient coverage.

206
207 The total number of years with data is shown in **Fig. 2a**. Coverage is below 100 years in the
208 North Irminger Sea and central Labrador Sea grid boxes; all other grid boxes have more than
209 100 years of coverage. The periods of missing data occur mainly during WWI and WWII, and
210 around 1900 (**Fig. 2b**). Multi-year data gaps are linearly interpolated, and attributed an error,
211 equal to the standard deviation of the whole time series (which is probably an overestimate).
212 A few missing end years (mostly 1896) in some boxes were filled by extending the first non-
213 missing value. As in Friedman et al. (2017), regional average error terms are constructed by
214 RMS averaging the error terms of their constituent individual boxes (which assumes that the
215 errors are uncorrelated in different boxes). The BINS SSS regional average fields are
216 compared with those of Friedman et al. (2017) in Appendix A for the common period 1896-
217 2013, showing differences most commonly within the ± 2 standard error ranges.

218
219 **Figs. 2c–2d** present an estimate of the average interannual grid box error as the mean
220 error term divided by the error term during the year when there was no coverage and the error

221 is maximum (1943, except in the West Greenland box, when it is 1917). This is thus an
222 indication of the ratio of noise relative to signal. This shows for S, the smallest relative errors
223 are in the central SPG. For temperature, the largest relative errors are in the western part of
224 the domain, where intra-annual variability and eddy variability are largest and not adequately
225 sampled in this data set. The normalized area-averaged error for SSS and SST is shown in
226 **Fig. 2e** and mirrors data coverage, with a larger value in the last year (2015) because error is
227 not reduced in that year by the filtering.

228

229 Density is computed from the annual binned SST and SSS values using EOS80 (UNESCO,
230 1981), and referring the annual anomalies to the March average T and S values. Because SST
231 is lower in winter and density becomes less sensitive to temperature variations when
232 temperature decreases, this choice slightly decreases the dependency of density on SST
233 compared to using an annual average reference. What motivated this choice was that we
234 wanted to have an estimate typical of winter conditions. Density errors are calculated from the
235 errors in interannual SSS and SST, assuming that they are uncorrelated.

236

237 **2.2 Seasonal dependency of the interannual variability**

238 The combination of the four seasonal time series to create an annual time series in each bin
239 implicitly assumes that the four seasons portray comparable signals/patterns of interannual
240 variability. This is particularly important if one season is missing, as is the case for the winter
241 season in some boxes for the first part of the records. To get a sense of how well this holds for
242 individual seasons, we combine the time series of individual bins into seven main multi-box
243 regions (**Fig. 3a**): the central Labrador Sea / West Greenland Shelf (3 grid boxes), the
244 southern Nordic seas (3 grid boxes), the north-east part of the SPG (5 grid boxes), the central
245 SPG (6 grid boxes), a large central area of the intergyre gyre south of 50°N (8 boxes), an
246 eastern Atlantic region (two boxes), and four boxes in the southwestern region around the
247 Grand Banks. The regions are chosen for similar source data and coherent variability signals
248 (except for the last one). For example, the two Norwegian Sea grid boxes contain mostly
249 Norwegian and Swedish surface data before WWII, and afterwards much hydrographic data.
250 The central Labrador Sea / West Greenland region contains hydrographic data from the Ice
251 Patrol and other cruises in the southwest, and a mix of hydrographic data and merchant ship
252 data until 1960 in the shelf box close to Greenland. South of 50°N, many different countries
253 contributed to the surface data before WWII, and the French data sources are a large
254 contribution since the mid-1970s. Three grid boxes are not included in any of these regions

255 due to low seasonal correlations in S and more varied data sources: the southwest Labrador
256 Sea, south Greenland/ southwest Irminger Sea, and the western area north of the Gulf Stream.
257

258 The seasonal correlations are summarized in **Table 3**. For the central SPG and north-east SG
259 regions (**Fig. 3b-c**, rows 1 and 2), the time series are most complete in all seasons. There, the
260 different seasons present rather similar variability, both for T and S, although for T, the
261 summer (June-August season) deviates at times from the other seasons and presents more
262 variance in the central SPG. Furthermore, after 1960, S anomalies in Sep-Nov in central SPG
263 tend to be below the ones in the other seasons, and the opposite before. This might either be
264 the sign of a real long-term seasonal change, or the effect of changes in the dataset around
265 1960 (transition in this region from bucket collection to other means of collecting water, for
266 example).

267
268 In the southern Nordic seas (**Fig. 3b-c**, bottom row), the different seasons also present rather
269 similar variability, at least since 1950. Before that, this remains largely true for SSS, albeit
270 with more variability in summer. For SST, this also holds before WWI, however between
271 1920 and 1940, the different seasons present non-coherent variability. We suspect that this
272 might be caused by inhomogeneities in the dataset, either spatially, with a southward shift of
273 the sampling in late autumn to early spring in this period, or due to lower data quality and
274 larger random as well as systematic errors affecting the results and lowering correlations,
275 such as the documented use of engine room or intake temperature data from Norwegian
276 vessels. For this reason and the limited winter sampling coverage, we chose not to incorporate
277 the winter season in the annual averages before 1947 in the two Norwegian Sea boxes.

278
279 In the central Labrador Sea / West Greenland Shelf region (**Fig. 3b-c**, row 3), on the other
280 hand, while there is some similarity between the different seasons in SST (less so for the
281 summer season), there is much less coherence in SSS, even during the last 50 years of more
282 regular sampling. Notice in particular the large differences between the spring and autumn
283 seasons, which originate from the two boxes in central Labrador Sea and along west
284 Greenland. These differences (and the much larger RMS variability in autumn) justify the
285 choice of not including the autumn season when combining the different seasons to create the
286 BINS time series in this region. In the Grand Banks region, however, we have less correlation
287 with winter season but with similar variance (**Table 3**), so that we chose to retain the winter
288 season. Elsewhere, the statistics associated with the comparison between the seasonal time

289 series (**Table 3**) justify the choice to have retained all seasons in order to reduce the sampling
290 uncertainty in the BINS interannual time series.

291

292 **2.3 Other products**

293 In the next section, we validate BINS with some of the few published compiled hydrographic
294 time series that are available in the SPG, either from repeated seasonal cruises or a
295 compilation of data from different origins: the Rockall Trough time series since 1975
296 (Holliday et al. 2015); South Icelandic stations (Sevlogsbanki 5 (Sb5) and Stokksnes 5 (St5))
297 since the 1970s; and the central Labrador Sea since 1941 (Yashayaev and Loder, 2009 2016).
298 Notice also the time series in the Norwegian Sea summarized by (Yashayaev and Seidov
299 2015). We will also compare BINS to gridded datasets. These include the $1^{\circ}\times 1^{\circ}$ EN4, version
300 4.2.1 (Good et al. 2013) fields from 1950-2015, as well as the 1950-2012 $1^{\circ}\times 1^{\circ}$ product of
301 Ishii et al. (2006), version 6.12 (referred to as ISHII). We also compared BINS to the
302 CORA5.0 fields (Cabanes et al. 2013) that have been extended to 1950, but that exclude
303 Nansen cast data, and thus cannot be used for the analysis of salinity before 1980, and have
304 much coarser spatial sampling in temperature than EN4 or ISHII. On the other hand, when
305 sampling is sufficient (mostly, in the northeastern part), results with CORA are comparable to
306 those of ISHII, and will not be discussed further. For long-term surface temperature, we also
307 compare BINS with the HadSST3 monthly product (Kennedy et al. 2011a and 2011b).

308

309

310 **3. Dataset validation**

311 In order to validate the BINS SST and SSS time series, we compare it first to local
312 compilations of high-quality hydrographic data. We also compare the binned time series to
313 different gridded time series to better characterize the properties of BINS. In addition,
314 Appendices A and B show comparisons with Friedman et al (2017) and Reverdin et al.
315 (2018), which are based on different binning (in time and space) of similar data sources.

316

317 **3.1 BINS and hydrographic time series**

318 The sites of the different time series are presented in Fig. 4a. The Rockall Trough and central
319 Labrador time series fit rather well within some BINS grid boxes, and the South Icelandic
320 time series are just to the north of one BINS grid box. The results based on hydrographic time
321 series often comprise vertical averages, excluding the near-surface (5 to 10 m thick) layer,
322 which is often determined by specifics of performing oceanographic stations and

323 technological limitations. This averaging depth range was not optimized, but in Rockall
324 Trough and near Iceland, it fits with the expected local maximum winter mixed layer. The
325 time series are then averaged over annual means (after removal of an annual cycle), and
326 smoothed with a 1-2-1 filter over successful years, similar to BINS.

327
328 The comparisons of these upper ocean time series with BINS are presented in **Fig. 4b** (with
329 statistics in **Table 4**). At each site BINS is correlated with the vertically integrated time series,
330 although the surface time series tend to have larger amplitudes. In the Labrador Sea, we
331 compared different ranges, with the best correlation for the comparison to 20-50 m, which is
332 within the mixed layer at least 6 months of the year and closely follows variations in seasonal
333 discharge of freshwater and its interannual variations. There is still a correlation when
334 considering the averages over the 20-200 m layer, which is within the winter mixed layer
335 most years. The thicker layer time series however present less extreme amplitudes than the
336 shallower ones. In the Rockall Trough, we also plot the surface time series based only on
337 winter data, which is also correlated with the vertically integrated time series. The largest
338 difference is found for South Iceland, with a low BINS S in the late 1980s that is not found in
339 the hydrographic time series. However, there is also a difference in latitude between the BINS
340 box and the site further north of the hydrographic time series.

341
342 Overall, we find that BINS surface time series are significantly correlated at 0-lag with upper
343 ocean time series from hydrography, both for T and S (correlation coefficients are in the
344 range 0.73 to 0.88 for the different time series of salinity or temperature), and exhibit very
345 comparable RMS variability, as described in **Table 4a**. The analysis of lag correlation
346 (summarized in **Table 4b**) indicates that the annually averaged BINS T and S precede by one
347 year the vertically integrated upper ocean analysis both at the Rockall and Southern Iceland
348 sites (for SST, the one-year lag correlation coefficients are respectively 0.80 and 0.89; and for
349 SSS, respectively 0.77 and 0.88), as expected from winter ventilation of the upper ocean and
350 time series analysis of the 20 to 40-year long weather-ship time series in the SPG (Reverdin et
351 al., 1997). Interestingly, this is not the case for the Norwegian Sea time series.

352

353 **3.2. BINS and gridded objective analyses**

354 We compare the BINS time series with other gridded products of hydrographic data at their
355 near-surface level that we average over the bins of the BINS analysis and filter in time to

356 correspond to the same time resolution (1-2-1 running average over successive years
357 December to November).

358

359 Because of the differences in sampling and mapping, the comparison in individual boxes with
360 BINS often presents a large scatter. We will thus average the time series over the same
361 domains as for the seasonal time series (**Fig. 3a**). [One difference is that the central SPG and
362 north-east SPG are combined into one SPG region]. Except in the Labrador Sea or near the
363 Grand Banks, the time series of the three products (**Fig. 5**) are well correlated (**Table 5**).
364 Correlation coefficients are larger for SST than for SSS, and for SSS, are usually larger
365 between ISHII and BINS than between EN4 and BINS. In the Labrador Sea (row 2), there are
366 differences both for SST and for SSS. For SSS, the large positive peak in BINS originates
367 from a data gap in one of the three bins that are averaged in the regional time series. On the
368 other hand, the altogether weaker SST-signal in ISHII in this box, something also found, but
369 to a lesser domain in the other regions might result from the objective mapping method used
370 or from more strident tests on outliers (Ishii and Kimoto, 2009). We also notice in the
371 southern Labrador Sea for the last 20 years smaller positive SSS (and SST) anomalies in
372 BINS than in the other two products, possibly a result of the thermosalinograph (TSG) data
373 incorporated in BINS and not in the other products (Reverdin et al. 2018). On the other hand,
374 near the Grand Banks during the last 20 years, T is higher in BINS than in the other products.
375 Elsewhere, the major variability is portrayed with similar amplitude and roughly at the same
376 time in the different products. For example in the SPG or southern Nordic seas, the salinity
377 time series portray the events sometimes described as great salinity anomalies (Belkin et al.
378 1998), the largest one in the central SPG happening in the early 1970s and referred to as the
379 Great Salinity Anomaly (Dickson et al. 1988). The figures also suggest larger differences
380 between the different products in the 1990s, in particular in the intergyre region.

381

382 The high overall agreement might result from a share of common hydrographic data
383 incorporated in the different products. However, there is also a large portion of specific data
384 that are not in all the products (T from mechanical or expandable bathythermographs are used
385 in EN4 and ISHII, but not in BINS, and different sets of surface data are in BINS, but not
386 used in EN4 and ISHII). There are also large methodological differences between the
387 different products, both in data selection and validation and in mapping techniques and how
388 seasonal anomalies are grouped or not. These comparisons, as well as the comparisons
389 presented in 3.1, thus reinforce our assessment of usually small sampling errors for the BINS

390 product during that period, and that the product captures the major events of interannual
391 variability at the grid box scale since 1950.

392

393 **3.3 BINS and HadSST3**

394 SST data in BINS and HadSST3 originate from different data streams. Although some data
395 are certainly incorporated in both products, the overlap is probably rather minimal, as
396 discussed in Reverdin et al. (1994). In particular, the data before 1920 north of 58°N in BINS
397 originate mostly from Danish and Norwegian sources which are not incorporated in
398 HadSST3. In the last two decades, data from TSGs on merchant vessels or from profiling
399 floats are a large contribution to BINS but are not included in HadSST3, whereas the drifter
400 SST data used in HadSST3 are mostly not in BINS. Even in the post-WWII to mid-1970s
401 period, when ocean weather ships contribute both to HadSST3 and BINS, temperature data in
402 the two products often originate from different sensors. BINS and HadSST3 also account for
403 biases and uncertainties differently. HadSST3 does not apply smoothing or interpolation
404 between neighboring grid cells, facilitating more direct regional comparisons.

405

406 We examine the monthly HadSST3 anomalies, version 3.1.1.0 (median realization). We area-
407 average the HadSST3 grid boxes into similar sub-regions similar to those shown in **Fig. 3a**.
408 The southwest Labrador Sea BINS grid box is also examined, which closely corresponds to
409 one HadSST3 grid box. Generally however, the large 5°×5° HadSST3 grid boxes do not align
410 directly with BINS grid boxes; the specific grid boxes are shown in **Fig. C1**. We average the
411 HadSST3 anomalies into 3-month seasons (skipping over missing values). The seasonal
412 anomalies are 1-2-1 smoothed (excluding endpoints) over successive years, and the seasonal
413 anomalies are averaged into Dec–Nov annual means (again skipping over missing values).
414 The winter season is not included in the southern Nordic, central Labrador, or southwest
415 Labrador annual means before 1947 to match the BINS coverage. Likewise, the fall season is
416 excluded from the central Labrador and southwest Labrador sub-regions. In general, we find
417 that the BINS data are more seasonally coherent at low frequencies before WWII (not
418 shown).

419

420 The HadSST3 and BINS regional time series are compared in **Fig. 6**, with the respective
421 1896-2015 and 1950-2015 correlations listed in **Table 6**. Overall, the products are very highly
422 correlated since 1950. In the first half of the record, the largest differences between the
423 products are generally found during periods of data gaps and linear interpolation near the end

424 of WWI and WWII, when BINS error estimates are also large. Among regions, the agreement
425 is less good in the southern Nordic seas (**Fig. 6b**) before the 1940s; we also find that the
426 HadSST3 SST in different seasons are also much less correlated there before the 1940s (not
427 shown). To some extent, the early differences come from the most poorly sampled
428 northwestern part of this region, but also from Danish and Norwegian data incorporated in
429 this product and not in HadSST3. Furthermore, in the 1930s, the differences might also
430 originate from Norwegian data interpreted in BINS as recorded continuously on a water line
431 inside the ship, and thus attributed a large positive bias. The central Labrador Sea / West
432 Greenland shelf (**Fig. 6a**) also shows less agreement before 1930. The products correspond
433 very well after the 1930s, although HadSST3 has less of a cooling in the 1980s. The two SPG
434 regions are both very strongly correlated over the length of the record (**Fig. 6c,d**). The
435 Southwest Labrador Sea grid box (**Fig. 6e**) generally corresponds well after the gap periods in
436 the early 1900s. In the regions further south (**Fig. 6f,g,h**), the two time series usually show a
437 good agreement, and are most of the time compatible with the BINS error estimate. The
438 largest differences are near the Grand Banks (**Fig. 6g**) before 1920, maybe as a result of
439 different seasonal coverage, but also in the 1990s, when BINS tends to present lower values.
440 Large differences are also found in the intergyre region (**Fig. 6f**) in the early 1950s. This is
441 the only place where the error estimate does not explain the difference between the curves,
442 maybe because, for a few years in the early 1950s, data in BINS for some of these boxes were
443 only available in the summer season, and thus are less representative of annual averages.

444

445 **4. Characteristics of T, S and surface density RMS variability**

446 In this section, we examine the linear trends and the RMS variability of the linearly detrended
447 interannual time series. The 1896–2015 BINS least-squares linear trends are shown in **Fig.**
448 **7a–7c**. We construct error estimates for the linear trends by block resampling using the time
449 series error estimates (resampled every 2 years to account for the 1-2-1 filter). The SSS trend
450 (**Fig. 7a**) shows a negative trend throughout most of the domain, except for the southwest and
451 part of the eastern boundary; the pattern is consistent with the 1896–2013 trend shown in
452 Friedman et al. (2017). The SST trend (**Fig. 7b**) shows strong warming in the Gulf Stream,
453 and to a lesser extent along the eastern margin. There is cooling in West Greenland / Central
454 Labrador, and near 52–55°N/30°W. However, error bars due to higher frequency variability or
455 sampling are often large. Both the increase in T and the decrease in S contribute to an overall
456 decrease in surface density (**Fig. 7c**).

457

458 The 1896–2015 HadSST3 trend is shown in (**Fig. 7d**). The trends are calculated for grid
459 boxes with 5 or fewer years with missing data. [Note that specific seasons are not removed
460 from the spatial grid boxes as for the HadSST3 indices discussed above]. Error terms for
461 HadSST3 are computed by multiplying the white-noise standard error of the slope by a factor
462 of $\sqrt{2}$ on account of the 1-2-1 filter. Like BINS, the HadSST3 trend shows significant
463 warming in the Gulf Stream extension, and also along the eastern margin. HadSST3 also
464 shows a region of cooling in the central SPG from 50-60°N, referred to as the ‘warming hole’
465 (Drijfhout et al., 2012), which is not well reproduced in BINS. Checking this particular
466 difference suggests that data in the pre-1917 period originated mostly from one group of data
467 from the ICES archive, often reported at set longitudes, and for which we applied an overall
468 correction of -0.2°C on temperature, but with little comparison data to define it (Reverdin et
469 al. 1994). Notice however that nonetheless, for the area average of the central SPG (**Fig. 6c**),
470 the difference between BINS and HADSST3 is not significant for that period.

471

472 The detrended RMS standard deviation of the interannual T and S time series (**Figs. 8a** and
473 **8b**) correspond to the expectation that large-scale temperature RMS standard deviation
474 diminishes from west to east; whereas the salinity RMS values are largest in the west, then
475 near 50°N with a decrease from west to east and from the southern to the northern SPG. This
476 pattern is indicative on one hand of larger air-sea and hydrographical/heat forcings in the
477 western SPG, and on the other hand of the presence during some winters of a rather shallow
478 mixed layer in the western SPG (even in the Labrador Sea, where deep convection also often
479 takes place (Yashayaev and Loder 2016)). In mid-ocean near 45-50°N, there might also be a
480 contribution of displacements of hydrographical fronts.

481

482 Surface density standard deviation (**Fig. 8c**) diminishes both from west to east and south to
483 north. The relative contributions of T and S to detrended density variability are shown in **Fig.**
484 **8d**. The contributions of SST and SSS to surface density variability are opposite in most
485 places (the effects of these variables on density variability tend to compensate each other, as
486 there is a positive correlation between SST and SSS variability on intra-annual to centennial
487 time scale in a large part of this domain). This also might explain a pattern of density rms
488 variability which presents weaker zonal gradients in the west than what is found in SST or
489 SSS.

490

491 Interestingly, in the westernmost part of the domain, except near the Gulf Stream, the salinity
492 contribution to density variability dominates over the temperature contribution (**Fig. 8d**),
493 whereas in the eastern and northern parts of the SPG, temperature dominates. Thus in the first
494 region, density tends to be positively correlated with SSS and SST, whereas in the other
495 region it is negatively correlated with SSS and SST. In the northern part, this was not
496 expected, as temperature is lower than further south, which would contribute to diminish the
497 contribution of temperature variations to density variations. However, this is counteracted by
498 a lower ratio of SSS over SST variability in the northern regions. In general, the SSS
499 contribution is slightly larger for the non-detrended data (not shown), due to its relatively
500 larger centennial trends in this part of the ocean (discussed in part 2 of this study).

501

502 Next, we show the total RMS variability contained at multidecadal frequencies. We low-pass
503 filter the time series using a spectral filter with half power at a period close to 15 years. **Fig. 9**
504 shows the percentage of the detrended interannual variance explained by the low-pass filtered
505 time series. The low-pass filtered variance percentage of SSS (**Fig. 9a**) and SST (**Fig. 9b**) are
506 less correlated spatially than overall RMS variability. For S, what striking is the lower
507 percentages in the western part (as low as 40%) than in the eastern part of the domain (larger
508 than 60%), whereas for T, it is mostly the northern part which has a larger percentage of
509 variance in the multidecadal frequencies, with percentage of detrended variance reaching up
510 to 80%. The low percentages of multidecadal variability for SSS in the northwest is
511 consistent with the penetration in the ocean interior of short-lived freshwater pulses
512 originating from the western shelves and the rim of the Labrador Sea (examples of recent
513 short-term SSS variability in this region are presented in Tesdal et al. (2018)).

514

515 The low-pass filtered time series explain comparatively less density variance, particularly in
516 the western SPG region (**Fig. 9c**, values usually less than 50%), which suggests coordinated
517 SSS and SST variations at lower frequencies with thus partial compensation on their
518 contribution to density changes. Indeed, the low-pass filtered correlation of SSS and SST
519 (**Fig. 9d**) is positive in most grid boxes, with large magnitudes in the central SPG, around
520 Greenland and in the south-west north of the Gulf Stream as well as around the Grand Banks.
521 Notice however, less significant correlation in other areas of the western, southern, and
522 eastern parts of the domain close to Europe. The relationship between SSS and SST variations
523 will be examined further in part 2 of this study.

524

525 **5. Discussion and Conclusions**

526 We have constructed a spatially averaged binned product BINS of interannual T, S, and
527 surface density for 1896-2015 north of 40°N, similar to what was done for S in Friedman et
528 al. (2017), but at a higher spatial resolution. We deliberately excluded the shelf regions
529 (except for one box along southwestern Greenland and one box on the southern Grand
530 Banks), which have different dynamics, variability and sampling issues (Reverdin et al.
531 2018). We have shown that in recent decades, the BINS interannual variability of SST and
532 SSS is coherent with published time series of upper ocean heat and salinity based on high
533 quality hydrographic data. It provides also rather similar results to an analysis of SST and
534 SSS variability along two ship tracks since 1993, based on the same in situ data set for its
535 main part (Reverdin et al. 2018).

536
537 Furthermore, BINS compares well with two gridded products of hydrographic data (EN4
538 from 1950-2015 and ISHII from 1950-2012), which use different data selection and mapping
539 methods. This suggests that in this region, the processing and spatial averaging of the
540 individual data in the spatial bins done in BINS is usually sufficient to portray the interannual
541 variability of SST on the grid box scale, at least since 1950, and that during this period, the
542 data set used does not contain large biases compared to these other data sets. On interannual
543 time scales, BINS SST also fits largely with what is portrayed in HadSST3, except in the pre-
544 WWII period for the Labrador Sea and the Norwegian Sea. Insufficient coverage or erroneous
545 data, either in HadSST3 or BINS, might be causing these discrepancies. Thus, more care
546 should be taken when interpreting variability in BINS before 1950.

547
548 The product combines time series of interannual variability in different seasons. This was
549 required to reduce sampling errors to a reasonable level (in particular pre-1950), but this
550 might also mask or alias important differences between the seasons. For example in the
551 northern part of the subtropical gyre (thus, south of the region we investigate) a strong
552 seasonal modulation in the recent trends towards increasing SSS has been documented,
553 associated to important seasonal changes of the hydrological cycle (Yu et al. 2017). In the
554 Labrador Sea, too, on an interannual time scale, large summer-time anomalies of SSS have
555 been documented in the recent decade (2008 and 2012) that don't have a clear counterpart in
556 other seasons (Tesdal et al. 2018). Such events, and also the seasonality of atmospheric
557 forcing, seasonal thermal stratification and mixing, could induce large interannual differences
558 between the anomalies in different seasons. Indeed, **Fig. 3** suggests that the differences

559 between seasons are large in the central Labrador Sea, but for the other open-ocean regions,
560 different seasons present more similar decadal and longer frequencies. Possibly, the
561 insufficient accuracy of binned time series and the change in collection and measurement
562 techniques could preclude further investigation of the seasonality of the decadal or longer
563 variability with this dataset. For example, in boxes of the southern SPG, we noticed SSS
564 differences in the autumn around 1960 that could be a data artifact. Differences with
565 HadSST3 taking place around 2000 could also be data-related (either in HadSST3 with
566 drifting buoy data becoming more numerous, or in BINS with the Argo data afterwards).
567 Nonetheless, a more homogeneous 25-year dataset in the central SPG resolving seasonal
568 variability along well sampled ship lines (Reverdin et al. 2018) also pointed out a similarity in
569 interannual variability for the different seasons, although interannual RMS variability tended
570 to be larger there too in early summer for SST.

571

572 This aside, the seasonal inhomogeneity in data coverage, with less sampling in winter in
573 particular before 1955, should have little impact on the reconstructed interannual time series.
574 We however warn that this might not be a wise substitute for the seasonal time series, in
575 particular for studies of winter conditions the central Labrador Sea or water mass formation,
576 but then obviously the available binned data present more gaps, and the original seasonal time
577 series have larger error bars.

578

579 Density is a non-linear function of temperature and salinity, with the character of non-
580 linearity increasing at low T. We deliberately emphasized winter characteristics when
581 estimating density from temperature and salinity time series, but use the BINS time series
582 which combine SST and SSS anomalies in the four (or three) different seasons. We noticed
583 that in some regions interannual variability has smaller amplitude in winter (Table 3
584 summarizing **Fig. 3**), thus this choice might result in an overestimation of winter-time
585 interannual density variability. It would be also misleading to use these density interannual
586 time series as a proxy for water density in other seasons.

587

588 We do not have a direct comparison of SSS BINS time series prior to the 1950s. We estimate
589 that they provide reasonable estimates for this early period mostly based on comparisons of
590 the SST BINS time series with HadSST3 binned data. However, before 1950 in the Nordic
591 Seas and Labrador Sea, the binned data have larger uncertainties and winter seasons are
592 poorly (or not at all) sampled. Thus, clearly our estimates for SSS are also less certain in these

593 regions prior to 1950. A possibility for further validation would be to compare with
594 independent proxy-based estimates of surface variability. Few such proxy time series have
595 been derived in this region resolving the multidecadal variability in SST and even less in
596 SSS, and their uncertainties are probably too large for such a validation (Hall et al.
597 (2010), for a site in the Iceland Basin (Gardar drift); Richter et al. (2009), for a site in the
598 northern Rockall Trough area (Feni drift); Moffa-Sanchez et al. (2017) and references
599 therein as well as Thornalley et al. (2018), for a site in the southeastern Labrador Sea).

600
601 BINS has sufficient spatial resolution to resolve large-scale patterns of RMS standard
602 deviation of the variability in SST, SSS and density (this was done after removing linear
603 trends). RMS variability in SSS and to a lesser degree in SST increases towards the west, with
604 the contribution to density of SSS variability dominating that of SST in the west, whereas in
605 the eastern and northern parts of the SPG, temperature typically dominates the variability. The
606 portion of variability in multidecadal frequencies (here with a cut-off at 15 years) could be
607 ascertained. It is found both for SST and SSS to be a very large part of interannual variability,
608 with lesser relative contribution in the west for SSS and in the southern part for SST. For
609 density, it is less prominent, except in the northeastern part of the SPG, where SST variability
610 contribution dominates and is very much at multidecadal frequencies (also see Holliday et al.
611 2015). The higher frequencies are more influenced by the sampling uncertainties, in particular
612 the effect of interpolation and interannual smoothing during data gaps in either specific
613 seasons or all seasons (near WWII, for example). Thus this analysis of the relative importance
614 of multidecadal variability is more dataset-dependent than the RMS variability

615

616

617 **Acknowledgements**

618 This is a contribution to the French SSS observation service, which is supported by French
619 agencies INSU/CNRS, IRD, CNES and IPEV, as well as from SOERE CTDO2. The Rockall
620 Trough time series were provided with support from the UK Natural Environment Research
621 Council (Extended Ellett Line program, National Capability). The station time series south of
622 Iceland were provided with Icelandic support. The annual oceanographic monitoring of the
623 Labrador Sea was initiated as a Canadian contribution to the World Ocean Circulation
624 Experiment in 1990 and is presently conducted as a core component of the Atlantic Zone Off-
625 Shelf Monitoring Program (AZOMP) run by the Bedford Institute of Oceanography of
626 Fisheries and Oceans Canada. The International Argo Program is part of the Global Ocean

627 Observing System (Argo, 2000). Argo data are available from the Coriolis Global Data
628 Center, Institut français de recherche pour l'exploitation de la mer (Ifremer). The HadSST3
629 and EN4 data were provided by the Met Office Hadley Center, and the ISHII data were
630 provided by the NCAR Research Data Archive. A.R.F. was supported by SOERE CTDO2
631 and the ERC funded project TITAN (EC-320691). L.C. acknowledges support from the
632 Swedish National Space Board (SNSB; Dnr 133/17). Comments by two reviewers and by
633 Semyon Grodsky were appreciated. The BINS product is available at
634 <https://dx.doi.org/10.6096/TSD-BINS-NASPG>.

635

636 **References**

- 637 Alory G, Delcroix T, Téchiné P, Diverrès D, Varillon D, Cravatte S, ... Roubaud F (2015)
638 The French contribution to the voluntary observing ships network of sea surface salinity.
639 Deep Sea Res I 105: 1–18. <http://doi.org/10.1016/j.dsr.2015.08.005>
- 640 Argo (2000) Argo float data and metadata from Global Data Assembly Centre (Argo GDAC).
641 SEANOE. <http://doi.org/10.17882/42182>
- 642 Belkin I M, Levitus S, Antonov J, Malmberg S.-A (1998) “Great Salinity Anomalies” in the
643 North Atlantic. Prog Oceanogr 41(1): 1–68. [http://doi.org/10.1016/S0079-6611\(98\)00015-](http://doi.org/10.1016/S0079-6611(98)00015-9)
644 9
- 645 Böning C W, Behrens E, Biastoch A, Getzlaff K, Bamber J L (2016) Emerging Impact of
646 Greenland Meltwater on Deepwater Formation in the North Atlantic Ocean. Nature
647 Geoscience. <https://doi.org/10.1038/ngeo2740>.
- 648 Boutin J, Chao T, Asher W, Delcroix T, Drucker R, Drushka K, Kolodziejczyk N, Lee T,
649 Reul N, Reverdin G, Schanze J, Soloviev A, Yu L, Anderson J, Brucker L, Dinnat E,
650 Santos-Garcia A, Jones W, Maes C, Meissner T, Tang W, Vinogradova N, Ward B (2016)
651 Satellite and In Situ Salinity : Understanding Near-Surface Stratification and Sub-footprint
652 Variability. Bull. Amer. Meteor. Soc., 97, 1391–1407, doi: 10.1175/BAMS-D-15-
653 00032.1.
- 654 Buckley M W, Marshall J (2016) Observations, Inferences, and Mechanisms of the Atlantic
655 Meridional Overturning Circulation: A Review. Reviews of Geophysics 54 (1):
656 2015RG000493. <https://doi.org/10.1002/2015RG000493>.
- 657 Cabanes C, Grouazel A, von Schuckmann K, Hamon M, Turpin V, Coatanoan C, Paris F, et al
658 (2013) The CORA Dataset: Validation and Diagnostics of in-Situ Ocean Temperature and
659 Salinity Measurements. Ocean Sci 9 (1): 1–18. <https://doi.org/10.5194/os-9-1-2013>.
- 660 Davis R E, Sherman J T, Dufour J (2001) Profiling ALACEs and other advances in
661 autonomous subsurface floats. J Atmos Ocean Tech 18: 982-993.
- 662 Dickson R R, Meincke J, Malmberg M S-A, Lee A J (1988) The “Great Salinity Anomaly”
663 in the northern North Atlantic 1968 – 1982. Prog Oceanogr 20: 103– 151.
- 664 Drijfhout S, van Oldenborgh G J, Cimadoribus A (2012) Is a Decline of AMOC Causing the
665 Warming Hole above the North Atlantic in observed and modeled warming patterns?” J
666 Clim 25: 8373–79. <https://doi.org/10.1175/JCLI-D-12-00490.1>.
- 667 Ebisuzaki W (1997) A method to estimate the statistical significance of a correlation when the
668 data are serially correlated. J Clim 10(9): 2147–2153.
- 669 Fofonoff N P (1985) Physical properties of seawater: A new salinity scale and equation of
670 state for seawater. J Geophys Res 90(C2): 3332–3342.
671 <https://doi.org/10.1029/JC090iC02p03332>.

672 Frankignoul C, Deshayes J, Curry R (2009) The Role of Salinity in the Decadal Variability of
673 the North Atlantic Meridional Overturning Circulation. *Climate Dynamics* 33(6): 777–93.
674 <https://doi.org/10.1007/s00382-008-0523-2>.

675 Friedman A R, Reverdin G, Khodri M, Gastineau G (2017) A new record of Atlantic sea surface
676 salinity from 1896 to 2013 reveals the signatures of climate variability and long-term trends.
677 *Geophys Res Lett* 44: 1866–1876. DOI:10.1002/2017GL072582.

678 Good S-A, Martin M J, Rayner N A (2013) EN4: Quality Controlled Ocean Temperature and
679 Salinity Profiles and Monthly Objective Analyses with Uncertainty Estimates. *J Geophys*
680 *Res* 118(12): 6704–16. <https://doi.org/10.1002/2013JC009067>.

681 Hall I R, Boessenkool K P, Barker S, McCave I N, Elderfield H (2010) Surface and deep
682 ocean coupling in the subpolar North Atlantic during the last 230 years. *Paleoceanogr* 25.
683 doi:10.1029/2009PA001886.

684 Holliday N P, Cunningham S A, Johnson C, Gary S F, Griffiths C, Read J F, Sherwin T
685 (2015) Multidecadal variability of potential temperature, salinity, and transport in the
686 eastern subpolar North Atlantic. *J Geophys Res* 120. doi:10.1002/2015JC010762.

687 Hughes, S. L., Holliday, N. P., Gaillard, F., and the ICES Working Group on Oceanic
688 Hydrography, 2012. Variability in the ICES/NAFO region between 1950 and 2009:
689 observations from the ICES Report on Ocean Climate. – ICES Journal of Marine
690 Science, doi:10.1093/icesjms/fss044.Icelandic hydrographic surveys ...

691 Ishii M, Kimoto M, Sakamoto K, Iwasaki S I (2006) Steric sea level changes estimated
692 from historical ocean subsurface temperature and salinity analyses. *J Oceanography*
693 62(2): 155-170.

694 Kennedy J J, Rayner N A, Smith R O, Parker D E, Saunby M (2011) Reassessing Biases and
695 Other Uncertainties in Sea Surface Temperature Observations Measured in Situ since
696 1850: 1. Measurement and Sampling Uncertainties. *J Geophys Res* 116.
697 doi:10.1029/2010JD015218.

698 Kennedy J J, Rayner N A, Smith R O, Saunby M, Parker D E (2011) Reassessing biases and
699 other uncertainties in sea-surface temperature observations since 1850: 2. Biases and
700 homogenisation. *J Geophys Res* 116 D14104. doi:10.1029/2010JD015220

701 Lavender K L, Davis R E, Owens W B (2000) Mid-depth recirculation observed on the
702 interior Labrador and Irminger seas by direct velocity measurements. *Nature* 607: 66- 69.

703 Lozier M S (2012) Overturning in the North Atlantic. *Ann Rev of Marine Sci* 4(1): 291–315.
704 <http://doi.org/10.1146/annurev-marine-120710-100740>

705 Moffa-Sanchez P, Hall I R (2017) North Atlantic variability and its links to European climate
706 over the last 3000 years. *Nature comm.* 8:1726, 555. doi:10.1038/s41467-017-01884-8.

707 Piron A, Thierry V, Mercier H, Caniaux G (2017) Gyre-scale deep convection in the subpolar
708 North Atlantic Ocean during winter 2014–2015. *Geophys Res Lett* 44: 1439–1447.
709 doi:10.1002/2016GL071895.

710 Polyakov I V, Bhatt U S, Simmons H L, Walsh D, Walsh J E, Zhang X (2005) Multidecadal
711 Variability of North Atlantic Temperature and Salinity during the Twentieth Century. *J*
712 *Clim* 18(21): 4562–4581. <http://doi.org/10.1175/JCLI3548.1>

713 Rahmstorf S, Box J E, Feulner G, Mann M E, Robinson A, Rutherford S, Schaffernicht E J
714 (2015) Exceptional twentieth-century slowdown in Atlantic Ocean overturning
715 circulation. *Nature Climate Change* 5(5): 475–480. <http://doi.org/10.1038/nclimate2554>

716 Reverdin G (2010) North Atlantic Subpolar Gyre Surface Variability (1895–2009). *J Clim*
717 23(17): 4571–4584. <http://doi.org/10.1175/2010JCLI3493.1>

718 Reverdin G, Cayan D, Dooley H, Ellett D, Levitus S, du Penhoat Y, Dessier A (1994) Surface
719 Salinity of the North-Atlantic - Can We Reconstruct Its Fluctuations Over the Last 100
720 Years. *Prog Oceanogr* 33(4): 303–346. [http://doi.org/10.1016/0079-6611\(94\)90021-3](http://doi.org/10.1016/0079-6611(94)90021-3)

721 Reverdin G, Cayan D, Kushnir Y (1997) Decadal variability of hydrography in the upper
722 northern North Atlantic 1948-1990. *J Geophys Res* 102: 8505-8531.

723 Reverdin G, Alory G, Diverres D, Bringas F, Goni G, Heilmann L, Chafik L, Szekely T,
724 Friedman A R (2018) North Atlantic subpolar gyre along predetermined ship tracks since
725 1993: a monthly data set of surface temperature, salinity, and density. *Earth Syst Sci Data*
726 10: 1403-1415. <https://doi.org/10.5194/essd-10-1403-2018>.

727 Rhein M, Kieke D, Hüttl-Kabus S, Roessler A, Mertens C, Meissner R, Klein B, Böning C W,
728 Yashayaev I (2011) Deep water formation, the subpolar gyre, and the meridional
729 overturning circulation in the subpolar North Atlantic. *Deep Sea Res II* 58: 1819-1832.
730 doi:10.1016/j.dsr2.2010.10.061

731 Richter T O., Peeters F J C, Weering T C E (2009) Late Holocene (0-2.4 ka BP) surface water
732 temperature and salinity variability, Feni Drift, NE Atlantic Ocean. *Quat Sci Rev* 28:
733 1941-1955.

734 Riser S C, Freeland H J, Roemmich D, Wijffels S, Troisi A, Belbéoch M, Gilbert D, et al.
735 (2016) Fifteen Years of Ocean Observations with the Global Argo Array. *Nature Climate*
736 *Change* 6 (2): 145–53. <https://doi.org/10.1038/nclimate2872>.

737 Roemmich, D, Johnson G C, Riser S, Davis R E, Gilson J, Owens W B, Garzoli S L, Schmid
738 C, Ignaszewski M (2009) [The Argo Program: Observing the global ocean with profiling
739 floats](https://doi.org/10.5670/oceanog.2009.36). *Oceanography*. 22:34-43. [10.5670/oceanog.2009.36](https://doi.org/10.5670/oceanog.2009.36)

740 Skliris N, Marsh R, Josey S A, Good S A, Liu C, Allan R P (2014) Salinity changes in the
741 World Ocean since 1950 in relation to changing surface freshwater fluxes. *Climate*
742 *Dynamics* 43(3-4): 709–736. <http://doi.org/10.1007/s00382-014-2131-7>

743 Tesdal J, Abernathey R P, Goes J I, Gordon A L, Haine T W (2018) Salinity Trends within
744 the Upper Layers of the Subpolar North Atlantic. *J Clim* 31: 2675–2698.
745 <https://doi.org/10.1175/JCLI-D-17-0532.1>

746 Thornalley D J R, Oppo D W, Ortega P, Robson J I, Brierley C M, Davis R, Hall I R, Moffa-
747 Sanchez P, Rose N L, Spooner P T, Yashayaev I, Keigwin L D (2018) Anomalously weak
748 Labrador Sea convection and Atlantic overturning during the past 150 years. *Nature* 556.
749 <https://doi.org/10.1038/s41586-018-0007-4>.

750 UNESCO (1981) The Practical Salinity Scale 1978 and the International Equation of State of
751 Seawater 1980. UNESCO technical papers in marine science 36: 25pp.

752 Williams R G, Roussenov V, Lozier M S, Smith D (2015) Mechanisms of heat content and
753 thermocline change in the subtropical and subpolar North Atlantic. *J Clim* 28: 9803-9813.
754 doi:10.1175/JCLI-D-15-0097.1.

755 Yashayaev I (2007) Hydrographic changes in the Labrador Sea, 1960-2005. *Prog Oceanogr*
756 73: 242-276. doi:10.1016/j.pocean.2007.04.015.

757 Yashayaev I, Loder J W (2009) Enhanced production of Labrador Sea Water in 2008,
758 *Geophys Res Lett* 36 L01606. doi:10.1029/2008GL036162.

759 Yashayaev I, Loder J W (2016) Recurrent replenishment of Labrador Sea water and
760 associated decadal-scale variability. *J Geophys Res* 121:8095-8114.
761 doi:10.1002/2016JC012046.

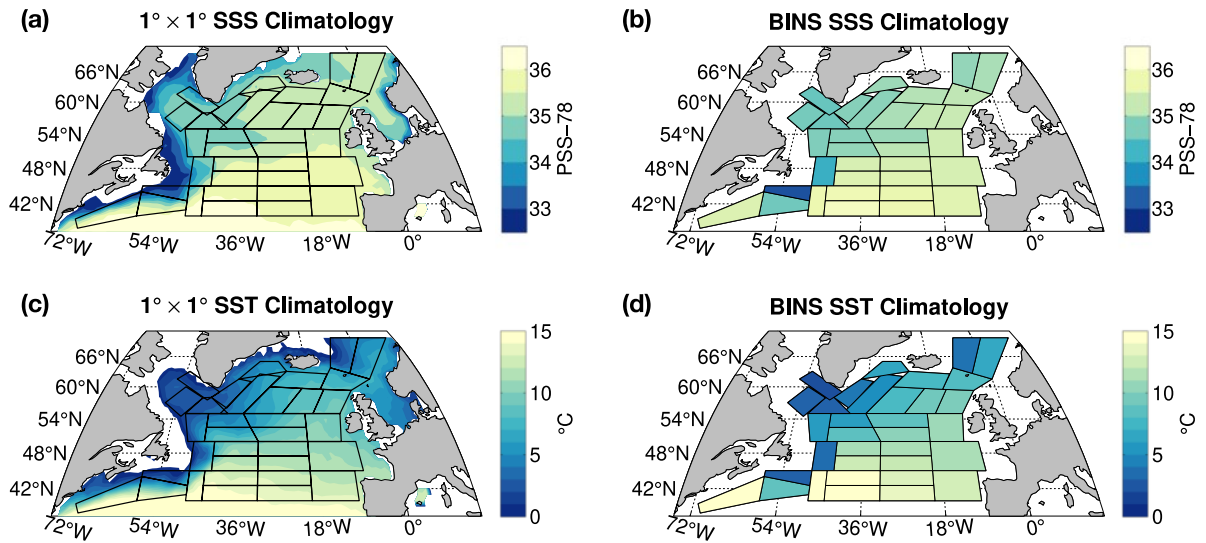
762 Yashayaev I, Loder J W (2017) Further intensification of deep convection in the Labrador
763 Sea in 2016. *Geophys Res Lett* 44:1429-1438, doi:/10.1002/2016GL071668.

764 Yashayaev I, Seidov D (2015) The role of the Atlantic water in multidecadal ocean
765 variability in the Nordic and Barents Seas. *Prog Oceanogr* 132: 68-127.

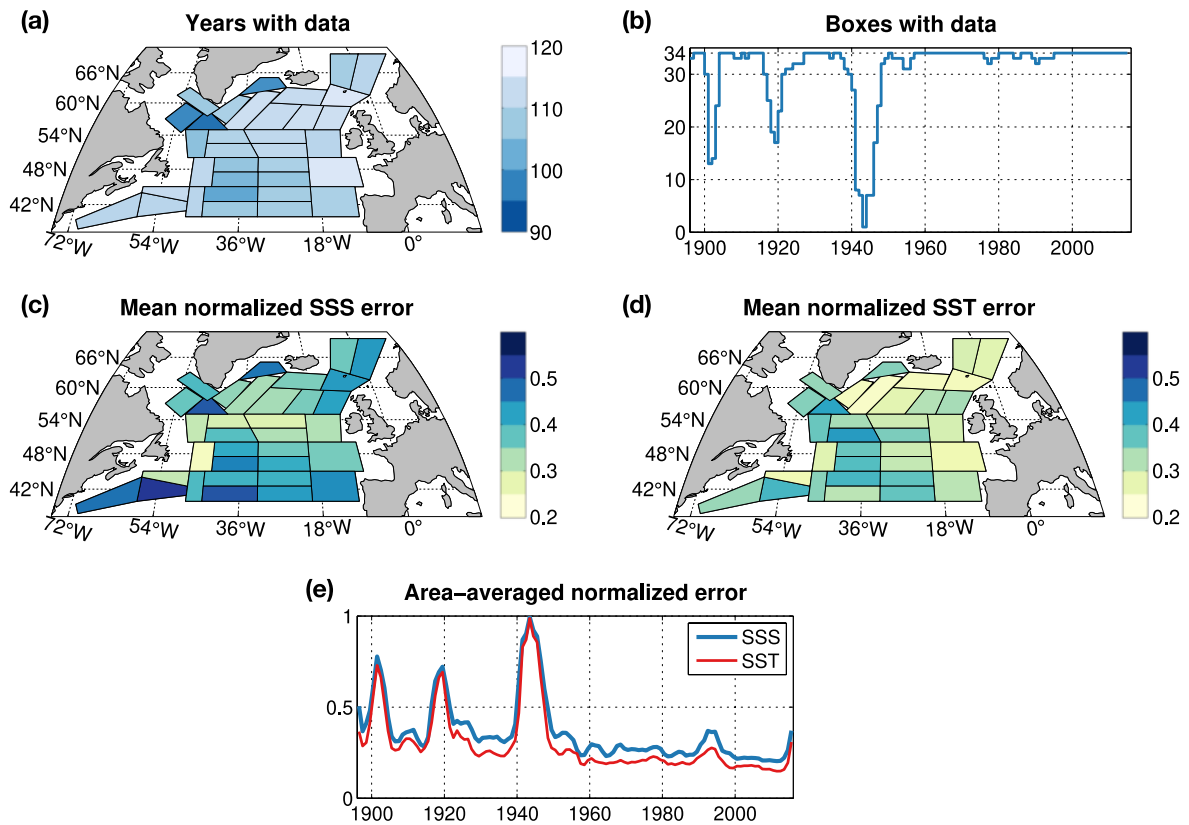
766 Yu L, Jin X Liu H (2017) Poleward Shift in Ventilation of the North Atlantic Subtropical
767 Underwater. *Geophysical Research Letters* 44. doi:
768 <https://doi.org/10.1002/2017GL075772>

769

770 **Figures.**

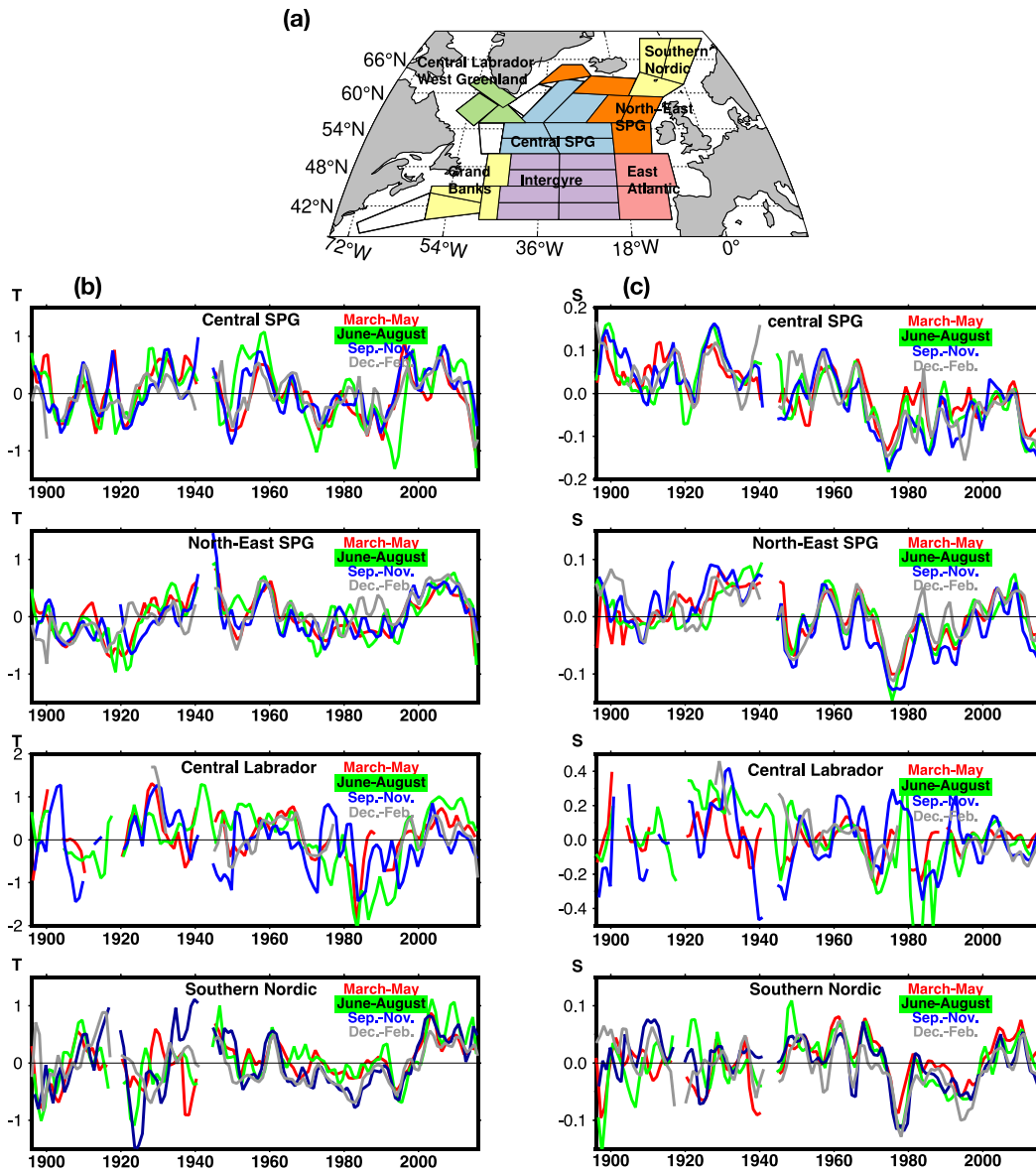


771 **Figure 1.** Gridded climatologies () (for 1896-2000). (a) and (c): 1°×1° March climatologies
772 for (a) SSS and (c) SST. (b) and (d): BINS March–May climatologies for (b) SSS and (d)
773 SST.
774
775



776
777
778
779
780
781
782
783

Figure 2. (a) Outline of BINS boxes, with number of years with data from 1896–2015 (out of a maximum of 120). (b) Time series of total BINS boxes with coverage (out of a maximum 34). The data counts in (a) and (b) are from after applying the 1-2-1 filter. (c) Mean annual SSS grid box error, defined as the mean grid box error divided by its maximum error (usually in 1943). (d) Same as (c), for SST. (e) Area-weighted normalized annual grid box error for SSS and SST.



784
785
786
787

Figure 3. (a) Regional domains described in the text. (b-c) Seasonal time series for (b) T (°C) and (c) S (PSS-78) averaged over four domains portrayed in (a): central SPG, north-east SPG, central Labrador Sea / West Greenland, and the southern Nordic Seas.

788 **Table 3.** Seasonal RMS and Pearson correlation coefficient. RMS values are in °C for T, and PSS-78 for S (×1000).
 789 Corresponding regions are shown on Fig. 3a.

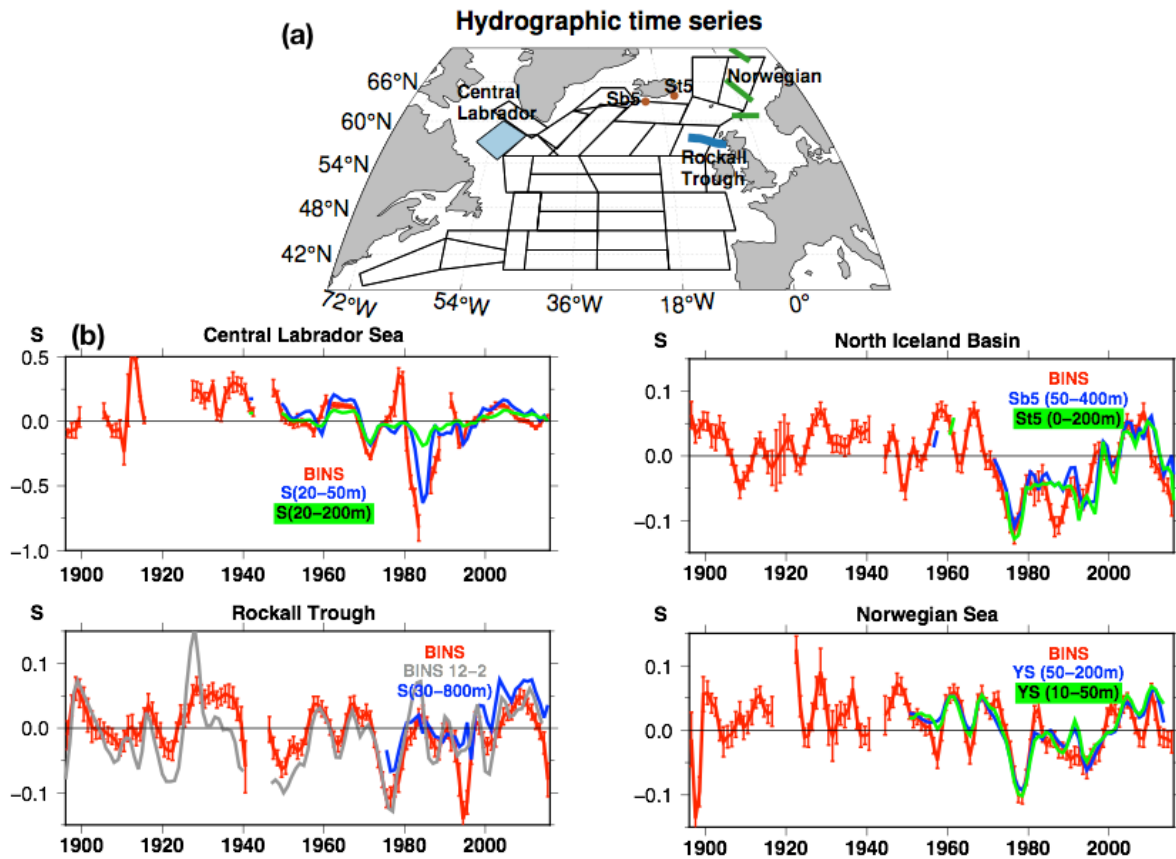
792	Region	Length	12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11
793	Central SPG	112-116 yrs	T RMS 0.304	0.407	0.505	0.404	S RMS 73	60	72	79
794			T corr.				S corr.			
795			12-2	0.58	0.52	0.57	12-2	0.72	0.77	0.78
796			3-5		0.55	0.71	3-5		0.79	0.82
797			6-8			0.62	6-8			0.86
798										
799	Region	Length	12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11
800	N-E SPG	113-115 yrs	T RMS 0.339	0.338	0.375	0.336	S RMS 42	41	44	55
801			T corr.				S corr.			
802			12-2	0.65	0.54	0.60	12-2	0.72	0.75	0.73
803			3-5		0.76	0.71	3-5		0.81	0.78
804			6-8			0.74	6-8			0.80
805										
806										
807			12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11
808	Southern Nordic		T RMS 0.358	0.351	0.425	0.524	S RMS 41	40	47	41
809			T corr.				S corr.			
810			12-2	0.31	0.40	0.61	12-2	0.50	0.43	0.66
811			3-5		0.69	0.41	3-5		0.66	0.52
812			6-8			0.57	6-8			0.61
813										
814			12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11
815	Central Labrador/ West Greenland*		T RMS 0.47	0.385	0.482	0.602	S RMS 130	95	131	162
816			T corr.				S corr.			
817			12-2	0.81	0.20	0.60	12-2	0.37	0.50	0.27
818			3-5		0.60	0.78	3-5		0.56	0.50
819			6-8			0.52	6-8			0.42

820											
821		12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11		
822	Grand Banks	T RMS	0.95	0.792	0.890	0.990	S RMS	148	156	132	150
823		T corr.					S corr.				
824		12-2	0.26	0.33	0.29	12-2	0.09	0.24	0.51		
825		3-5		0.56	0.21	3-5		0.49	0.33		
826		6-8			0.44	6-8			0.50		
827											
828		12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11		
829	Intergyre	T RMS	0.49	0.42	0.61	0.49	S RMS	87	76	82	100
830		T corr.					S corr.				
831		12-2	0.56	0.19	0.42	12-2	0.64	0.65	0.73		
832		3-5		0.44	0.53	3-5		0.51	0.55		
833		6-8			0.42	6-8			0.59		
834											
835		12-2	3-5	6-8	9-11	12-2	3-5	6-8	9-11		
836	East Atlantic	T RMS	0.37	0.37	0.41	0.51	S RMS	53	61	55	54
837		T corr.					S corr.				
838		12-2	0.59	0.41	0.57	12-2	0.40	0.53	0.56		
839		3-5		0.57	0.52	3-5		0.59	0.34		
840		6-8			0.38	6-8			0.51		

842 * small number of common points, and varying number of boxes; thus not reliable, in particular for S.

843 Both for Southern Nordic and Central Labrador, very different number of years in winter and other seasons

844 (only on the order of 60 years with winter data in some of the boxes included in the average)



845
 846 **Figure 4.** (a) Map showing hydrographic time series locations with BINS grid boxes. (b)
 847 Comparison for S of BINS (in red) with hydrographic time series low passed 1-2-1 over
 848 successive years; Central Labrador Sea (top; green and blue for different layer averages),
 849 Rockall Trough (top middle; the green curve is from BINS but for the winter (Dec-Feb.)
 850 season; blue is the vertically averaged (30-800m) S time series), North Iceland Basin (lower
 851 middle; two vertically averaged station time series are presented); and eastern Norwegian Sea
 852 (bottom; the blue and green time series is a compilation of the offshore data (median
 853 averaged) in 3 Norwegian Sea sections presented in Yashayaev and Seidov (2015), for two
 854 different depth ranges). One standard deviation rms error bars are added on the BINS series.

855 **Table 4a.** Hydrographic sections and BINS: RMS variability for S (PSS-78) and T (°C), and lag-0 correlation coefficient.

856

		RMS (S)			RMS (T)				
		corr (S)			corr (T)				
857	Region	Years S	section	BINS	Years T	section	BINS		
858	S Rockall (30–800m)	37	.040	0.048	0.73	37	0.284	0.350	0.80
859	S Iceland (50–400m)	47	.050	0.043	0.81	47	0.413	0.391	0.87
860	Central Labrador (20–50m)	69	.155	0.169	0.76	67	0.715	0.716	0.74
861	Norwegian Sea (10-50m)	64	.039	0.038	0.82	64	0.31	0.31	0.87

862

863 **Table 4b.** Hydrographic section lag correlation with BINS.

864

864	Region	Lag	Years (S)	corr. (S)	Years (T)	corr. (T)
865	South Rockall	-2	35	0.57	35	0.56
866	(30–800m)	-1	36	0.70	36	0.65
867		0	37	0.73	37	0.80
868		+1	37	0.77	37	0.80
869	(BINS leads)	+2	37	0.68	37	0.72

870

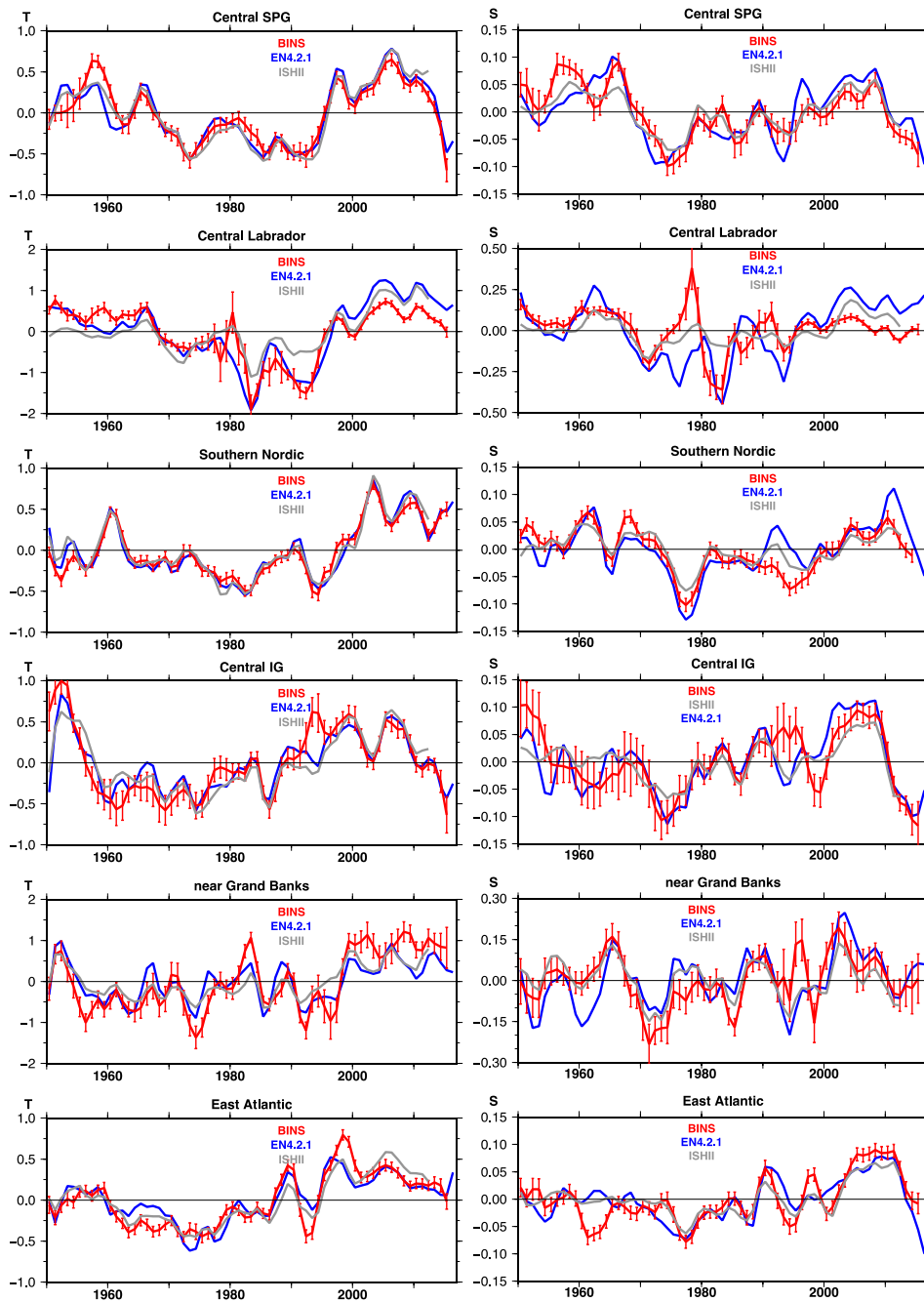
871

871	Region	Lag	Years (S)	corr. (S)	Years (T)	corr. (T)
872	South Iceland	-2	45	0.54	44	0.59
873	(50–400m)	-1	46	0.68	45	0.74
874		0	47	0.81	45	0.87
875		+1	47	0.88	45	0.89
876	(BINS leads)	+2	47	0.83	45	0.81

877

878

878	Region	Lag	Years (S)	corr. (S)	Years (T)	corr. (T)
879	Eastern Norwegian Sea	-2	64	0.48	64	0.70
880	(10-50m)	-1	64	0.72	64	0.81
881		0	64	0.82	64	0.87
882		+1	64	0.74	64	0.83
883	(BINS leads)	+2	64	0.54	64	0.72

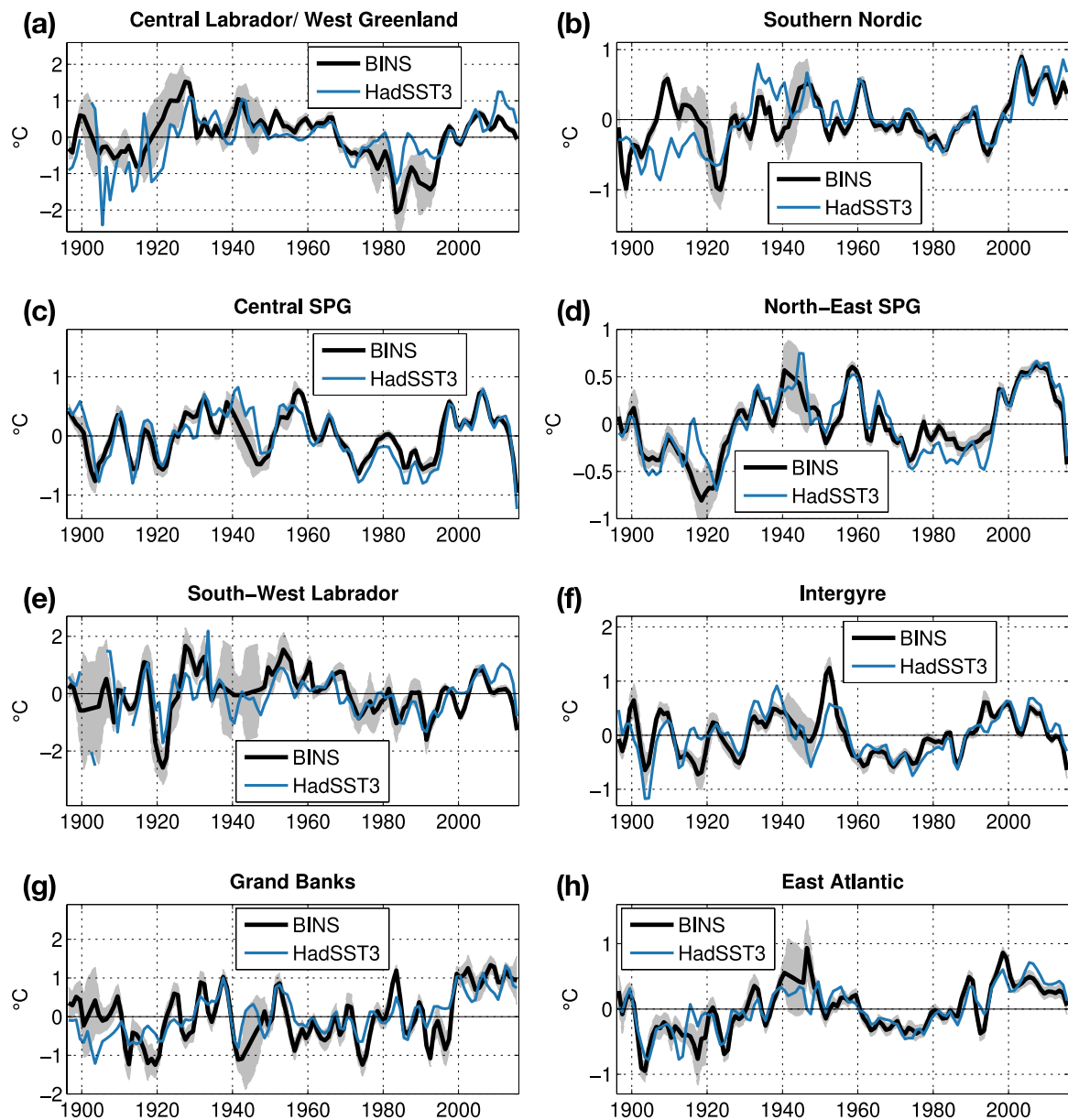


884
885
886
887
888
889

Figure 5. Comparison of regional T (left, °C) and S (right, PSS-78) time series from BINS with EN4 (1950-2015) and ISHII (1950-2012) surface products. The regional domains are the ones presented in Fig. 3. Here, the SPG region includes both the central SPG and the north-east SPG (and the south Greenland/southwest Irminger Sea box). One standard deviation rms error bars are added on the BINS series.

890 **Table 5.** RMS variability and correlation with BINS: (a) EN4, (b) ISHII.

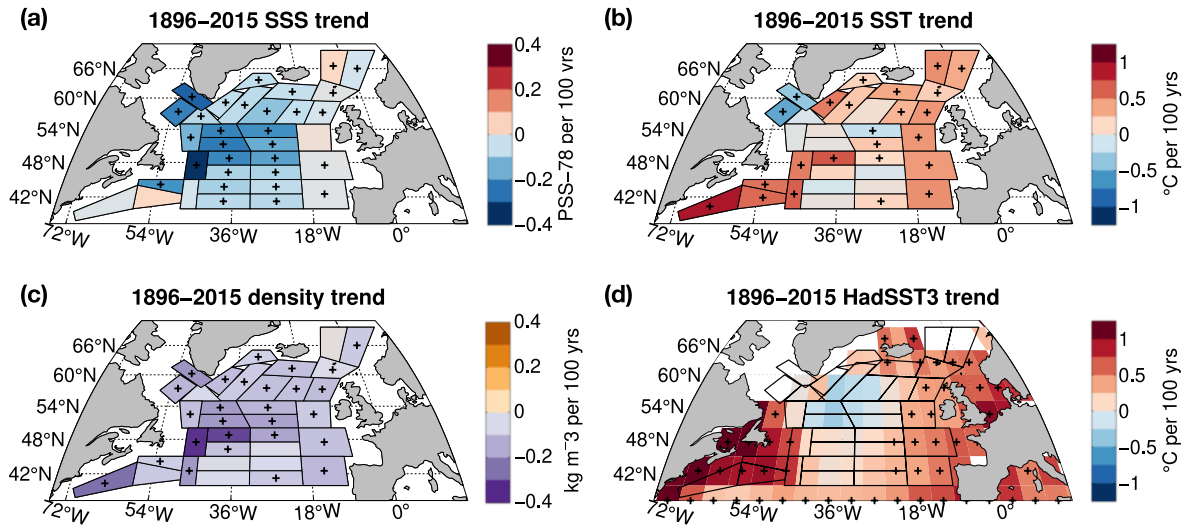
(a) EN4 / BINS		RMS (S)			corr (S)	RMS (T)		corr (T)
		(PSS-78)				°C		
Region	Time period	EN4	BINS			EN4	BINS	
894	SPG	1950–2015	0.050	0.046	0.78	0.36	0.34	0.93
895	Central Labrador/ West Greenland	1950–2015	0.173	0.121	0.49	0.76	0.63	0.87
897	Central Labrador/ West Greenland	excluding 1976–1979	0.169	0.112	0.67	0.78	0.64	0.87
899	Southern Nordic S	1950–2015	0.045	0.039	0.75	0.34	0.32	0.96
900	Intergyre	1950–2012	0.057	0.057	0.78	0.33	0.43	0.85
901	Grand Banks	1950–2012	0.098	0.114	0.40	0.47	0.69	0.42
902	East Atlantic	1950–2012	0.037	0.042	0.73	0.26	0.32	0.84
<hr/>								
(b) ISHII / BINS		RMS (S)			corr (S)	RMS (T)		corr (T)
		(PSS-78)				°C		
Region	Time period	ISHII	BINS			ISHII	BINS	
908	SPG	1950–2012	0.034	0.045	0.90	0.38	0.34	0.95
909	Central Labrador/ West Greenland	1950–2012	0.075	0.124	0.55	0.50	0.64	0.74
911	Central Labrador/ West Greenland	excluding 1976-79	0.077	0.115	0.61	0.52	0.65	0.75
913	Southern Nordic	1950–2012	0.028	0.040	0.91	0.35	0.32	0.9
914	Intergyre	1950–2012	0.034	0.045	0.71	0.34	0.34	0.62
915	Grand Banks	1950–2012	0.068	0.124	0.45	0.41	0.65	0.42
916	East Atlantic	1950–2012	0.030	0.040	0.73	0.28	0.32	0.65



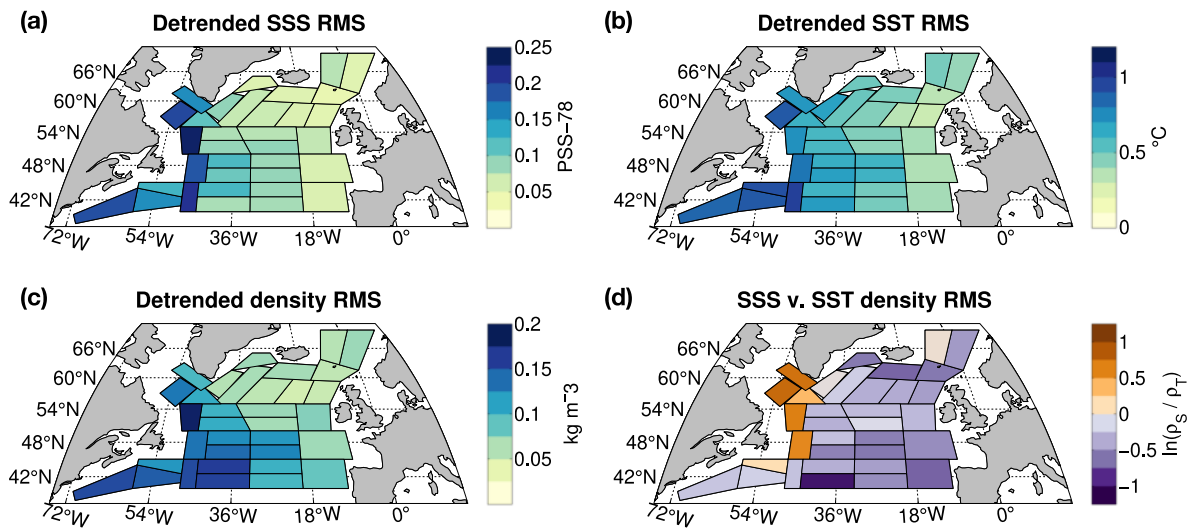
917
 918 **Figure 6.** Area-averaged BINS and HadSST3 SST anomalies: (a) central Labrador / West
 919 Greenland, (b) southern Nordic, (c) central SPG, (d) north-east SPG, (e) south-west Labrador,
 920 intergyre, (g) Grand Banks / Labrador Current, (h) east Atlantic. Shading indicates ± 2
 921 BINS error terms. The corresponding regions are shown in **Fig. C1**.
 922

923 **Table 6.** HadSST3 correlation with BINS SST

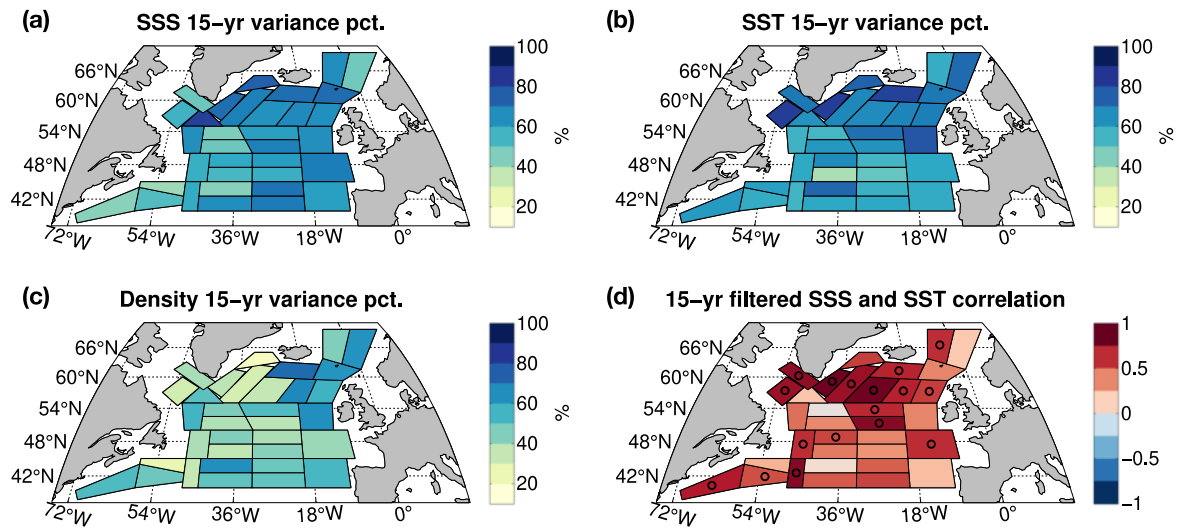
924	Region	Time period	corr	Time period	corr
925	Central SPG	1896–2015	0.84	1950–2015	0.95
926	N-E SPG	1896–2015	0.87	1950–2015	0.93
927	Central Labrador/ West Greenland	except 1900–01, 1910	0.57	1950–2015	0.72
929	SW Labrador	except 1900–01, 1904–1905, 1912	0.56	1950–2015	0.71
931	Southern Nordic	1896–2015	0.65	1950–2015	0.93
932	Intergyre	1896–2015	0.74	1950–2015	0.84
933	Grand Banks	1896–2015	0.67	1950–2015	0.82
934	East Atlantic	1896–2015	0.82	1950–2015	0.89
935					
936					



937
 938 **Figure 7.** Trend over 1896–2015 (a) SSS; (b), SST; and (c) density; per 100 years. Pluses
 939 indicate where the slope magnitude is larger than twice the estimated error. (d) HadSST3
 940 trend, per 100 years.
 941



942
 943 **Figure 8.** Detrended interannual (1-2-1 filtered) RMS variability for (a) SSS, (b) SST, and (c)
 944 density (ρ). (d) Ratio of the contributions of detrended SSS and SST to density RMS
 945 variability (plotted as a logarithm).
 946



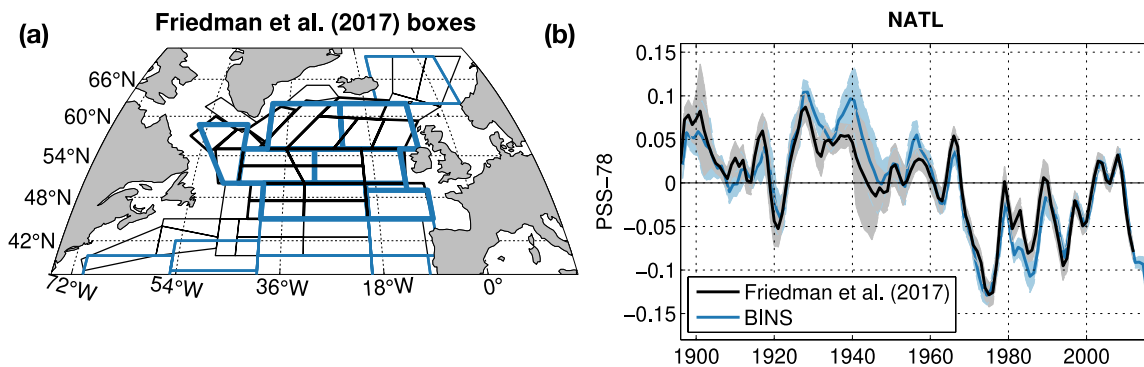
947
 948
 949
 950
 951
 952
 953

Figure 9. Percentage of detrended interannual (1-2-1) variance explained by the 15-year low-pass filtered time series for (a) SSS, (b) SST, and (c) density. (d) Detrended correlation coefficient between the low-pass filtered SSS and SST. Circles indicate where the correlation is significant at $p < 0.05$, estimated with a random-phase bootstrap test to account for serial autocorrelation (Ebisuzaki, 1997).

954 **Appendix A. Comparison with Friedman et al. (2017)**

955 The BINS boxes mainly use the same underlying SSS data as the large boxes north of 40°N in
956 Friedman et al. (2017). The datasets cover a similar area (**Fig. A1a**), though as mentioned
957 previously, the BINS boxes more carefully avoid shelf regions (except for southwest
958 Greenland and the southern part of the Grand Banks). Most of the source data used in the two
959 analyses are the same. Additionally, BINS also incorporates two small datasets from the
960 1900s and 1910s, plus a few recent transects (and 2014–2015). There are more gaps in the 19
961 BINS time series than in Friedman et al. (2017): in particular in 1918-1921, and during and
962 just after WWII; these gaps are linearly interpolated in the smaller boxes. Also, when
963 sampling is poor, but varying geographically within the larger boxes, it is possible that some
964 spatial variability is aliased in the temporal variability in the larger boxes.

965
966 **Fig. A1b** compares the NATL index from Friedman et al. (2017), area-averaged SSS from
967 45°–62°N, with SSS averaged over a similar area in BINS. [NATL from Friedman et al.
968 (2017) is only plotted through 2012, as 2013 was subject to endpoint smoothing]. The two
969 products are very highly correlated ($r=0.94$, 1896–2012), and compatible considering the
970 differences in area and error estimates. Greater differences are found for smaller regions in
971 the first half of the record, particularly during the gap years mentioned above (not shown)
972



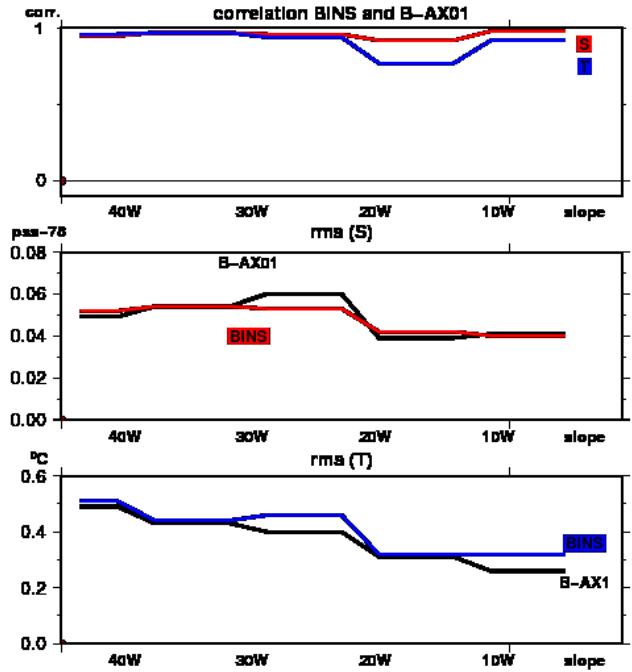
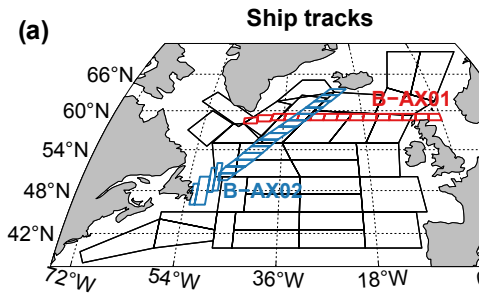
973 **Figure A1.** (a) Grid boxes from Friedman et al. (2017) (blue) and BINS (black). Thick lines
974 indicate the NATL region. (c) Comparison of NATL index from Friedman et al. (2017) and
975 BINS grid boxes. Anomalies are from 1896–2012; shading indicates ± 2 error terms.
976
977

978 **Appendix B. Comparison with Reverdin et al. (2018)**

979 As a check on the BINS time series, we compare the boxes with time series constructed with
980 mostly similar data, monthly binned along two ship routes since mid-1993, intersecting near
981 $59.5^{\circ}\text{N}/32^{\circ}\text{W}$: AX02 between Iceland and southern Newfoundland and AX01 between the
982 North Sea and southern Greenland, mostly along 59.5°N (Reverdin et al. 2018), shown in **Fig.**
983 **B1a**. Time series along AX02 start in July 1993 with few gaps, whereas for AX01 some large
984 data gaps were filled until late 1997. These time series, referred to as B-AX01 and B-AX02,
985 provide increased spatial resolution at seasonal time scales, and portray very coherent
986 variability where they intersect.

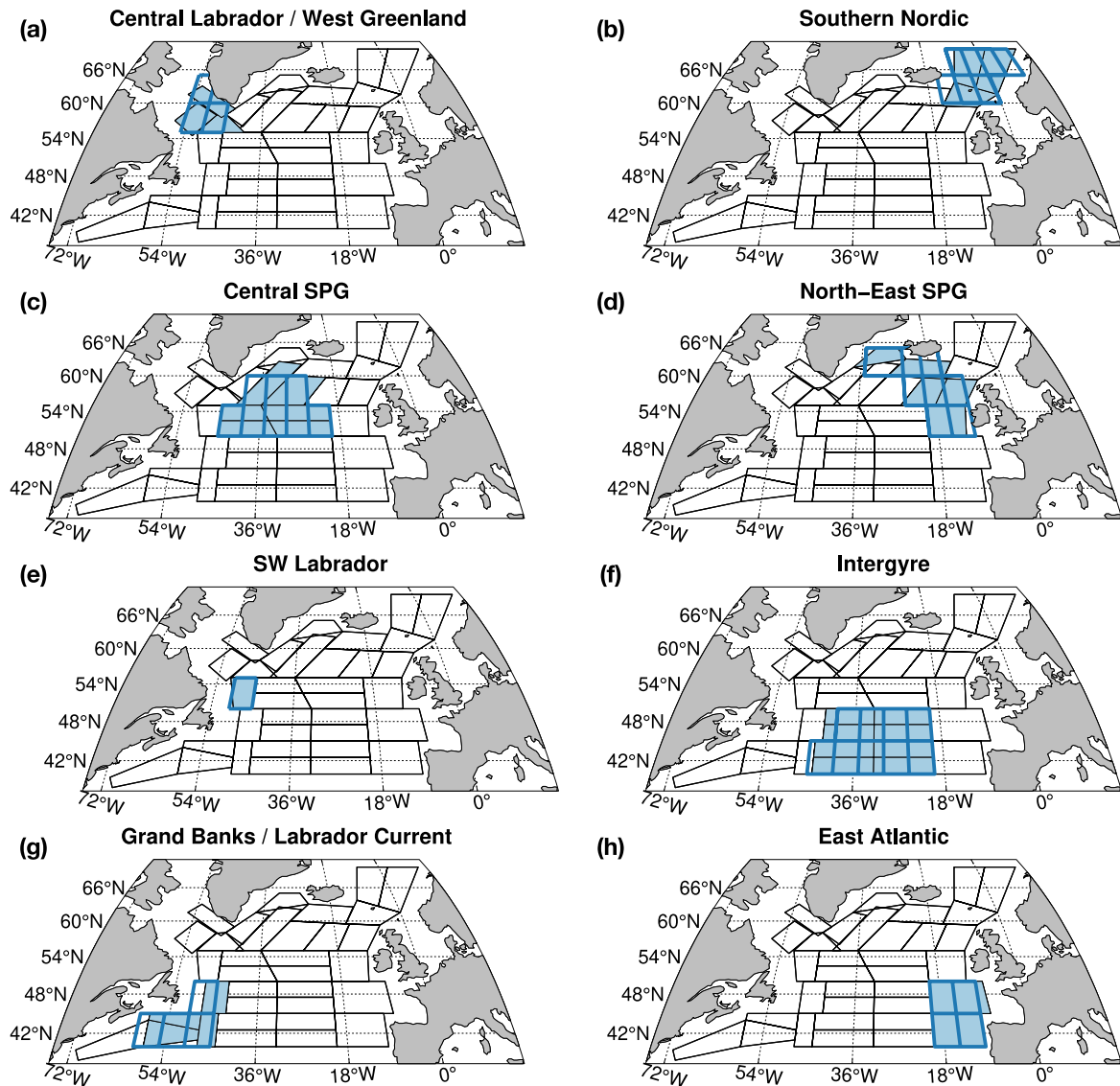
987
988 We illustrate the comparison of interannual variability of T and S between B-AX01 and BINS
989 for overlapping boxes in the common period 1993-2015 (**Fig. B1b**). For S, the corresponding
990 filtered RMS variability is larger in B-AX01 than in BINS by up to 20%, but with very high
991 correlation coefficients (all larger than 0.95). The smaller RMS amplitudes of salinity in
992 BINS probably result from the larger box sizes and the resulting spatial averaging. RMS
993 variability is more similar for T, but with a slightly smaller correlation in the Iceland Basin
994 (0.80) where gaps in 1993-1997 were the most common in B-AX01. Altogether, the
995 comparisons for the two ship tracks suggest that the method used for BINS in the box
996 averaging to produce interannual variability yields correct results when data coverage is
997 sufficient.

998



999
1000
1001
1002
1003
1004
1005

Figure B1. (a) AX01 (red) and AX02 (blue), from (Reverdin et al. 2018). (b) Comparison of B-AX01 with BINS, 1993–2015. The monthly time series of B-AX01 have been yearly-averaged (Dec-Nov), low-pass (1-2-1) filtered over successive years, and then averaged over the bins in BINS: correlation coefficient (top) for T (blue) and S (red), and RMS (middle for S and bottom for T) (B-AX01 in black, BINS in blue (T) and red (S)).



1007
 1008
 1009
 1010
 1011
 1012
 1013

Figure C1. Locations of BINS and HadSST3 grid box regions used in the Section 3.3. Thick blue lines show the $5^{\circ} \times 5^{\circ}$ HadSST3 grid boxes; shading shows the corresponding BINS grid boxes: (a) central Labrador / West Greenland, (b) southern Nordic, (c) central SPG, (d) north-east SPG, (e) south-west Labrador, (f) intergyre, (g) Grand Banks / Labrador Current, (h) east Atlantic.