

Revised circulation scheme north of the Denmark Strait

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Abstract

The circulation and water mass transports north of the Denmark Strait are investigated using recently collected and historical in-situ data along with an idealized numerical model and atmospheric reanalysis fields. Emphasis is placed on the pathways of dense water feeding the Denmark Strait Overflow Water plume as well as the upper-layer circulation of freshwater. It is found that the East Greenland Current (EGC) bifurcates at the northern end of the Blosseville Basin, some 450 km upstream of the Denmark Strait, advecting overflow water and surface freshwater away from the boundary. This “separated EGC” flows southward adjacent to the previously identified North Icelandic Jet, indicating that approximately 70% of the Denmark Strait Overflow Water approaches the sill along the Iceland continental slope. Roughly a quarter of the freshwater transport of the EGC is diverted offshore via the bifurcation. Two hypotheses are examined to explain the existence of the separated EGC. The atmospheric fields demonstrate that flow distortion due to the orography of Greenland imparts significant vorticity into the ocean in this region. The negative wind stress curl, together with the closed bathymetric contours of the Blosseville Basin, is conducive for spinning up an anti-cyclonic gyre whose offshore branch could represent the separated EGC. An idealized numerical simulation suggests instead that the current is primarily eddy-forced. In particular, baroclinic instability of the model EGC spawns large anti-cyclones that migrate offshore and coalesce upon reaching the Iceland continental slope, resulting in the separated EGC. Regardless of the formation mechanism, the recently obtained shipboard data and historical hydrography both indicate that the separated EGC is a permanent feature of the circulation north of the Denmark Strait.

Keywords: Denmark Strait, East Greenland Current, North Icelandic Jet, Blosseville Basin, Denmark Strait Overflow Water, Arctic freshwater export

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1. Introduction

The meridional exchange across the Greenland-Scotland Ridge is of key importance for the North Atlantic climate system. Warm, saline Gulf Stream-origin waters flow northward across the ridge into the Nordic seas, release heat to the atmosphere, and are transformed into dense overflow waters. These waters return southward by flowing through gaps in the ridge as overflow plumes, which form the lower limb of the Atlantic Meridional Overturning Circulation (AMOC). As part of the horizontal circulation, fresh surface waters from the Arctic Ocean are fluxed southward along the western boundary of the Nordic seas and across the ridge. Most of this freshwater, as well as the densest portion of the AMOC – the Denmark Strait Overflow Water (DSOW) – pass between Greenland and Iceland. As such, the Denmark Strait is a critical and complex choke point in the subpolar circulation, and the pathways and water mass transports north of the strait help dictate the magnitude of the exchange between the Nordic seas and the North Atlantic Ocean.

The climatic importance of the deep overflows across the Greenland-Scotland Ridge was first hypothesized by Cooper (1955), and about a decade later remarkably accurate estimates of the overflow transports had been made (see Dickson et al., 2008, for an overview of the early measurements). The most recent observations of the DSOW near the sill indicate a mean transport of 3.4 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3/\text{s}$, Jochumsen et al., 2012). The first definitive scenario for the source of the DSOW was put forth by Swift et al. (1980) who suggested that the water comprising the plume originated from open-ocean convection in the central Iceland Sea (Swift and Aagaard, 1981). A later study then steered the community towards the idea that the transformation of Atlantic inflow into DSOW occurs primarily within the boundary current system of the Nordic seas (Mauritzen, 1996). This notion has persisted, supported by a recent study using over 50 years of historical hydrographic data (Eldevik et al., 2009).

Consistent with this, it was noted that Atlantic-origin water, modified by some exchange with the Greenland and Iceland seas, comprised the bulk of the overflow (Rudels et al., 2002, 2005). Geochemical tracer data suggest that the DSOW is a complex mixture of a large set of water masses originating from different regions. However, it is generally believed that Atlantic-origin water forms the major part of the overflow (Tanhua et al., 2005, 2008; Jeansson et al., 2008), with some contribution from the Iceland Sea (Olsson et al., 2005). One common link in all of these studies is that the primary pathway by which the source waters of the DSOW enter the strait is the East Greenland Current (EGC, Fig. 1). There is increasing evidence, however, that this may not be the case.

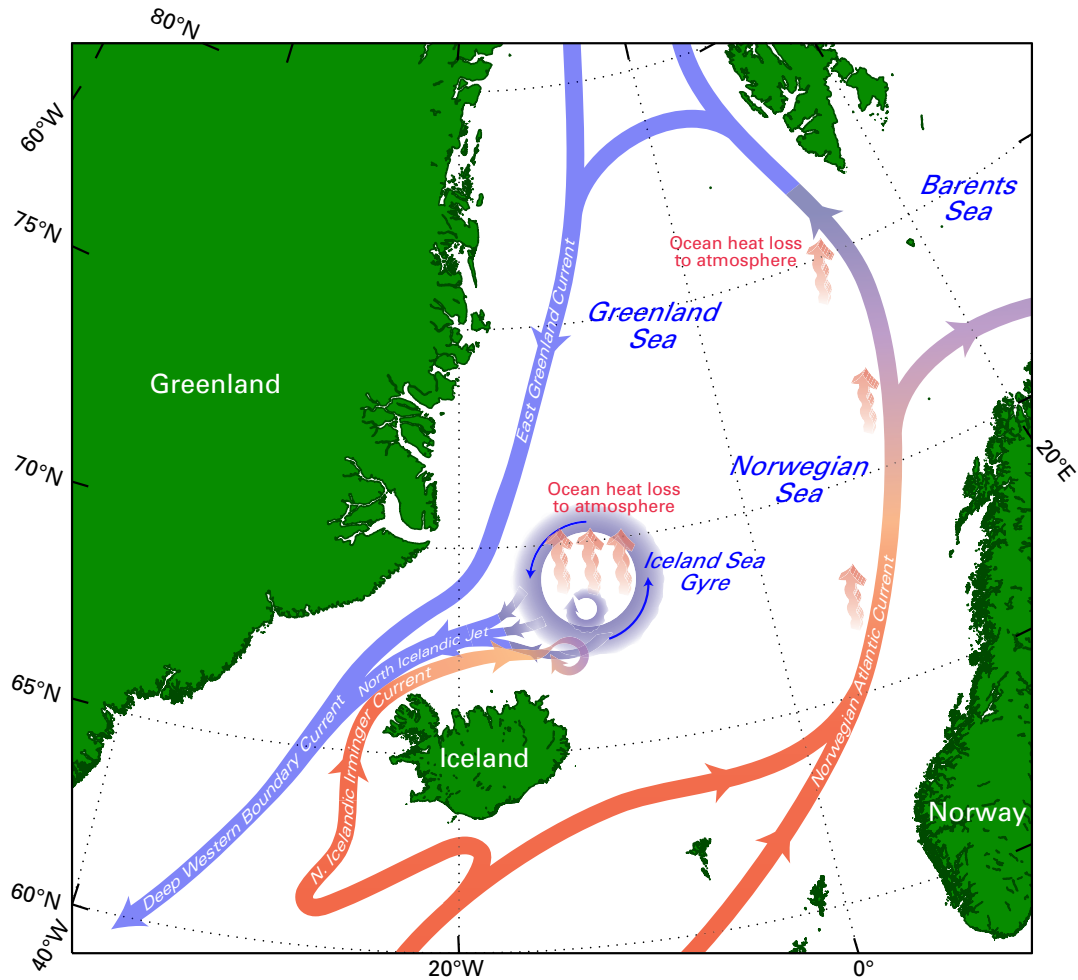


Figure 1: Two overturning circulation schemes in the Nordic seas in which warm, light surface waters (red colors) are transformed into cold, dense overflow waters (blue colors). The first scheme is the boundary current loop around the perimeter of the Nordic seas identified by Mauritzen (1996). The second scheme is the recently hypothesized interior loop north of Iceland (Våge et al., 2011b).

33 A previously unknown current flowing along the continental slope north of Iceland in the direc-
 34 tion of the Denmark Strait was identified by Jónsson (1999) and Jónsson and Valdimarsson (2004),
 35 which is now referred to as the North Icelandic Jet (NIJ, Fig. 1). These studies found that the NIJ
 36 was potentially of sufficient strength to account for most of the transport of DSOW across the sill if
 37 some entrainment of ambient water was included. However, it was subsequently suggested that the
 38 NIJ is not an independent current, but rather a branch of the EGC. For example, hydraulic theory
 39 predicts that as the deep-reaching EGC approaches the Denmark Strait and encounters shoaling
 40 bathymetry, the Iceland continental slope becomes a dynamical western boundary due to the topo-

41 graphic beta effect arising from the sloping bottom, forcing the EGC to cross from the Greenland
42 side to the Iceland side (e.g. Pratt, 2004). Another possibility, present in the model simulations
43 of Köhl et al. (2007), is that the NIJ is a branch of the EGC that bifurcates upstream of the strait as
44 part of a time dependent process. According to their model, the total volume transport of the two
45 branches is constant and the DSOW may be supplied by either branch.

46 By contrast, other model studies suggest that, at times, the NIJ may not be directly related to
47 the EGC. Köhl (2010) argued that the magnitude of the cyclonic wind stress curl in the region
48 dictates the source of the NIJ. In particular, when the curl is strong the NIJ originates from the
49 EGC, but during periods of weak curl the NIJ emanates from the northern Iceland Sea. In the
50 latter case the current stems from an offshoot of the weakening cyclonic circulation in the Iceland
51 Sea. In some of the idealized configurations of a similar model, Käse et al. (2009) found that the
52 NIJ can originate from a southward flow of dense water along the Jan Mayen Ridge northeast of
53 Iceland. Such temporal switching between sources of the overflow water has also been noted in
54 observational studies (Rudels et al., 2003; Olsson et al., 2005; Holfort and Albrecht, 2007).

55 A recent investigation has shed further light on the nature and source of the NIJ, which suggests
56 that the dense current is neither a branch of the EGC nor does it originate from the northern Iceland
57 Sea or the Jan Mayen Ridge. Data from two extensive hydrographic/velocity surveys along the
58 Iceland slope in 2008 and 2009 imply that the NIJ advects both the densest overflow water into
59 the Denmark Strait as well as roughly half of the total overflow transport (1.5 ± 0.2 Sv, Våge
60 et al., 2011b). Våge et al. (2011b) traced the current upstream as far as the northeast corner of
61 Iceland, where it weakened considerably (recently collected unpublished data support this view as
62 well). Its distinct hydrographic properties provide additional evidence that the NIJ is independent
63 from the EGC. Using an idealized numerical simulation, Våge et al. (2011b) argued that the NIJ
64 mainly originates along the north slope of Iceland as a deep limb of a local overturning cell whose
65 upper branch is the North Icelandic Irminger Current (NIIC, Fig. 1), which flows northward on
66 the eastern side of the Denmark Strait. Specifically, the warm, salty inflow is exchanged laterally
67 with dense water transformed via air-sea interaction in the interior Iceland Sea. The dense water
68 subsequently sinks near the boundary to form the NIJ. These results place a renewed emphasis on
69 the Iceland Sea as a potential contributor to the AMOC.

70 While the transport of dense overflow water through the Denmark Strait is reasonably well
71 quantified (Jochumsen et al., 2012), the flow of buoyant freshwater through the strait remains
72 largely unknown. Only recently have estimates of the EGC freshwater volume transport been

73 made (Holfort et al., 2008). The ability to measure this transport using moored instruments is
74 hampered by the presence of ice. However, Holfort and Meincke (2005) successfully deployed
75 moorings at 74°N east of Greenland, approximately halfway between the Fram Strait and the
76 Denmark Strait, resulting in the first EGC total (liquid and solid) freshwater flux estimate of 40-
77 55 mSv (referenced to a salinity of 34.9). In the Fram Strait an extensive mooring array has
78 been maintained since 1997, and a decade-long time series of liquid freshwater flux was recently
79 presented by de Steur et al. (2009). Some synoptic measurements of liquid freshwater flux along
80 the east coast of Greenland also exist (e.g. Nilsson et al., 2008; Sutherland and Pickart, 2008) as
81 well as estimates of the solid freshwater export from the Arctic Ocean through the Fram Strait (e.g.
82 Kwok et al., 2004).

83 One of the important aspects of freshwater in the high-latitude climate system is its ability to
84 influence deep convection. However, in order to impact the convective activity in the Nordic seas
85 and subpolar North Atlantic, there must be a flux of freshwater from the boundary current into
86 the interior convective regions. There are two known direct export pathways of freshwater from
87 the EGC into the Nordic seas: The Jan Mayen Current north of the Jan Mayen Fracture Zone and
88 the East Icelandic Current north of Iceland. However, these pathways together account for only
89 a small fraction of the freshwater flux through the Fram Strait (15 mSv or about 13%, Jónsson,
90 2007; Dickson et al., 2007). Additional exchange between the EGC and the interior is thought
91 to be minor (Aagaard and Carmack, 1989; Nilsson et al., 2008), leaving the bulk of the liquid
92 freshwater to remain within the EGC (this is not necessarily the case for the solid freshwater, see
93 Jones et al., 2008; Dodd et al., 2009, 2012). Freshwater volume transports of the EGC through
94 the Denmark Strait have also been obtained by constructing budgets for the Nordic seas (Dickson
95 et al., 2007; Segtnan et al., 2011), but these values represent residuals in the basin-wide balances.

96 While the importance of the Denmark Strait for the North Atlantic climate system is well es-
97 tablished, the complex circulation in and upstream of the strait is not fully understood. Oceanog-
98 raphers have been aware of the northern overflows for a century (Nansen, 1912), yet a consensus
99 has not been reached regarding the origin of their source waters. Closed heat and freshwater bud-
100 gets for the Nordic seas will not be attainable until reliable transport measurements throughout the
101 water column have been made in the vicinity of the strait. The motivation for the present study
102 is to advance our understanding of the shallow and deep circulation in this critically important
103 area. Using a collection of oceanic and atmospheric data sets, together with a numerical model,
104 we present new aspects of the water mass pathways and dynamics in this region, with emphasis

105 on the EGC system. We provide evidence of a heretofore unknown interior branch of the EGC
106 north of the Denmark Strait that impacts the supply of both buoyant freshwater and dense overflow
107 water to the strait. We refer to this current as the “separated EGC”. Our primary objectives are to
108 establish the existence of the separated EGC, quantify its structure and transport, and elucidate the
109 dynamics by which it is formed, including the role of atmospheric forcing.

110 The structure of the paper is as follows. The various data sets and methods are presented in
111 Section 2. We use a collection of high-resolution synoptic realizations of a hydrographic/velocity
112 transect from Greenland to Iceland in Section 3, along with historical hydrographic data in Sec-
113 tion 4, to investigate the circulation north of the Denmark Strait. The wind forcing is described
114 using atmospheric reanalysis fields in Section 5. Finally, a numerical model is used in Section 6 to
115 explore the internal and external forcing mechanisms responsible for the separated branch of the
116 EGC and the associated interior flux of overflow water and freshwater.

117 **2. Data and methods**

118 *2.1. Synoptic sections*

119 Four synoptic hydrographic/velocity realizations of a transect from Greenland to Iceland across
120 the Blosseville Basin (Fig. 2) are considered in the study. The transect, known as the Kögur
121 section, was occupied in August 2004, October 2008, August 2011, and August 2012. Some
122 aspects of two of the occupations have been previously presented (Sutherland and Pickart, 2008;
123 Våge et al., 2011b). The hydrographic measurements were obtained using a Sea-Bird conductivity-
124 temperature-depth (CTD) instrument, and velocities were measured using acoustic Doppler current
125 profilers (ADCPs): An upward- and downward-facing lowered ADCP system in August 2012,
126 and vessel-mounted ADCPs on the remaining cruises. The vessel-mounted ADCP instrument
127 malfunctioned at the western end of the August 2004 occupation, and hence no velocity data were
128 obtained on the Greenland shelf during that cruise. Vertical sections of potential temperature,
129 salinity, potential density, and ADCP velocity for each cruise were constructed using Laplacian-
130 spline interpolation with a grid spacing of 2 km by 10 m. From the temperature and salinity fields,
131 the relative geostrophic flow normal to each section was calculated, which was then referenced
132 by matching this to the vertically averaged ADCP velocities over the common depth interval at
133 each horizontal grid point (Pickart et al., 2005). The accuracies of the pressure, temperature,
134 salinity, and absolutely referenced velocity fields are estimated to be 0.3 dbar, 0.001 °C, 0.002,
135 and 3.6 cm s⁻¹, respectively. See Våge et al. (2011b) for further details of the data processing

136 and Appendix A for a description of the de-tiding procedure and the methodology for calculating
137 transports, including errors.

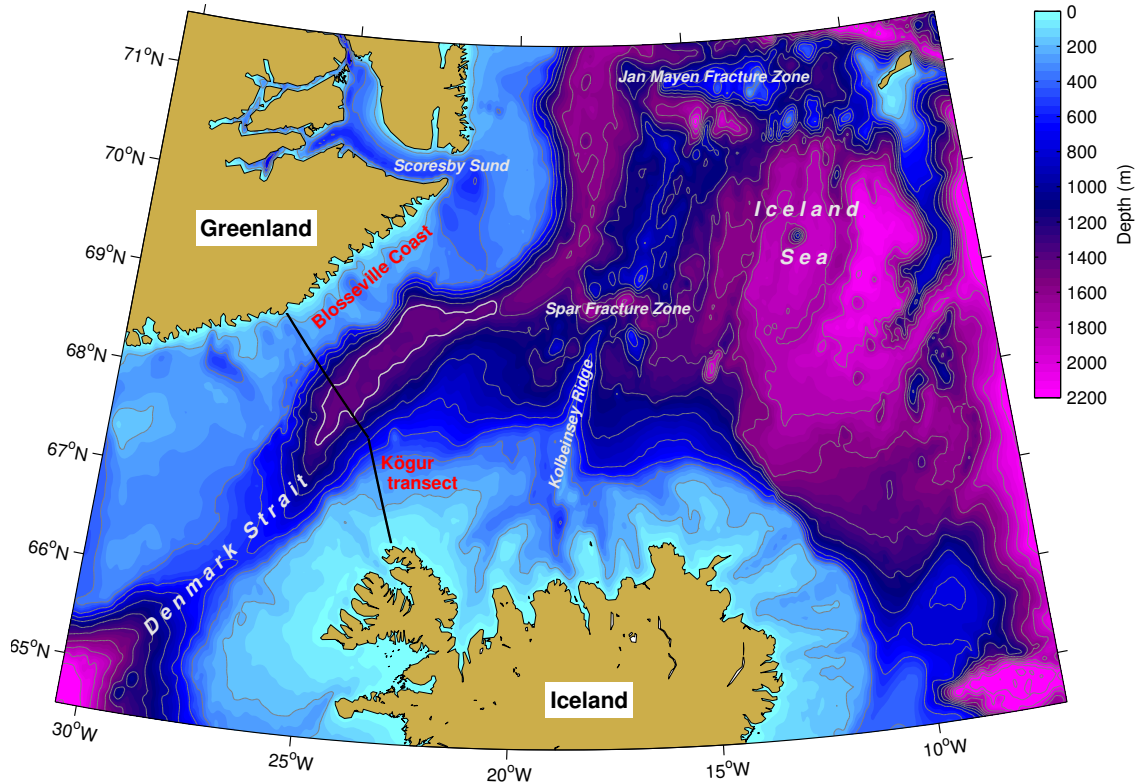


Figure 2: Bathymetry of the Iceland Sea and the Denmark Strait from the ETOPO2 2-minute elevation data base. The closed white contour is the 1400 m isobath roughly delineating the Blosseville Basin, and the black line identifies the Kögur transect (named after the Kögur mountain on the northwest coast of Iceland). Depth is contoured with 200 m increments starting at 200 m.

138 2.2. Historical hydrography

139 The historical hydrographic data employed in this study cover the period from 1980 to the
140 present and were acquired from the archives of the Marine Research Institute of Iceland, the Inter-
141 national Council for the Exploration of the Seas (ICES), the World Ocean Database, the Norwegian
142 Iceland Seas Experiment (NISE) database (Nilsen et al., 2008), and the Argo global program of
143 profiling floats (using only delayed-mode data, which have been corrected for drift in the conduc-
144 tivity and pressure sensors, Wong et al., 2003). The profiles from these different sources were
145 combined into a single product, hereafter referred to as the historical hydrographic data set. Only
146 observations obtained during the summer half-year period of May through October (66% of the

147 total number of measurements) are considered here due to a dearth of wintertime data on the Green-
148 land continental shelf and slope. See Appendix B for details of the quality control and gridding of
149 the historical hydrography.

150 *2.3. Meteorological fields*

151 We employ two different reanalysis products in this work. The Interim Reanalysis (ERA-I)
152 from the European Center for Medium Range Weather Forecasts is a global product (Dee et al.,
153 2011). We use the 0.75° interpolated 6-hourly fields for the period from January 1979 to December
154 2011. Comparison with aircraft and ship observations in the southeast Greenland region show good
155 agreement with ERA-I (Renfrew et al., 2009; Harden et al., 2011).

156 For a higher resolution view of the surface wind field in the region of interest, we also make
157 use of the North American Regional Reanalysis (NARR) from the U.S. National Meteorological
158 Center (Mesinger et al., 2006). The NARR is a regional dataset that covers the North American
159 continent as well as adjoining oceanic regions including southeast Greenland and the Irminger
160 Sea, with lateral boundary conditions provided by the NCEP-2 global reanalysis. The NARR has
161 a horizontal resolution of approximately 32 km. For this paper we use the full 3-hourly resolution
162 data set for the period from January 1979 to December 2011. Recent studies of the flow distortion
163 around the topography of southern Greenland indicate that the NARR surface fields are in good
164 overall agreement with both aircraft and buoy observations (Moore et al., 2008; Renfrew et al.,
165 2009).

166 *2.4. Idealized simulations*

167 A high-resolution regional general circulation model is implemented to aid in the interpretation
168 of the observational results and to provide insights on the dynamics of the circulation. The model
169 is the MITgcm (Marshall et al., 1997), which solves the hydrostatic primitive equations of motion
170 on a fixed Cartesian, staggered C-grid in the horizontal and at constant depths in the vertical.
171 Bottom topography is treated with a partial cell, which provides a high-resolution representation
172 of the bottom topography while retaining accuracy in the calculation of the horizontal pressure
173 gradient (Adcroft et al., 1997). The model uses a linear equation of state $\rho = \rho_0 + \beta(S - S_0)$,
174 where $\beta = 0.8 \text{ kg m}^{-3}$, $\rho_0 = 1026.5 \text{ kg m}^{-3}$ is a reference density, and $S_0 = 32.5$ a reference
175 salinity. For simplicity, only salinity is considered (temperature is constant).

176 The model domain is a channel oriented along the east coast of Greenland, extending 864 km
177 in the along-boundary direction, y , and 480 km in the offshore direction, x (Fig. 3, rotated counter-
178 clockwise by 33°). The Coriolis parameter is $f = 1.3 \times 10^{-4} \text{ s}^{-1}$, taken to be constant. The

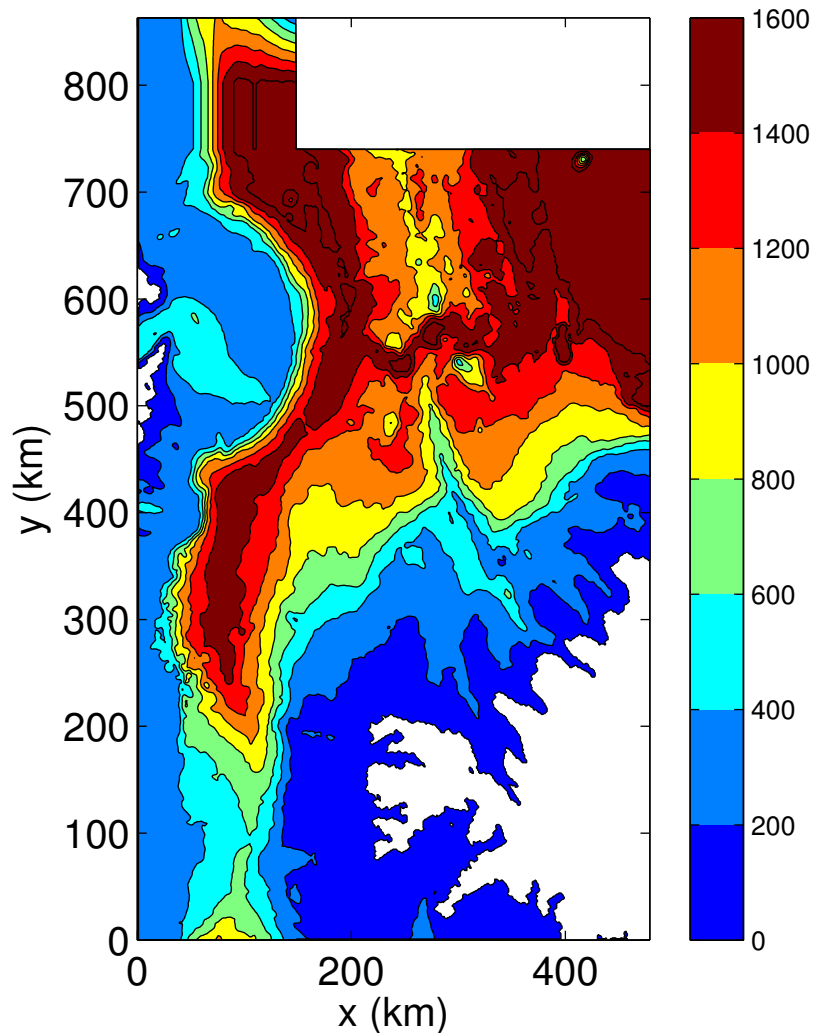


Figure 3: Model domain and bottom topography. The white regions are land.

179 bottom topography has been linearly interpolated to the model grid from the ETOPO2 2-minute
 180 elevation data base and then smoothed with a 5-point filter. The model is initialized with fresh,
 181 stratified water on the shelf (bottom depths shallower than 400 m) with a minimum salinity of 32
 182 at the surface, which varies exponentially with vertical scale of 250 m towards a deep salinity of
 183 35. The basin interior (bottom depths greater than 575 m) is filled with unstratified water with a
 184 salinity of 35. The salinity at each depth in the transition region between 400 and 575 m bottom
 185 depth is linearly interpolated between the shelf and interior values at each depth. The stratification
 186 in the region between 740 and 840 km in y and the region offshore of 350 km in x are restored
 187 towards this initial state with a time scale of 5 days. This provides a source of freshwater on the

188 shelf and saline water in the basin interior. The long-time evolution is not qualitatively sensitive to
 189 the details of the initial profile, provided that freshwater is initialized over the shelf. The bottom
 190 topography in the region between 804 and 864 km is linearly interpolated from the value at $y = 0$
 191 to that at $y = 804$ km. In this way, the flow out of the southern end of the channel provides the
 192 inflow from the north and mass is conserved, while the restoring of salinity maintains freshwater
 193 on the shelf. The region offshore of this northern restoring region is a solid boundary, provided to
 194 prevent large-scale recirculations in the interior. The offshore boundary of the model is placed far
 195 from the region of interest, and salinity restoring for $x > 350$ km is used to minimize influences of
 196 Ekman transport interacting with the solid eastern boundary.

197 The horizontal resolution of the model is 1 km in both the x and y directions. There are 30
 198 levels in the vertical, with grid spacing increasing from 5 m over the upper 20 m to 250 m near the
 199 bottom. The maximum bottom depth is approximately 1500 m. The model incorporates second
 200 order vertical viscosity and diffusivity with coefficients of $10^{-5} \text{ m}^2 \text{ s}^{-1}$. Horizontal viscosity is
 201 parameterized as a second order operator with the coefficient A_h determined by a Smagorinsky
 202 closure as

$$A_h = (\nu/\pi)^2 \Delta^2 D, \quad (1)$$

203 where Δ is the grid spacing, and D is the deformation rate defined as $D = \left[(u_x - v_y)^2 + (u_y + \right.$
 204 $\left. v_x)^2 \right]^{1/2}$, u and v are the horizontal velocities, subscripts indicate partial differentiation, and $\nu = 1$.
 205 There is no explicit diffusion of salinity.

206 3. Synoptic transects north of the Denmark Strait

207 The Kögur transect (Fig. 2) extends from Greenland to Iceland across the Blosseville Basin,
 208 hence capturing all of the advective pathways into the Denmark Strait. The transect is sufficiently
 209 far upstream that distinct deep pathways can be distinguished prior to forming the merged DSOW
 210 plume that subsequently overflows the sill. To investigate the conditions north of the strait, we first
 211 examine the mean flow patterns and hydrography, and then inspect the individual realizations in
 212 more detail. Finally, transport estimates for each pathway are presented.

213 3.1. Mean velocity structure

214 The EGC flows southward from the Fram Strait, roughly paralleling the east Greenland shelf
 215 break, as part of the cyclonic boundary current system transiting the perimeter of the Nordic seas.
 216 Freshwater exported from the Arctic in the form of Polar Surface Water (PSW, see Rudels et al.

217 (2005) and Våge et al. (2011b) for water mass definitions) constitutes most of the upper part of the
218 EGC, while Atlantic-origin overflow water is advected at depth. At the Kögur transect the EGC
219 is evident as a surface-intensified shelf break current centered near 185 km with a deep extension
220 over the Greenland continental slope (Fig. 4a).

221 Farther to the east, between 75 and 130 km over the deeper part of the Iceland continental
222 slope, another southward-flowing surface-intensified current is evident. This transports primarily
223 water of similar hydrographic properties as the EGC (Fig. 4b, c) and is located near the shallow
224 hydrographic front separating the Polar and Atlantic waters. We propose that this current is a
225 separated branch of the EGC and offer two hypotheses to explain its existence. The first hypothesis
226 is that the current is wind-driven, and the second is that it is eddy-driven. (As discussed below,
227 these two ideas are not mutually exclusive.) With regard to wind forcing, while most of the Nordic
228 seas is subject to cyclonic wind stress curl, the region encompassing the Blosseville Basin is in fact
229 characterized by anti-cyclonic wind stress curl. Due to the closed bathymetric contours within the
230 basin (Fig. 2), this would tend to force an anti-cyclonic circulation. In this case the separated EGC
231 is the southward-flowing branch of the gyre, while the northward flow between the separated EGC
232 and the shelf break EGC¹ would be the return branch of the gyre (Fig. 4a). This gyre scenario is
233 presented in detail in Section 5. With regard to eddy forcing, the numerical simulations presented
234 below indicate that baroclinic instability of the shelf break EGC just north of the Blosseville Basin
235 results in a continuous spawning of anticyclonic eddies. These eddies then migrate offshore and
236 equatorward until encountering the shoaling topography of the Iceland continental slope, at which
237 point they coalesce to form a surface-intensified southward current. This eddy scenario is discussed
238 in Section 6. Either way, it appears that the separated EGC is a permanent and substantial part of
239 the circulation north of the Denmark Strait.

240 The separated EGC abuts, but is dynamically distinct from, the NIJ (in the 2004 occupation
241 the two currents were also geographically separated). The NIJ is characterized by isopycnals
242 that diverge westward from Iceland in the middle of the water column, consistent with a mid-
243 depth intensified southward-flowing current (Våge et al., 2011b). By contrast, the separated EGC
244 is surface intensified with uniformly downward sloping isopycnals to the west. This transition
245 in isopycnal slope is taken to represent the boundary between the NIJ and the separated EGC
246 (indicated by the dashed line in Figure 4). The other major current present in the Kögur section

¹From here on we use the term shelf break EGC to distinguish the EGC from the separated EGC in the Blosseville Basin.

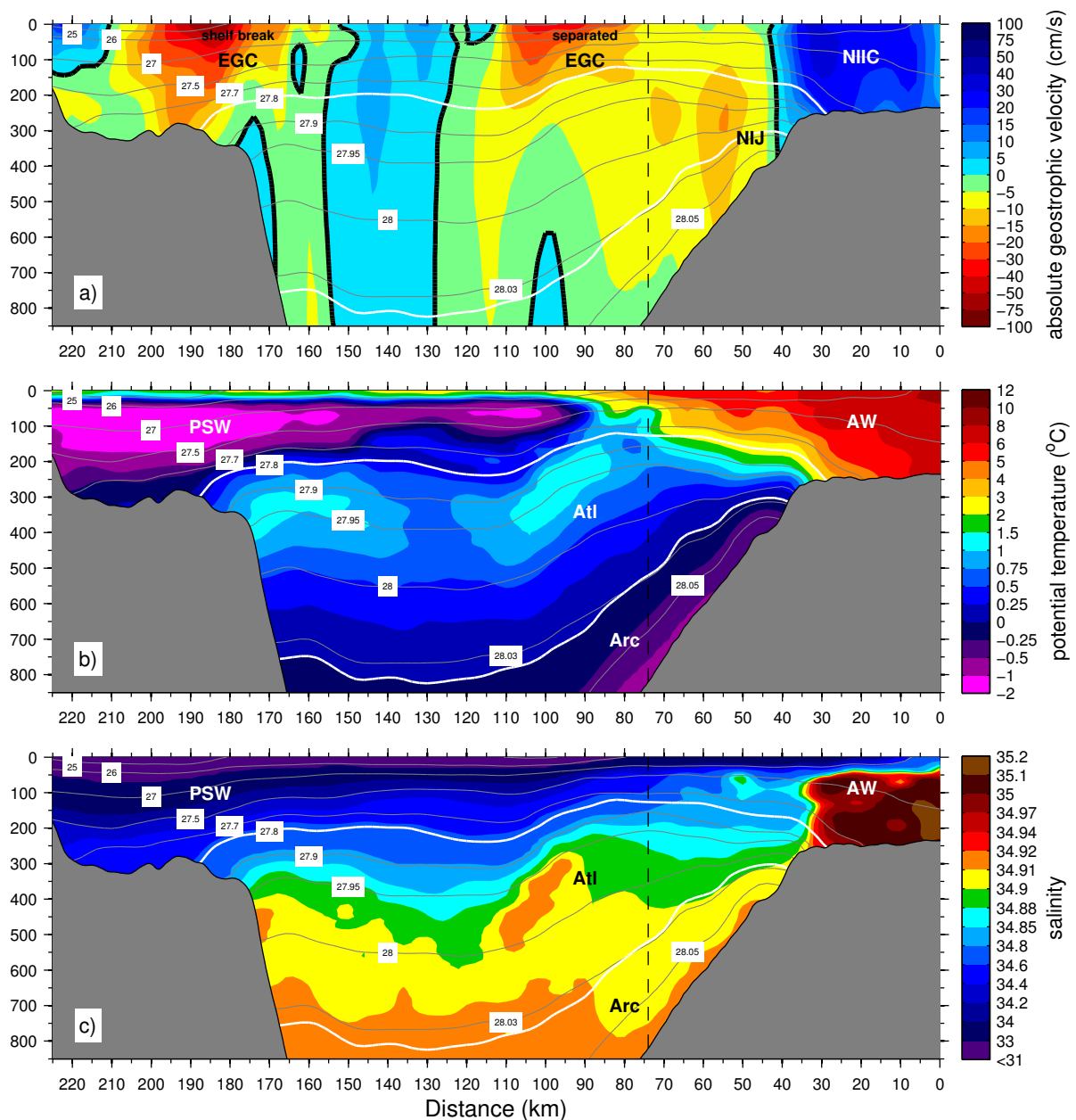


Figure 4: Mean vertical sections along the Kögur transect. (a) Absolutely referenced geostrophic velocity (color, cm/s, equatorward flow is negative), overlain by potential density (thin gray contours, kg/m³). The 27.8 kg/m³ isopycnal and the 0°C isotherm are highlighted in white. The dashed vertical line indicates the boundary between the separated EGC and the NIJ. The thick black line in (a) is the zero velocity contour. The acronyms are: EGC = East Greenland Current; NIJ = North Icelandic Jet; NIIC = North Icelandic Irminger Current; PSW = Polar Surface Water; AW = Atlantic Water; Atl = Atlantic-origin overflow water; Arc = Arctic-origin overflow water. (b) Potential temperature (color, °C) and (c) Salinity (color).

247 is the NIIC, which is the poleward extension of the Irminger Current. It advects warm, saline
248 Atlantic Water (AW) northward through the Denmark Strait in the vicinity of the shelf break, and
249 influences the climate and ecosystem north of Iceland (e.g. Jónsson and Valdimarsson, 2012b).
250 The NIIC is also believed to be the surface component of a regional overturning loop whose lower
251 branch consists of Arctic-origin overflow water transformed within the central Iceland Sea and
252 transported equatorward by the NIJ (Fig. 1; Våge et al., 2011b).

253 3.2. Mean hydrographic structure

254 In the mean Kögur section, cold, fresh PSW dominates the upper water column on the Green-
255 land shelf. This water mass extends far offshore into the interior of the Blosseville Basin, more
256 than halfway across the transect (Fig. 4b, c). Warm, saline AW is found over the Iceland shelf and
257 slope.

258 Below the surface layer, water denser than $\sigma_\theta = 27.8 \text{ kg/m}^3$ (the upper white contour in Fig-
259 ure 4) has traditionally been identified as overflow water (e.g. Dickson and Brown, 1994). Follow-
260 ing Våge et al. (2011b), we distinguish two types of overflow water: Atlantic- and Arctic-origin
261 overflow waters, warmer and colder than 0°C , respectively (the 0°C contour is highlighted white
262 in Figure 4). These names refer to the geographic domain in which the transformation from sur-
263 face to overflow water takes place (Swift and Aagaard, 1981). The Atlantic-origin overflow water
264 is identified by an intermediate maximum in temperature and salinity, and is primarily found be-
265 tween the 27.9 and 28.0 kg/m^3 isopycnals. In the mean Kögur transect this layer extends from
266 the Greenland continental slope and shoals towards the surface front of the separated EGC. Two
267 cores of Atlantic-origin overflow water, near 100 and 170 km, appear to be associated with the
268 separated EGC and the shelf break EGC, respectively. On the eastern side of the Kögur section
269 the Arctic-origin overflow water is banked up high on the Iceland continental slope. This forms
270 the densest component of the DSOW plume (generally denser than 28.03 kg/m^3), and is primarily
271 supplied by the NIJ (Våge et al., 2011b).

272 The PSW and Atlantic-origin overflow water are generally associated with the EGC and hence
273 largely confined to the Greenland shelf and slope (this is particularly true for the shallow waters,
274 Aagaard and Carmack, 1989; Nilsson et al., 2008). However, our data reveal that, associated
275 with the separated EGC, these water masses are clearly found in the interior of the Blosseville
276 Basin as well (Fig. 4b, c). Regardless of the process by which the separated EGC is formed,
277 this current provides another means of shelf–basin exchange in addition to the direct advective
278 pathways represented by the Jan Mayen Current and the East Icelandic Current.

279 *3.3. Temporal variability*

280 The major flow features present in the mean Kögur transect – the two branches of the EGC,
281 the NIJ, and the NIIC – are evident as well in each of the individual realizations (Fig. 5).² On the
282 western side of the transect one sees the shelf break EGC, although it is variable in magnitude from
283 section to section. This is likely due in part to the strength of the wind forcing. For example, in
284 October 2008 the shelf break EGC was quite strong because it was under the influence of northerly
285 winds, while in August 2012 the winds were predominantly from the south, which likely resulted
286 in the poleward flow at depth that “split” the shelf break EGC in two. The deep extension of the
287 shelf break EGC, which carries the overflow water, seems to vary both in strength and in lateral
288 position in the individual realizations.

289 On the eastern side of the Kögur section the NIIC was sampled on three occasions (the August
290 2011 occupation did not extend onto the Iceland continental shelf), evident as a surface-intensified
291 poleward flow. In October 2008 and August 2012 the current was located near the shelf break,
292 while in August 2004 it was found farther inshore on the shelf. Seaward of this, in all four re-
293 alizations, the NIJ was situated near the 650 m isobath, which is also the depth of the Denmark
294 Strait sill. This is consistent with the results of Våge et al. (2011b) who demonstrated that the NIJ
295 feeds the DSOW. At times the NIJ consisted of distinct filaments of equatorward flow, the reasons
296 for which remain unclear. However, we note that the NIJ’s characteristic isopycnal divergence and
297 mid-depth intensification were present in each realization.

298 Seaward of the NIJ, in the eastern part of the Blosseville Basin over the deep Iceland conti-
299 nental slope, the separated EGC is present in each occupation of the Kögur transect. This current
300 resides at the hydrographic front where the bulk of the PSW layer ends. As is true for the shelf
301 break branch of the EGC, the lateral salinity gradient dominates the temperature gradient, resulting
302 in a density front that supports a surface-intensified southward-flowing jet. One sees that the pen-
303 etration depth of the separated EGC can be quite extensive. Interestingly, the first two realizations
304 of the Kögur section (2004 and 2008) are more reminiscent of the above-mentioned gyre scenario.
305 In those two cases there is a well-defined deep-reaching poleward flow to the west that could be
306 construed as the northward branch of the gyre. By contrast, the latter two realizations of the tran-
307 sect (2011 and 2012) suggest the presence of eddies. In particular, note the lens of cold PSW
308 centered at 105 km in the 2011 occupation (Figs. 6c and 7c), corresponding to an anti-cyclonic

²This provides justification for computing a mean section based only on four realizations.

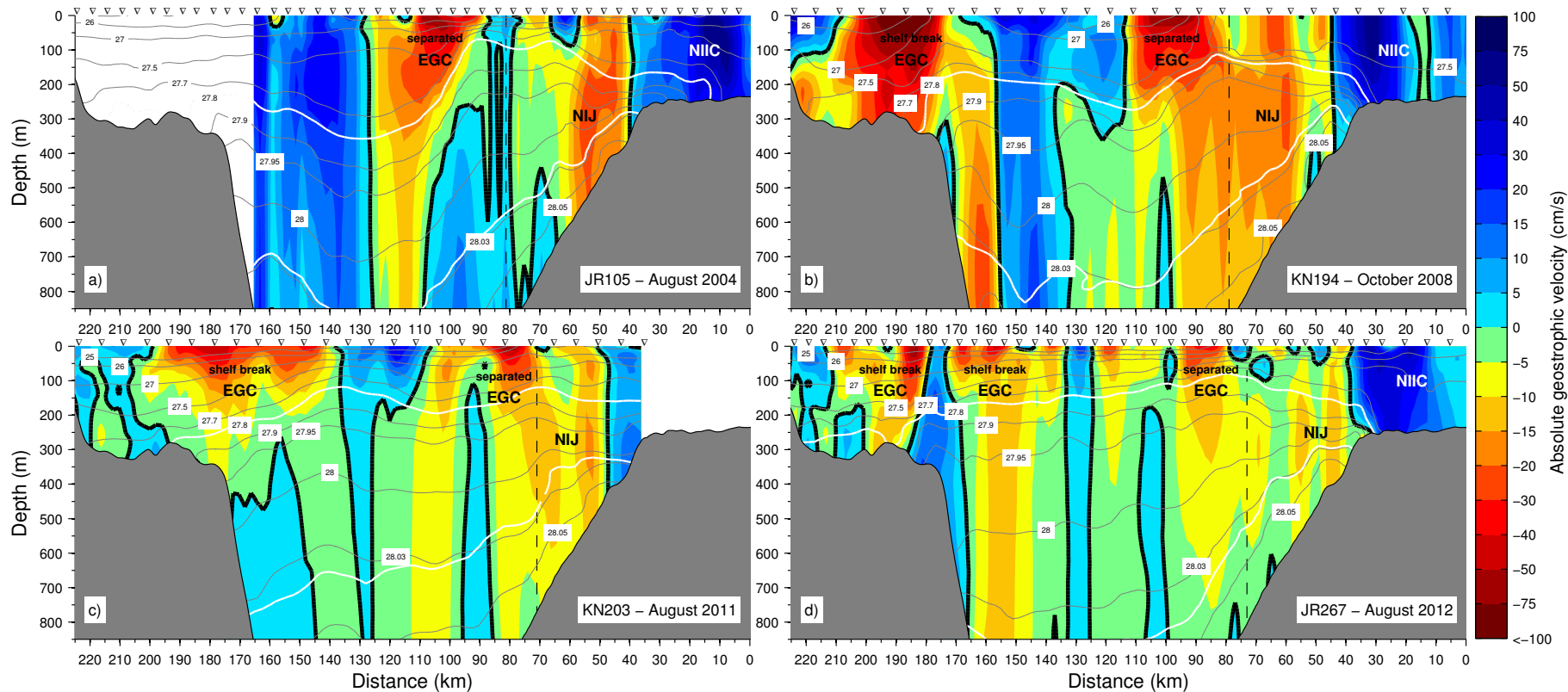


Figure 5: Vertical sections along the Kögur transect of absolute geostrophic velocity (color, cm/s, equatorward flow is negative) overlain by potential density (contours, kg/m^3). The panels show the (a) August 2004, (b) October 2008, (c) August 2011, and (d) August 2012 realizations. The 27.8 kg/m^3 isopycnal and the 0°C isotherm are highlighted in white. The dashed vertical line indicates the approximate boundary between the separated EGC and the NIJ. The black line is the zero velocity contour.

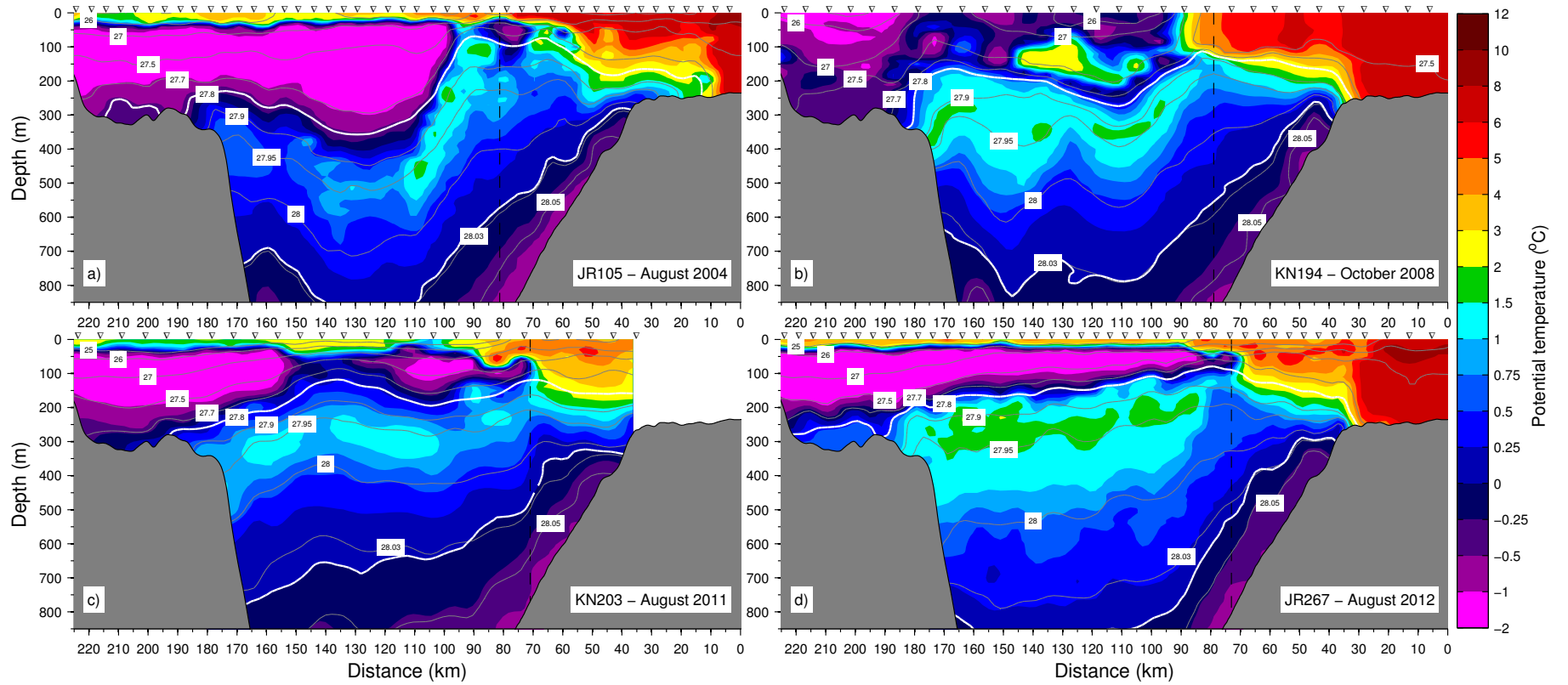


Figure 6: Same as Figure 5, but for potential temperature (color, °C).

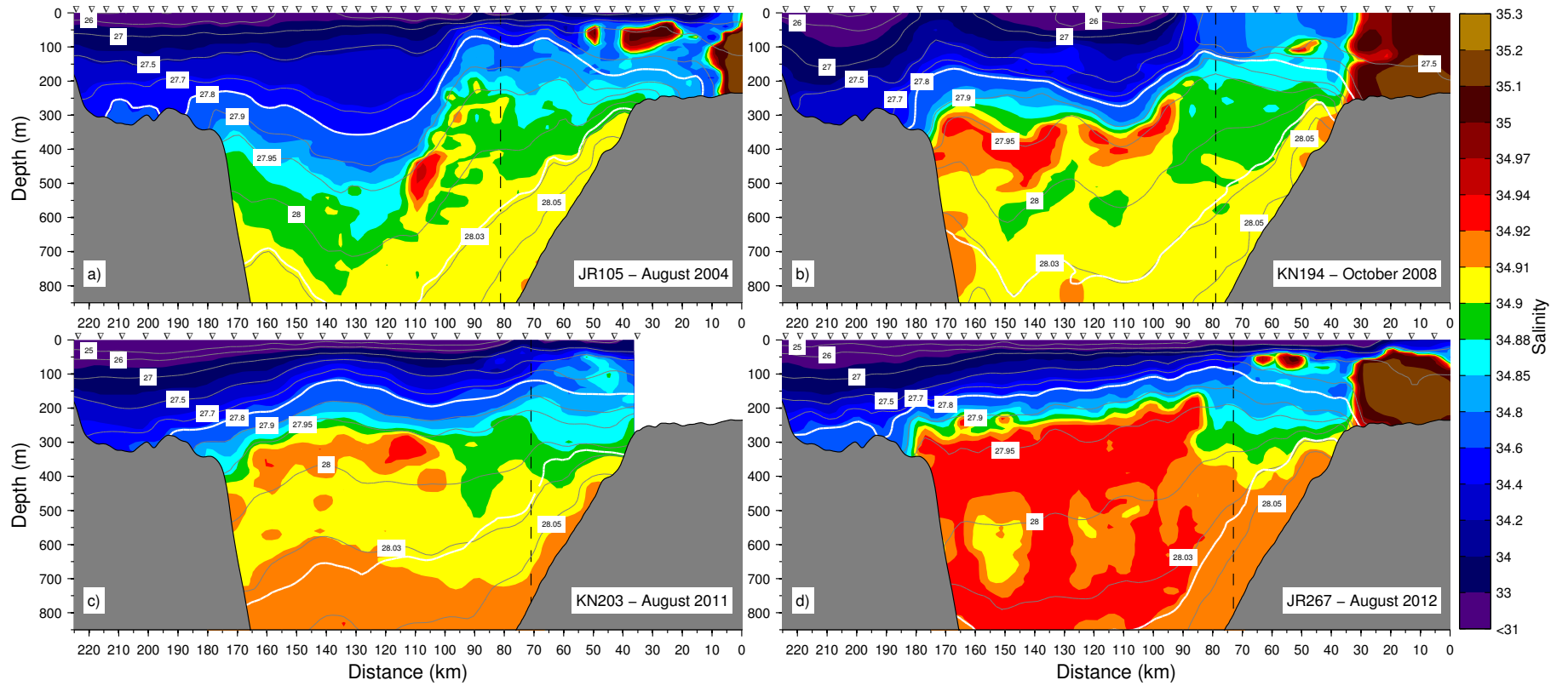


Figure 7: Same as Figure 5, but for salinity (color).

309 surface-intensified circulation (Fig. 5c). This is indicative of a large eddy that might have been
310 shed from the shelf break EGC. These two scenarios, eddy-forced vs. wind-forced, are elaborated
311 on later in the paper.

312 3.4. *Transports*

313 We now document the volume transport, and, where appropriate, the freshwater transport of
314 each of the equatorward currents present in the Kögur transect. This demonstrates the importance
315 of the NIJ as well as the separated EGC – both newly revealed features – in the circulation system
316 north of the Denmark Strait. Averaged over all four realizations, the shelf break EGC transported
317 108 ± 24 mSv and the separated EGC 29 ± 7 mSv of freshwater relative to a reference salinity of
318 34.8 (Fig. 8a). Hence, nearly one quarter of the total freshwater transport within the EGC system
319 takes place in the interior of the Blosseville Basin, which must be considered in order to obtain
320 accurate estimates of the freshwater export through the Denmark Strait. Most (>95%) of this
321 freshwater transport takes place within the upper 200 m of the water column, while a negligible
322 amount is transported by the NIJ.

323 There have been no mooring-based estimates of liquid freshwater transport in the vicinity of
324 the Denmark Strait because of the inherent risks due to pack-ice and icebergs. As such, the most
325 relevant numbers for comparison were obtained assuming balanced Nordic seas freshwater bud-
326 gets. Relative to reference salinities of 35.2 and 34.9, Dickson et al. (2007) and Segtnan et al.
327 (2011) computed freshwater fluxes through the Denmark Strait of 151 and 130 mSv, respectively.
328 Our corresponding numbers for the composite EGC are 223 ± 37 mSv and 159 ± 28 mSv. While
329 the estimates of Dickson et al. (2007) and Segtnan et al. (2011) are total freshwater flux (liquid and
330 solid), only small amounts of sea ice were encountered during our realizations, hence it is reason-
331 able to compare our estimates with the earlier studies. The volume transport of water lighter than
332 $\sigma_{\theta} = 27.8$ kg/m³ across the Kögur transect is 2.6 ± 0.3 Sv in the shelfbreak EGC and 1.3 ± 0.1 Sv
333 in the separated EGC. The combined transport is substantially larger than the 1.3 Sv of surface
334 outflow in the EGC reported by Hansen and Østerhus (2000).

335 The hydrographic analysis of Mauritzen (1996) suggested that the main path of overflow water
336 into the Denmark Strait is along the continental slope of east Greenland via the EGC. Our results
337 imply instead that the majority of the overflow water approaches the strait from the Iceland con-
338 tinental slope through a combination of the NIJ and the separated EGC. While this is consistent
339 with predictions from hydraulic theory (e.g. Pratt, 2004), we will show in Sections 5 and 6 that
340 the separated EGC has not simply switched to the Iceland side due to shoaling bathymetry of the

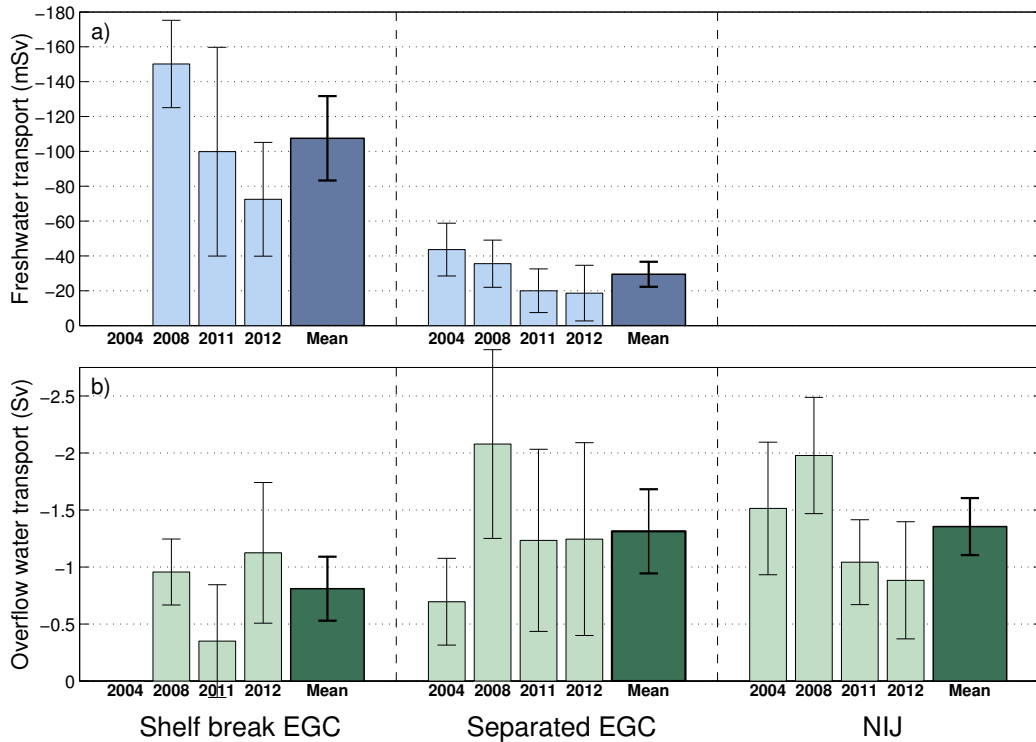


Figure 8: Transport estimates across the Kögur transect (equatorward flow is negative) for the shelf break EGC (left), the separated EGC (middle), and the NIJ (right). Panel a) shows freshwater transports (relative to a reference salinity of 34.8), and panel b) shows overflow water transports ($\sigma_\theta \geq 27.8 \text{ kg/m}^3$ and depth $< 650 \text{ m}$). The thin bars represent the 4 realizations and the thick bar is the mean value. The black lines are error bars.

341 strait. Furthermore, Våge et al. (2011b) have previously documented that the NIJ originates along
 342 the Iceland continental slope far upstream of the Denmark Strait. Hence the topographic beta effect
 343 does not seem to play an important role in the pathways of dense water leading to the sill. Our cal-
 344 culated average overflow water transports ($\sigma_\theta \geq 27.8 \text{ kg/m}^3$ and depth $< 650 \text{ m}$) are $0.8 \pm 0.3 \text{ Sv}$
 345 in the shelf break EGC, $1.3 \pm 0.4 \text{ Sv}$ in the separated EGC, and $1.4 \pm 0.3 \text{ Sv}$ in the NIJ (Fig. 8b).
 346 The combined sum of 3.5 Sv corresponds well with the long-term mean overflow water transport
 347 through the Denmark Strait of 3.4 Sv estimated by Jochumsen et al. (2012), and our NIJ overflow
 348 water transport is in good agreement with the previous estimate of $1.5 \pm 0.2 \text{ Sv}$ from two synoptic
 349 surveys (Våge et al., 2011b).

350 In the gyre scenario outlined above, some of the equatorward transport of freshwater and over-
 351 flow water in the separated EGC would recirculate before reaching the Denmark Strait sill. The
 352 poleward return flow is weaker, however, resulting in a net equatorward throughput. Based on the
 353 present set of sections, the poleward return flow is approximately half of the equatorward flow in

354 the separated EGC. Even if no permanent gyre exists in the Blosseville Basin, the synoptic sec-
355 tions indicate that there is some transient northward flow between the shelf break EGC and the
356 separated EGC. This would tend to reduce the composite EGC contribution to the total overflow
357 transport, although this smaller total transport is still in good agreement with the value calculated
358 by Jochumsen et al. (2012) when considering measurement errors. The northward flow would
359 similarly tend to reduce the composite EGC freshwater transport, but the resulting net throughput
360 is still higher than the previous calculations that assume a balanced Nordic seas freshwater budget.
361 It is worth noting that sparse measurements on the inner Greenland shelf indicate that the freshest
362 water is found adjacent to the coast, possibly associated with a southward flow. Such a coastal
363 current could have a significant freshwater transport, but this cannot be properly assessed with the
364 present set of sections.

365 The poleward flow of AW through the Denmark Strait in the NIIC has been monitored for a
366 number of years using moorings located roughly 80 km to the east of the Kögur transect. The
367 long-term mean transport at that location (known as the Hornbanki line) is 0.88 Sv (Jónsson
368 and Valdimarsson, 2012b). Delineating the AW by a salinity greater than 34.9 and temperature
369 warmer than 3°C (after Swift and Aagaard, 1981), we obtain a three-section mean NIIC transport
370 of 1.1 ± 0.2 Sv. AW is the only warm and saline water mass along the Kögur transect (Fig. 4), and
371 the transport estimate is not very sensitive to the precise criteria used. Our larger value of NIIC
372 transport may be due to synoptic variability, or there may be leakage of AW from the NIIC between
373 the two locations (Jónsson and Valdimarsson, 2005). On the other hand, the mooring-derived value
374 at the Hornbanki line could be an underestimate due to inadequate spatial coverage (Jónsson and
375 Valdimarsson, 2012b).

376 The process of aspiration, by which dense water upstream of a ridge is raised above sill level
377 to participate in an overflow, is believed to occur in the Strait of Gibraltar and in the Faroe Bank
378 Channel (Stommel et al., 1973; Kinder and Parrilla, 1987; Hansen and Østerhus, 2007). In our
379 occupations of the Kögur line the geostrophic velocities are generally weak at depths greater than
380 the Denmark Strait sill. In fact, the mean transport below 650 m is not significantly different from
381 zero. This implies that aspiration is not important in the Denmark Strait. There was, however,
382 considerable variability in the deep flow from section to section (not shown).

383 4. Historical hydrography north of the Denmark Strait

384 The synoptic sections presented above reveal that the EGC has two branches north of the Den-
385 mark Strait, and, accordingly, PSW occupies much of the interior Blosseville Basin. Furthermore,
386 a significant fraction of the Atlantic-origin overflow water is transported by the separated EGC.
387 We now use the historical hydrographic data set to demonstrate that our results are consistent with
388 previous measurements. The dynamic topography of the sea surface relative to 500 m is shown in
389 Figure 9, which reveals features of the upper ocean circulation in the region. The broad minimum
390 in the central Iceland Sea indicates the presence of a cyclonic gyre, and is consistent with earlier
391 findings using a more sparse hydrographic data set (Swift and Aagaard, 1981) and direct velocity
392 measurements (Voet et al., 2010). In the northwest part of the domain, along the east Greenland
393 shelf break (marked by the gray contour following the 500 m isobath in Figure 9), a sharp increase
394 in dynamic height is evident which implies surface-intensified equatorward flow. This is the shelf
395 break EGC. However, at the northern end of the Blosseville Basin near 69°N the high values of
396 dynamic topography extend into the interior. Here the separated EGC splits from the shelf break
397 EGC. Farther south, over the length of the Blosseville Basin (marked by the closed gray contour
398 following the 1400 m isobath in Figure 9), the two branches can be identified by regions of en-
399 hanced gradients in dynamic topography: Along the Greenland shelf break, and in the interior
400 along the base of the Iceland continental slope. Interestingly, a consistent northward return flow is
401 not visible in the mean map of dynamic topography. If it exists, it may not appear in the mean due
402 to a weaker baroclinic signature than the surface-intensified separated EGC (see Figure 4) or due
403 to sparse data in the western Blosseville Basin. We note that inspection of along-track altimeter
404 sea surface height also does not reveal a northward flow, but this product is not expected to perform
405 well in partially ice-covered waters and on such small spatial scales.

406 The distribution of polar waters north of the Denmark Strait closely resembles the dynamic
407 topography. Near-surface potential temperature and salinity fields (Fig. 10) show that the PSW
408 is closely confined to the shelf north of 69°N, whereas in the Blosseville Basin this water mass
409 spreads far into the interior to the base of the Iceland slope. (A slightly deeper level is considered
410 for temperature than for salinity in order to avoid the summertime surface warming due to insola-
411 tion, evident in Figure 4.) Presence of PSW over the deep Iceland slope is regularly observed at
412 the Kögur transect, and some of this water mass is likely mixed into the AW transported by the
413 NIIC along the shelf north of Iceland (Jónsson and Valdimarsson, 2012b).

414 While PSW was confined to the Greenland shelf north of 69°N, the Atlantic-origin overflow

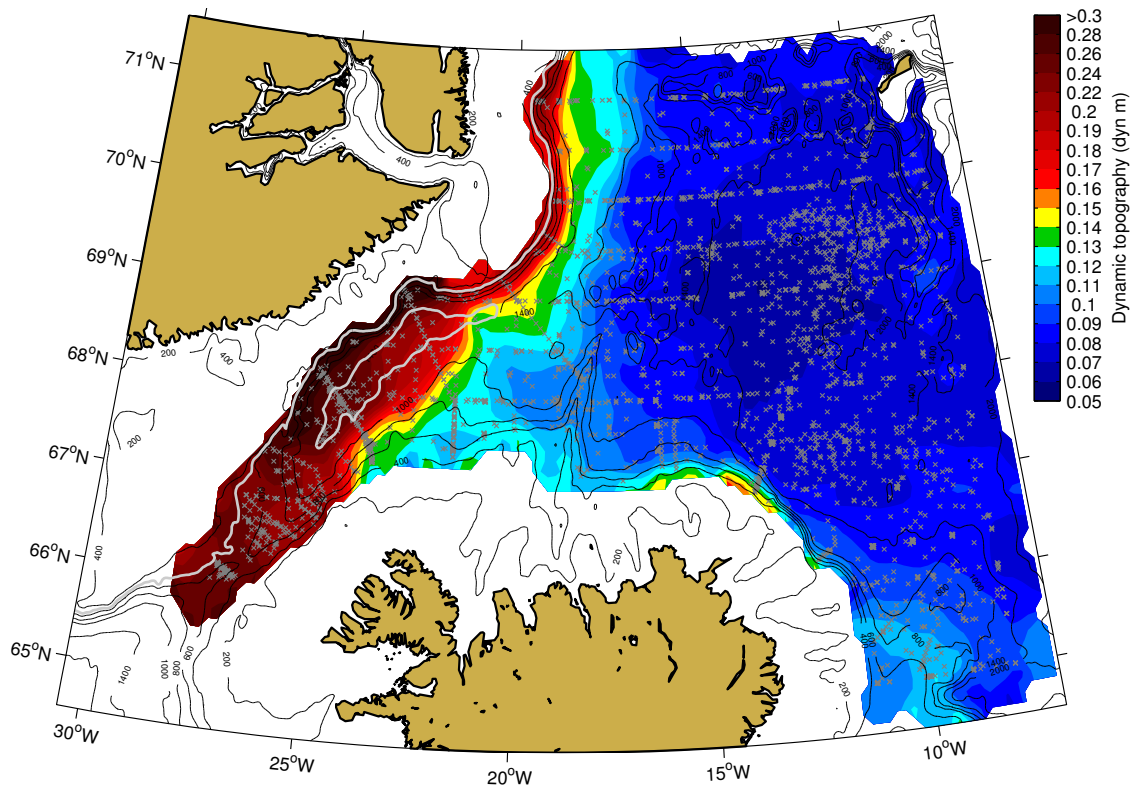


Figure 9: Dynamic height of the surface relative to 500 db. The gray lines are the 500 m depth contour along the east Greenland shelf break and the closed 1400 m depth contour delineating the Blosseville Basin. The gray crosses mark the locations of data points. The 200 m, 400 m, 600 m, 800 m, 1000 m, 1400 m, and 2000 m isobaths are contoured as black lines.

415 water transported by the EGC in this region extends some distance offshore (Fig. 11). The tem-
 416 perature and salinity maxima indicate, however, that the core of the overflow water is found along
 417 the continental slope. (The strongest signature of the Atlantic-origin overflow water is found in the
 418 density range $27.9 \text{ kg/m}^3 < \sigma_\theta < 28.0 \text{ kg/m}^3$, used to isolate this water mass in Figure 11.) South
 419 of 69°N , temperature and salinity maxima along the slope are no longer visible, and Atlantic-
 420 origin overflow water is observed throughout the Blosseville Basin. The hydrographic properties
 421 of the overflow water in this region have been modified, probably by the same process forming the
 422 separated EGC.

423 The historical hydrography supports the inference from the synoptic Kögur sections that the
 424 separated EGC is a permanent feature in the Blosseville Basin and that it provides a means of shelf-
 425 basin exchange impacting the circulation of both surface and overflow water masses north of the
 426 Denmark Strait. However, it is difficult to discern the details of how these water masses approach

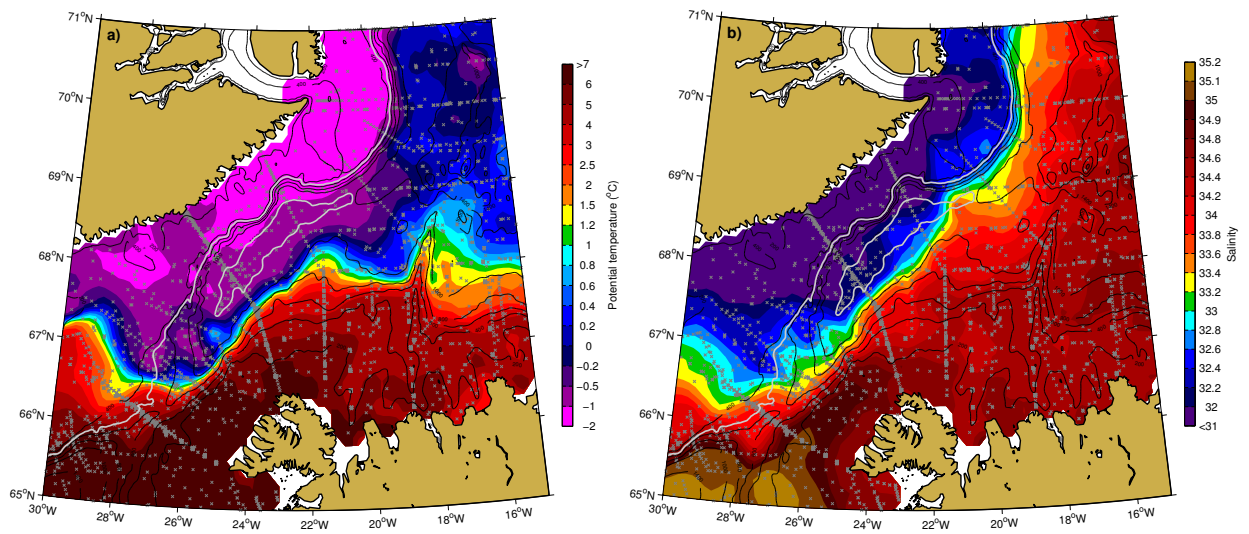


Figure 10: As in Figure 9, but for near-surface potential temperature (a, vertically averaged between 50 and 100 m) and salinity (b, vertically averaged between 10 and 30 m).

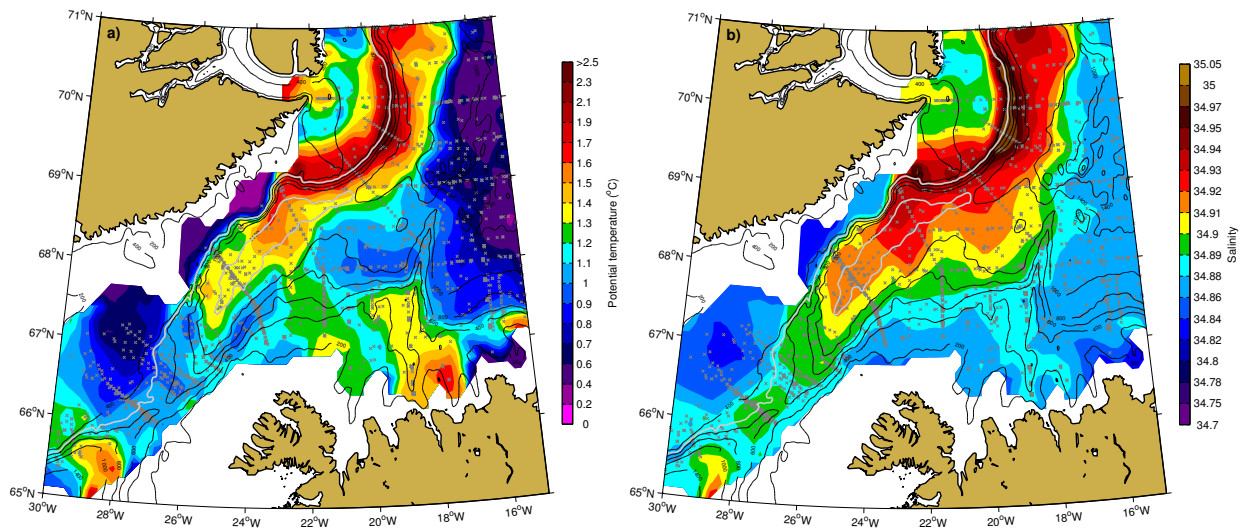


Figure 11: As in Figure 10, but for overflow waters (maximum value between 27.9 and 28.0 kg/m³).

427 and pass through the Denmark Strait from this analysis. In the vicinity of the sill warm, saline AW
428 from the Irminger Current dominates the surface layer, even extending at times onto the Greenland
429 shelf, and undiluted PSW appears to be found only near the coast. In the case of the overflow layer
430 the two-branch EGC system and the NIJ, observed as distinct pathways at the Kögur transect near
431 the southern end of the Blosseville Basin, merge prior to forming the DSOW plume exiting the
432 Denmark Strait. The relative transports of these overflow pathways likely influence the final water
433 mass composition of the plume (e.g. Rudels et al., 2003). In the next two sections we investigate
434 the processes by which the separated EGC is formed, which will enable us to better understand the
435 associated impacts on the overflow product and on the freshwater budget of the region.

436 **5. Atmospheric forcing**

437 The atmospheric circulation over the western subpolar North Atlantic is dominated by the
438 Icelandic Low, a semi-permanent region of low pressure situated southwest of Iceland in the lee
439 of southern Greenland (Serreze et al., 1997). The cyclonic circulation around the Icelandic Low
440 and the associated positive wind stress curl can be seen in Figure 12. The curl attains its maximum
441 value to the east of Greenland's southernmost point, Cape Farewell, as a result of the increased
442 surface wind speeds in this region that are associated with westerly tip jets (Doyle and Shapiro,
443 1999; Moore and Renfrew, 2005; Våge et al., 2009). The low Froude number of the surface
444 circulation, resulting from the high and steep topography of Greenland (Fig 12), leads to so called
445 barrier winds along its east coast that are characterized by southerly flow directed parallel to the
446 coast (Moore and Renfrew, 2005; Harden et al., 2011; Moore et al., 2012).

447 As can be seen from Figure 12a, surface wind speeds tend to be lowest near the coastline and
448 over sea ice as a result of an increase in surface roughness (Moore, 2003; Petersen and Renfrew,
449 2009). Due to this reduction in wind speed from the nearshore ice, a narrow band of negative
450 wind stress curl extends along the entire east coast of Greenland that is embedded in the gener-
451 ally positive wind stress curl of the broader subpolar North Atlantic (Fig. 12b). In the southeast
452 portion of Greenland there are two regions of particularly steep coastal topography coupled with
453 strong curvature of the coastline: One near 69°N and the other near 66°N. The former is associ-
454 ated with the Watkins Range, that contains Greenland's highest mountain, and is in the vicinity
455 of the Blosseville Basin. In these regions, there is an acceleration of the barrier winds because
456 the flow is being forced to move around the obstacles (Harden and Renfrew, 2012; Moore, 2012).
457 This localized flow distortion enhances the anti-cyclonic circulation. The combination of the lo-

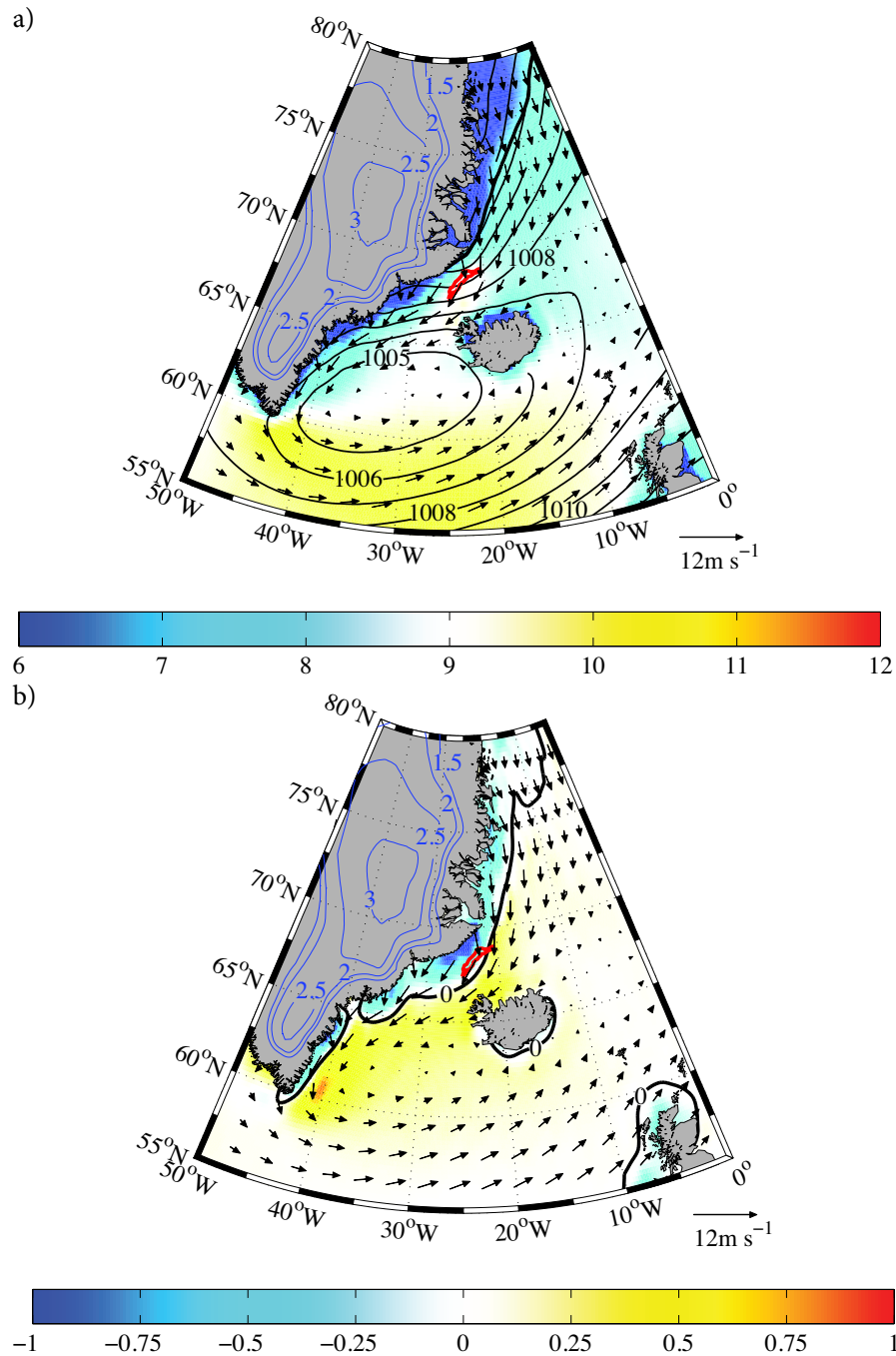


Figure 12: The annual mean (a) sea-level pressure (contours, mb), 10 m wind (vectors, m/s) and 10 m wind speed (color, m/s); and (b) curl of the wind stress (color, 10^{-6} N/m^2) and the 10 m wind (vectors, m/s) from the ERA-I for the period 1979-2011. In (a) the thick black line represents the annual mean 50% sea ice concentration contour. In (b) the zero isoline of the curl of the wind stress is indicated by the thick black contour. The red contour delineates the approximate extent of the Blossville Basin as represented by the 1400 m isobath. The blue contours represent the height of the topography over Greenland (km).

458 cal topographic steering together with the reduction in nearshore wind speed due to ice causes a
459 particularly strong region of negative wind stress curl over the Blosseville Basin (Fig. 13).

460 The sea ice along east Greenland expands southwards during the winter and retreats north-
461 wards during the summer (Wadhams, 1981), and so the coastal region of low surface wind speed
462 undergoes a similar cycle. Also, the Icelandic Low is deepest during the winter months and so the
463 barrier winds are intensified during this period. Consequently, the negative wind stress curl over
464 the Blosseville Basin is strongest in fall/winter (Fig. 13a) and weakest in spring/summer (Fig 13b).
465 Note, however, that the curl remains negative even when the circulation is weak, and, as a result,
466 the annual mean wind stress curl is negative in this region (Fig. 13c).

467 The strong negative wind stress curl over the Blosseville Basin, in conjunction with the closed
468 bathymetric contours of the basin, is conducive for spinning up a local anti-cyclonic ocean gyre.
469 The model simulation of Spall and Pickart (2003) demonstrated that positive wind stress curl east
470 of Cape Farewell (Fig. 12b) was capable of driving the cyclonic Irminger Gyre (Våge et al., 2011a).
471 This was true even though the wind forcing nearly vanishes during the summer months. The weak
472 stratification of the Irminger Sea, in conjunction with the bathymetry of the continental slope,
473 resulted in a “flywheel” effect whereby the seasonal input of vorticity from the atmosphere to the
474 ocean was able to maintain a nearly steady gyre. At depth, the Blosseville Basin is characterized
475 by similarly weak stratification adjacent to the continental slopes of Greenland and Iceland, and
476 the atmospheric circulation imparts strong negative vorticity to the ocean for nearly half the year.
477 As such, there is reason to suspect that an anti-cyclonic gyre should be maintained in the basin,
478 consistent with the mean absolute geostrophic velocity section of Figure 4.

479 As demonstrated in the next section, there is an additional aspect of the wind forcing that
480 appears to be of importance for the formation of the separated EGC. To first order the barrier winds
481 parallel the shelf break of East Greenland, and, as such, there is an onshore Ekman transport in the
482 surface layer. This shoreward flow helps maintain the hydrographic front associated with the shelf
483 break EGC. As noted above, the sharp bend in the coastline near 69°N steers the wind towards
484 the southwest (which is part of the reason for the negative wind stress curl over the Blosseville
485 Basin). However, because the orientation of the continental slope changes so abruptly near 69°N,
486 the wind cannot adjust quickly enough to remain parallel to the shelf break at this location. This is
487 demonstrated in Fig. 14 which documents the degree to which the wind parallels the 500 m isobath
488 (roughly the shelf break) at each latitude in the domain of interest. One sees that at the northern
489 end of the Blosseville Basin the winds are more than 40° offset from the direction of the shelf

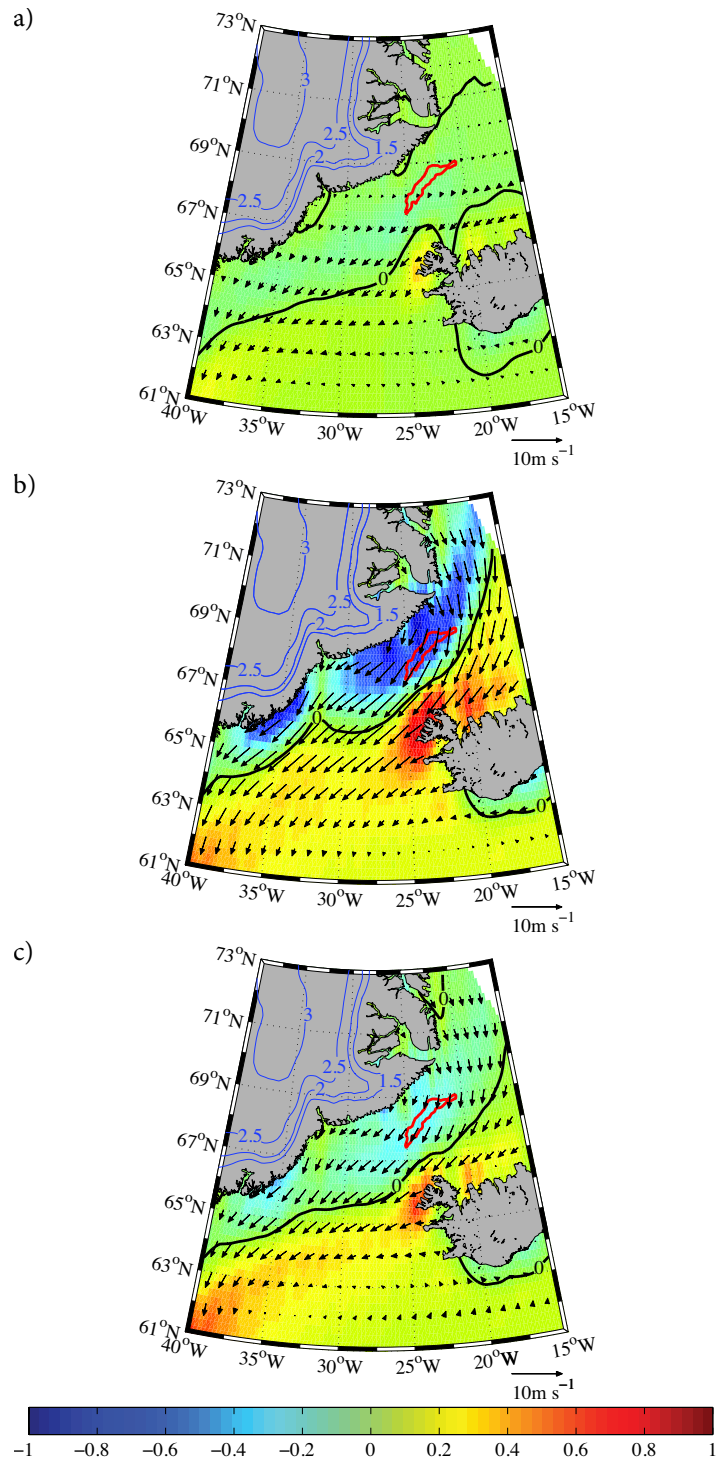


Figure 13: The curl of the wind stress (color, 10^{-6} N/m^2) and the 10 m wind (vectors, m/s) for (a) October; (b) June; and (c) the annual mean from the NARR for the period 1979-2011. The zero isoline of the curl of the wind stress is indicated by the thick black contour. The red contour delineates the approximate extent of the Blosseville Basin as represented by the 1400 m isobath. The blue contours represent the height of the topography over Greenland (km).

490 break. Consequently, the onshore component of the Ekman transport is reduced in this region. The
491 ramifications of this are explored below.

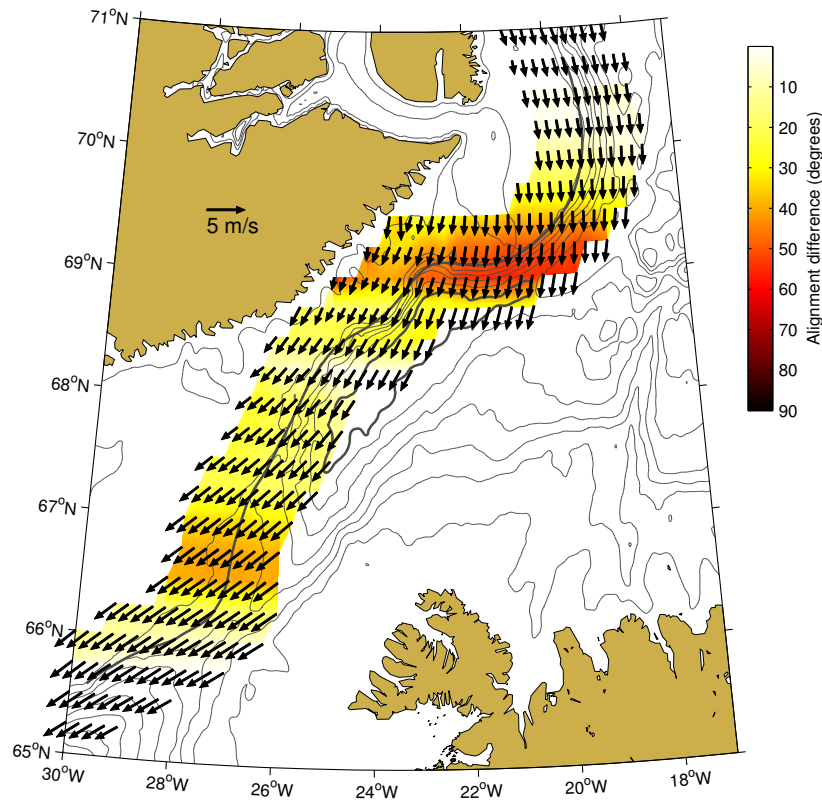


Figure 14: Difference in angle between the mean wind direction and orientation of the shelf break (color) as a function of latitude. The shelf break is taken to be the 500 m isobath, and the calculation is carried out along a 50 km swath centered around that isobath. The mean 10 m wind vectors from NARR, used in the calculation, are shown.

492 **6. Numerical simulation of the East Greenland Current north of the Denmark Strait**

493 We now use the idealized numerical model, described in Section 2.4, to examine the interaction
494 between the East Greenland Current and the basin interior, and to investigate the cause of the
495 separated EGC. In an effort to understand the most basic aspects of this interaction, we force the
496 model with a steady wind stress (no surface forcing of heat and freshwater). The wind forcing is
497 derived from the annual mean of the high resolution NARR product (Section 5), and the model
498 has been run for a period of 2.5 years. The mean sea surface salinity over the final 2 years of
499 integration is shown in Figure 15. There is a sharp gradient in salinity near the shelf break at high
500 latitudes (although there is some spreading of saline waters onto the shelf between $y = 500$ km

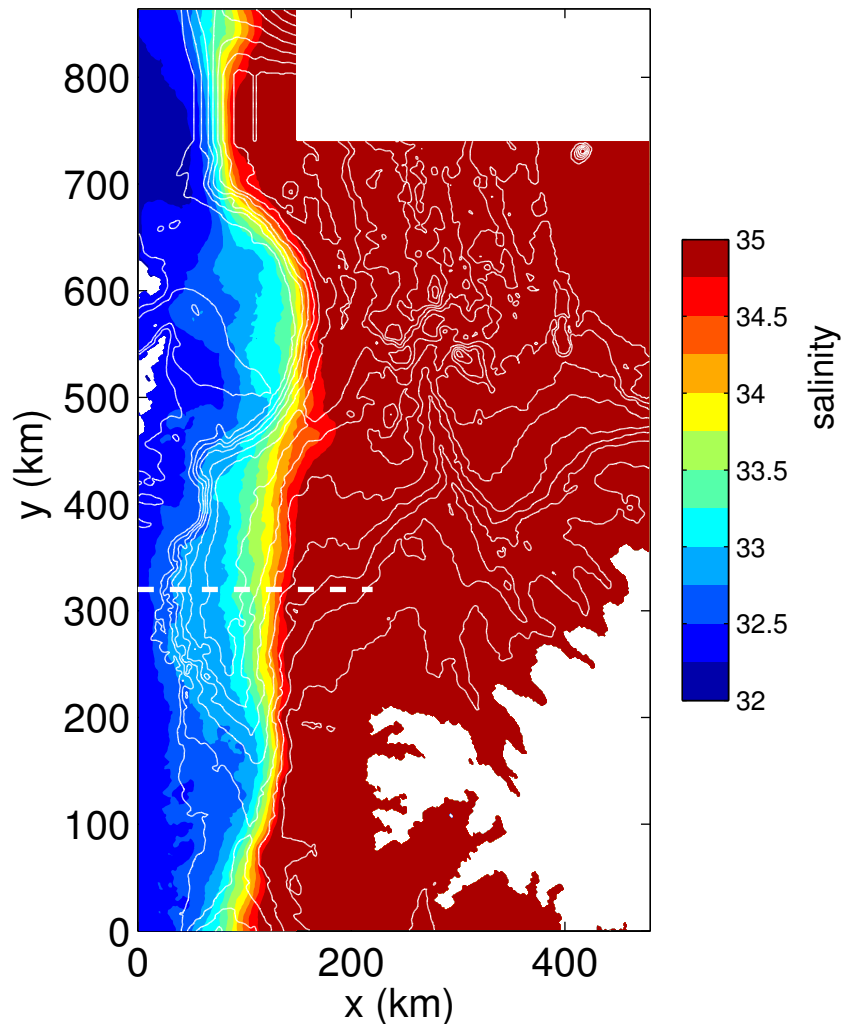


Figure 15: Mean sea surface salinity over final 2 years of integration, with bottom topography (white contours, contour interval = 200 m). The dashed white line is the location of the vertical section in Figure 16.

501 and 700 km). This gradient corresponds to a maximum in along-shelf velocity, i.e. a shelf break
 502 current. However, near $y = 500$ km the freshwater begins to shift offshore of the shelf break
 503 such that by $y = 300$ km waters fresher than 33.5 are found almost 100 km offshore of the shelf
 504 break. This freshwater remains offshore of the shelf edge all the way to the Denmark Strait (near
 505 $y = 100$ km, $x = 100$ km in the model).

506 6.1. Salinity and velocity

507 A representative synoptic section of the along-channel velocity and salinity from the model
 508 at $y = 320$ km are shown in Figure 16 for day 360. This location roughly corresponds with the

509 Kögur section. There is a maximum southward velocity of just over 1 m s^{-1} over the shelf break,
510 which decreases towards the bottom (Fig. 16a). There is also southward flow exceeding 50 cm s^{-1}
511 centered about 100 km offshore of the shelf break, over the eastern side of the Blosseville Basin.
512 This flow is more surface trapped, with a vertical scale only on the order of 200 m. In between
513 these two southward flows is a northward current of 50 cm s^{-1} near $x = 80 \text{ km}$. While it is tempting
514 to associate the southward flow above the Iceland continental slope with the separated EGC, the
515 two oppositely flowing jets in the interior are in fact the signature of an anti-cyclonic ring of shelf
516 water that separated from the shelf and carried freshwater across the Blosseville Basin (Fig. 16b).
517 This structure is reminiscent of some of the synoptic sections along the Kögur line (e.g. Figs. 5c
518 and 6c) and suggests that the features seen in the data may be large anti-cyclonic rings of shelf
519 water (i.e. larger than the deformation radius).

520 The sea surface salinity on day 770 (Fig. 17) demonstrates that the shelf break jet south of
521 $y = 500 \text{ km}$ is very time-dependent, and is dominated by meanders and eddies with horizontal
522 scales of $O(50 \text{ km})$. These freshwater eddies penetrate well off the boundary, giving rise to the
523 offshore shift in the mean position of the sea surface salinity front. The region between the shelf
524 break and the eastern side of the Blosseville Basin is highly time dependent and dominated by
525 eddies and filaments.

526 The spin-up of the offshore front, and the space and time scales of the variability, are demon-
527 strated by a plot of along-channel velocity and salinity at 17.5 m depth as a function of x and
528 time (Fig. 18). The fresh water initially confined to the shelf spreads rapidly offshore until around
529 day 200, when the offshore front equilibrates near $x = 120 \text{ km}$. Coincident with this salinity gra-
530 dient is a region of southward flow with strength $O(10 - 20 \text{ cm s}^{-1})$. There is also a region
531 of stronger southward flow shoreward of $x = 50 \text{ km}$, which is the meandering shelf break jet.
532 The region between the shelf break jet and the offshore front is dominated by flow reversals of
533 $O(20 \text{ cm s}^{-1})$. They generally occur in concert with southward flow farther offshore, separated
534 by a freshwater anomaly. This is the signature of freshwater lenses that have been shed from the
535 boundary current and have propagated offshore, as seen in Figures 16 and 17. The dashed line
536 at day 360 in Figure 18 shows that the features seen in the synoptic section (Fig. 16) are quite
537 common.

538 The model suggests then that the separated EGC arises from eddies that coalesce when they
539 encounter the Iceland continental slope. In this scenario the northward flow between the shelf break
540 EGC and separated EGC is simply the recirculation associated with freshwater eddies shed from

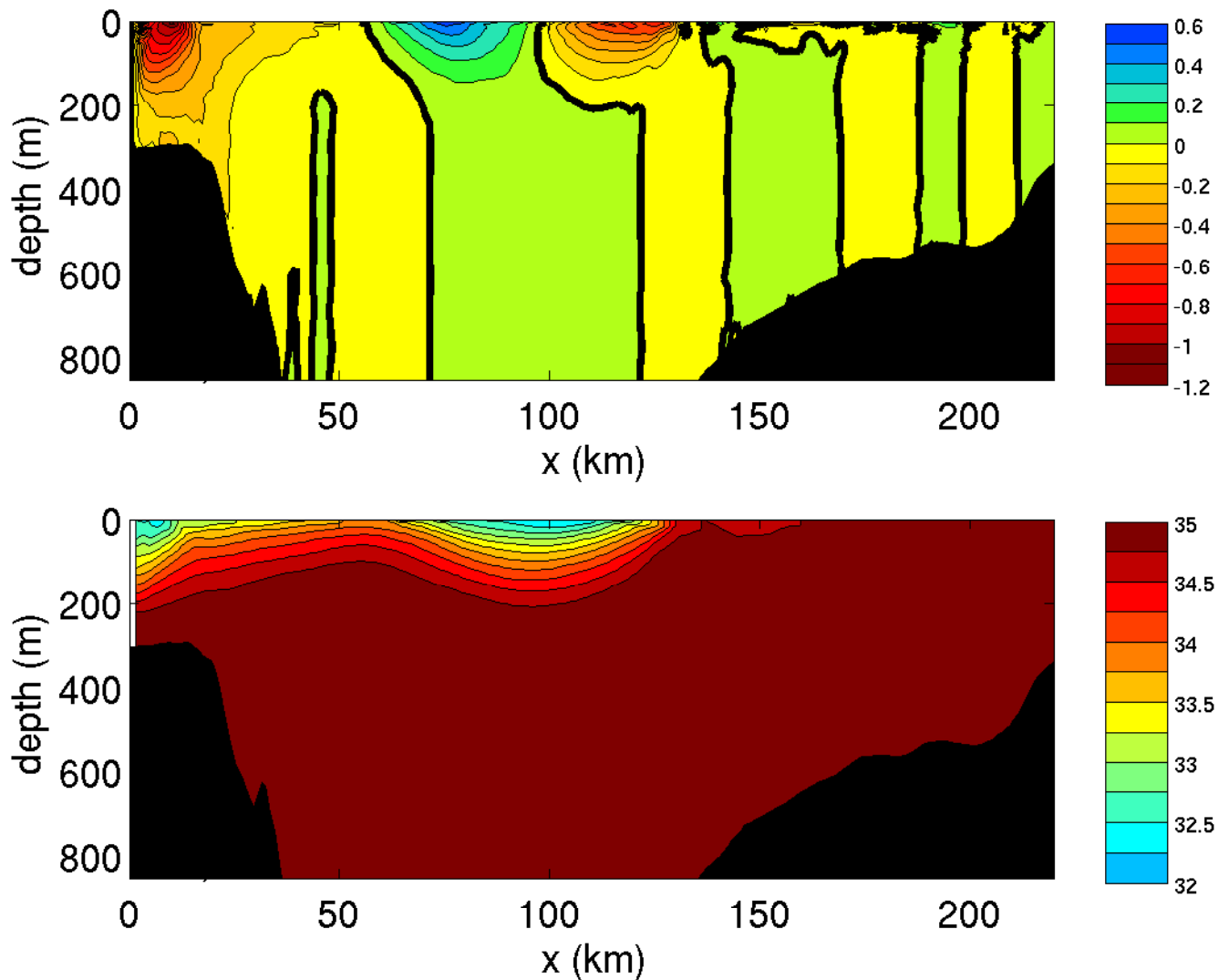


Figure 16: Synoptic zonal sections of meridional velocity (upper panel, c.i. = 0.1 m s^{-1} , bold line is the zero velocity contour) and salinity (lower panel, c.i. = 0.25) on day 360.

541 the shelf break jet. Furthermore, the model suggests that synoptic sections across the Blosseville
 542 Basin should occasionally reveal isolated anti-cyclonic eddies, which, as noted above, seems to be
 543 the case. The alternate hypothesis of a wind-driven anti-cyclonic gyre, where the offshore branch
 544 is the separated EGC, is not supported by the model.³ However, one must keep in mind that the
 545 data are spatially and temporally sparse, which makes it difficult to distinguish between these two

³Note that the model does have anti-cyclonic wind stress curl over the Blosseville Basin (Fig. 13c).

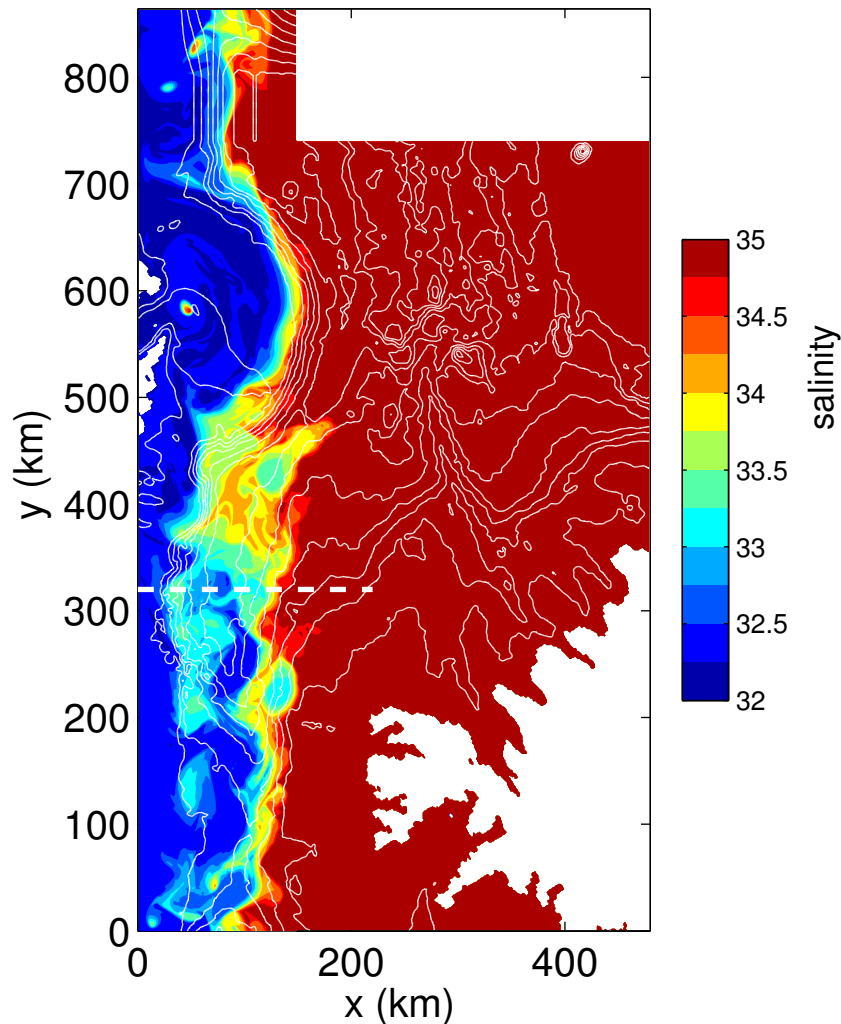


Figure 17: Sea surface salinity on day 770, with bottom topography (white contours, c.i. = 200 m).

546 possibilities; a definitive conclusion will require more field data.

547 6.2. Role of the wind

548 Although the curl of the wind stress seems not to play a central role in the model, the wind
 549 stress itself is important in maintaining the shelf break jet and in determining where the freshwater
 550 is able to penetrate offshore of the boundary. A calculation was run with the same initial conditions
 551 and restoring but with no wind stress. Freshwater extends farther offshore than for the case with
 552 wind (not shown). As noted earlier, for the most part the wind is parallel to the coast (Fig. 14), so
 553 the Ekman transport is directed onshore. This advects dense, saline water towards the shelf break

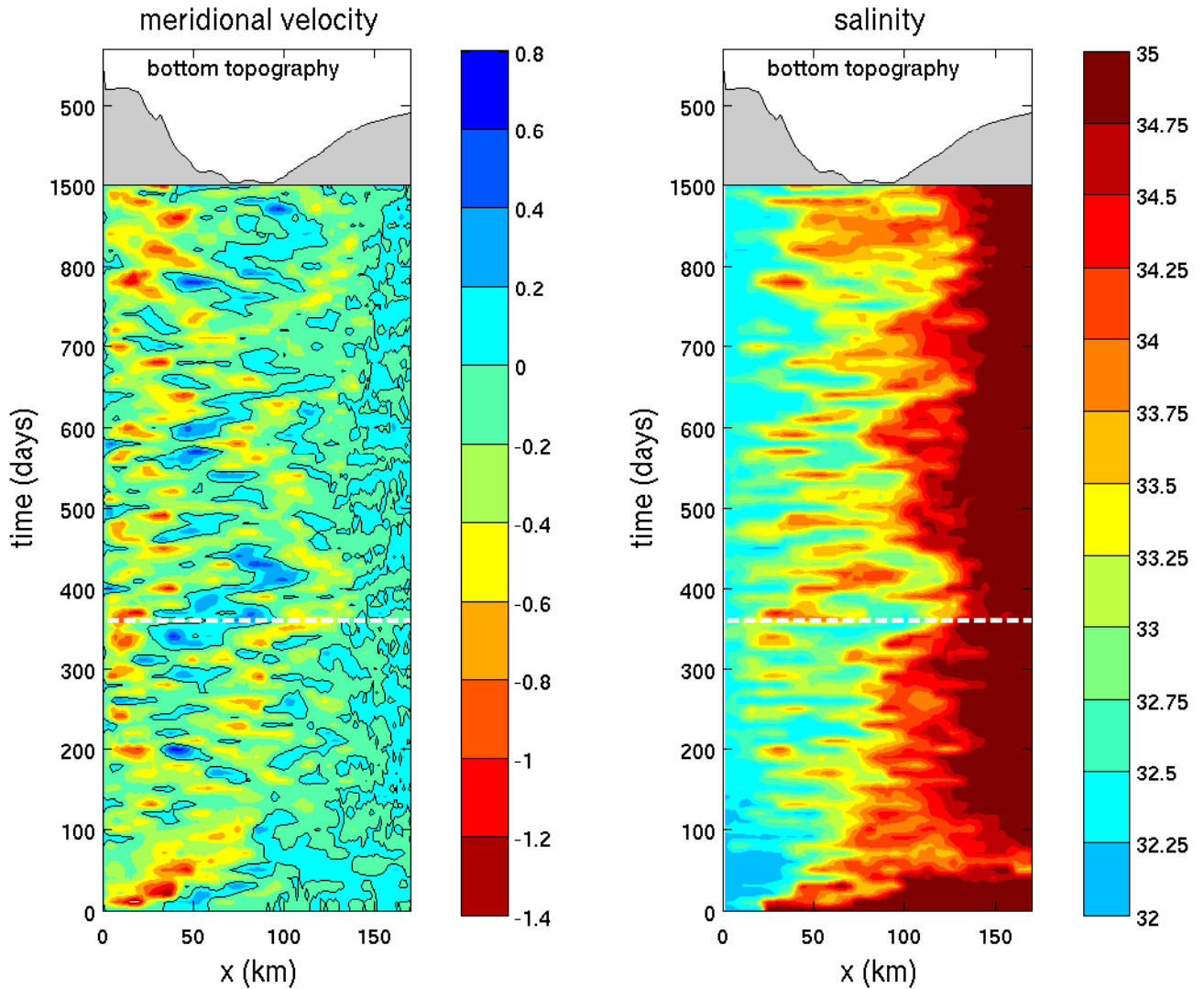


Figure 18: Time/x plots of meridional velocity (left panel, c.i. = 0.1 m s^{-1} , bold line is the zero velocity contour) and salinity (right panel, c.i. = 0.25) at 20 m depth and $y = 320 \text{ km}$. The bottom topography at this location is indicated at the top. The dashed white line is day 360, when the sections in Figure 16 were taken.

554 and acts to maintain the baroclinicity of the jet. Frontal instability acts to reduce the horizontal
 555 density gradient by advecting freshwater offshore near the surface. The absence of wind forcing
 556 thus allows the eddy fluxes to carry the fresh water farther offshore.

557 Analysis of the energy conversion terms indicates that the meanders and eddies are formed
 558 primarily by baroclinic instability. The conversion rate from potential to eddy kinetic energy was
 559 calculated in the region of the Blosseville Basin (20 to 150 km in x and 200 to 500 km in y) and

560 averaged between 10 and 540 m depth. For the case with wind forcing, the average conversion
561 rate is positive at about $1.2 \times 10^{-8} \text{ m}^2\text{s}^{-3}$. This corresponds with eddy formation via baroclinic
562 instability. The case with no wind starts at a level above the case with wind forcing, but rapidly
563 decreases by about an order of magnitude. This weak level of energy conversion persists for the
564 remainder of the calculation. This is consistent with the model result that, in the absence of wind,
565 the early growth of instabilities spreads the freshwater more effectively over the basin interior,
566 reducing the potential energy in, and instability of, the shelf break front. In contrast, when the
567 wind is present the shelf break front is maintained by a local balance between frontogenesis by
568 onshore Ekman transport and frontolysis by baroclinic instability. This allows for a more continued
569 extraction of energy by the eddies from the front.

570 The influence of the Ekman transport on the offshore transport of freshwater is further demon-
571 strated by calculations with an idealized coastline. In this scenario, the shelf topography is repre-
572 sented by two straight regions offset by 80 km (Fig. 19). The first bend in the continental slope,
573 near 650 km, is meant to represent the change in the orientation of the East Greenland shelfbreak
574 near 69°N (Fig. 2), whereas the second bend farther to the south (in the opposite direction) is nec-
575 essary to smoothly join the upper and lower boundaries of the model (as was done for the earlier
576 model). The initial conditions for this simulation were specified the same as for the cases described
577 above, and no restoring of salinity was used.

578 The sea surface salinity on day 100 is shown for a case with a uniform meridional wind stress
579 of amplitude -0.05 Nm^{-2} (Fig. 19a) and for an otherwise identical calculation with no wind
580 (Fig. 19b). The shelf break front develops meanders in both cases, but the offshore transport is
581 confined to the region where the topography is not parallel to the wind stress for the case with
582 wind, while it is more uniformly distributed along the shelf break jet in the absence of wind. This
583 demonstrates the importance of the abrupt change in orientation of the boundary north of the Blos-
584 seville Basin and the fact that the wind cannot adjust quickly enough to remain parallel to the shelf
585 break in this region (Fig. 14). Note that, for the case with wind, eddies also develop where the to-
586 pography bends back to the east near 250 km. This indicates that it is not simply inertial overshoot
587 that causes the anti-cyclones to form at the northern end of the Blosseville Basin. Frontal instabil-
588 ities are less inhibited where the Ekman transport is not perpendicular to the front, supporting the
589 previous interpretation. One also sees in Figure 19 that the salinity front is located farther offshore
590 everywhere in the case with no wind, which is consistent with the previous model results using the
591 realistic topography and wind stress.

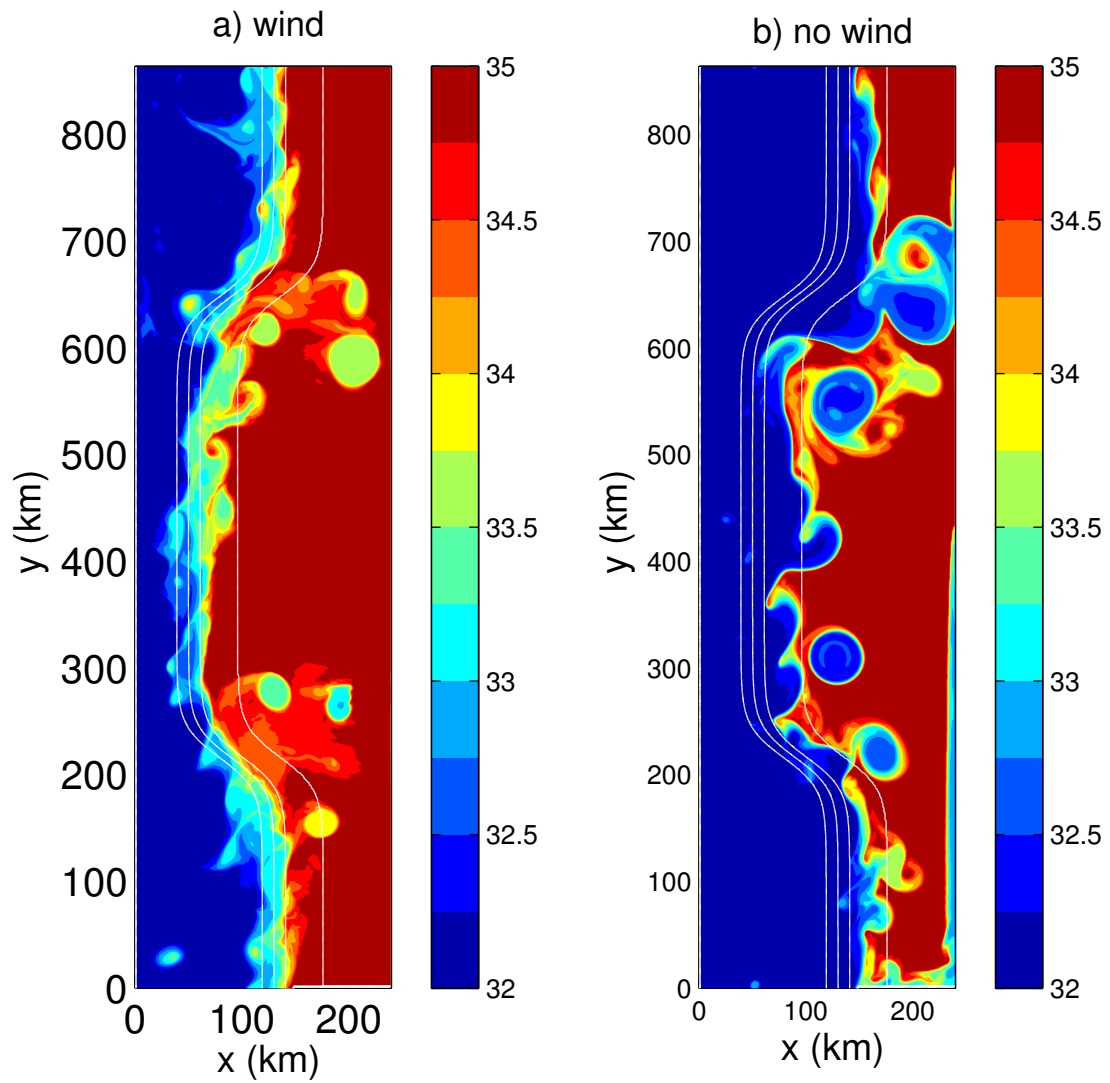


Figure 19: Sea surface salinity (color) on day 100 for an idealized shelf. The white contours indicate the bathymetry. a) The case of uniform meridional wind stress of -0.05 Nm^{-2} . b) The case of no wind. The offshore freshwater flux is enhanced in the two regions where the wind is not parallel to the shelfbreak in a) compared to the case with no wind in b).

592 7. Discussion

593 Our results show that the separated EGC provides a means for transporting freshwater as well
594 as overflow water from the western boundary into the interior of the Blosseville Basin. What is
595 the fate of the freshwater? Two direct pathways of freshwater export from the EGC into the inte-
596 rior Nordic seas, the Jan Mayen and East Icelandic currents, have been previously identified (e.g.
597 Dickson et al., 2007). While the Jan Mayen Current is to some extent topographically steered by
598 the Jan Mayen Fracture Zone (Bourke et al., 1992), Figure 2 shows no corresponding bathymetric
599 feature farther south that would cause the East Icelandic Current to diverge from the EGC. Sparse
600 current meter measurements at the Kögur transect led Jónsson (1999) to conclude that a partial
601 recirculation of the EGC in the Denmark Strait was not the source of the East Icelandic Current
602 and that it instead originates north of the transect. The East Icelandic Current has been traced
603 upstream only as far as the Spar Fracture Zone (a gap in the Kolbeinsey Ridge north of Iceland,
604 Fig. 2; Jónsson, 2007; Jónsson and Valdimarsson, 2012a). It is possible then that the mechanism
605 generating the separated EGC in the northern part of the Blosseville Basin is also the source of
606 freshwater to the East Icelandic Current. If so, a portion of the freshwater exported off the Green-
607 land shelf in the Blosseville Basin would be advected by the East Icelandic Current into the Iceland
608 Sea. The resulting reduction in surface salinity and increase in stratification could in turn reduce
609 the extent of wintertime convection in the Iceland Sea Gyre, with possible consequences for the
610 AMOC. Another ramification of the freshwater export from the boundary is that less of the undi-
611 luted PSW remains within the shelf break EGC equatorward of the Denmark Strait (Fig. 10) which
612 could potentially impact the convective regions in the Irminger and Labrador seas.

613 Our revised circulation scheme is shown schematically in Figure 20. Included in the figure are
614 findings from the present study as well as results from the study of Våge et al. (2011b). As seen in
615 the schematic, and at odds with the Mauritzen (1996) circulation scheme, the source waters of the
616 Denmark Strait Overflow Water plume primarily approach the sill along the Iceland continental
617 slope. These waters are advected in roughly equal proportions by the separated EGC and the
618 North Icelandic Jet (NIJ). By contrast, the shelf break EGC appears to transport a significantly
619 smaller fraction (roughly 30%) of overflow water along the Greenland slope. This is perhaps to be
620 expected within the framework of hydraulic theory (e.g. Pratt, 2004). However, the NIJ originates
621 far upstream (northeast of Iceland, Fig. 20) and the separated EGC transposes to the Iceland slope
622 in the northern end of the Blosseville Basin, not due to shoaling bathymetry in the vicinity of
623 the sill. At this point it remains uncertain how much water (if any) recirculates anti-cyclonically

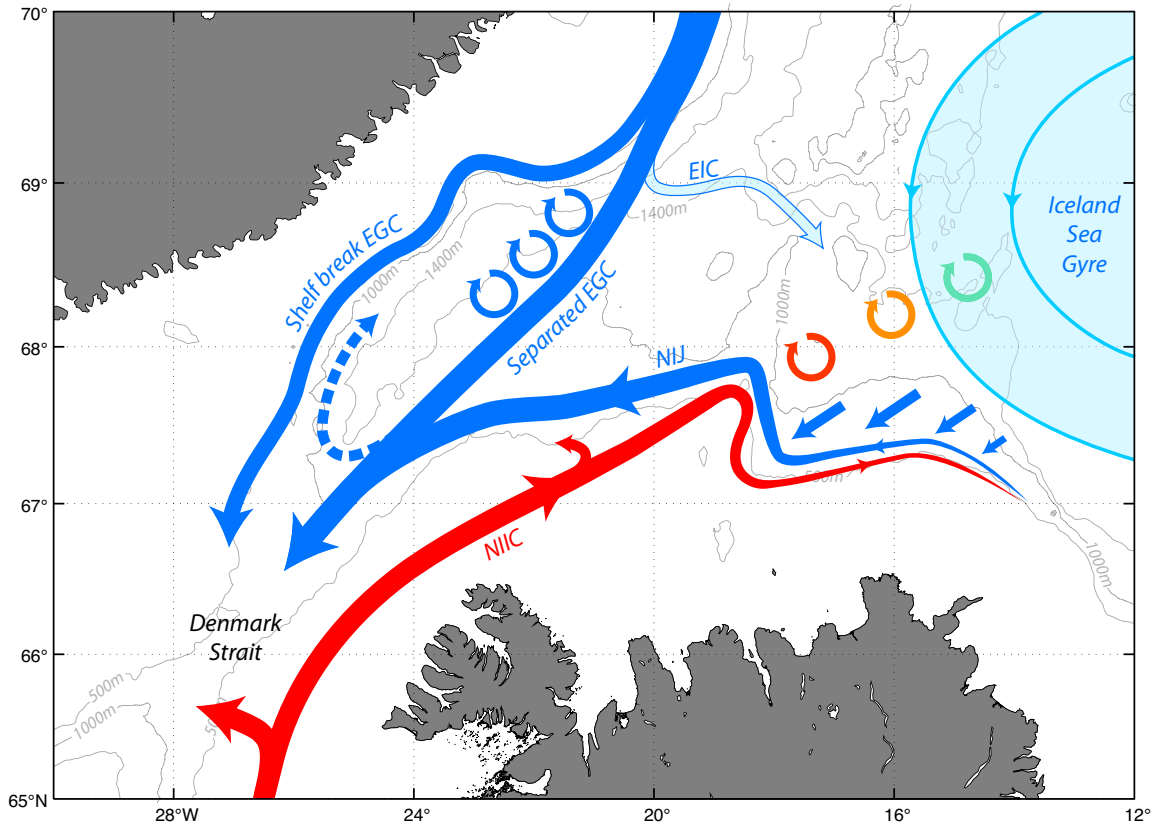


Figure 20: Schematic circulation in the area northeast of the Denmark Strait, presented in the text. The East Greenland Current (EGC) bifurcates north of the Blossville Basin and the offshore branch joins with the North Icelandic Jet (NIJ) to provide most of the dense water feeding the Denmark Strait Overflow Water plume. The shelf break EGC provides the other portion. The separated EGC is believed to be formed by anti-cyclonic eddies that coalesce, with perhaps a wind-driven anti-cyclonic recirculation north of the sill (dashed line). As discussed in Våge et al. (2011b), the NIJ represents the lower limb of a local overturning loop: The inflowing North Icelandic Irminger Current – advecting warm Atlantic Water – forms eddies that are cooled by the atmosphere and disintegrate in the Iceland Sea Gyre. The dense water so formed progresses back towards the boundary (represented by the short blue arrows) and sinks to form the NIJ. A possible pathway of the upper-layer East Icelandic Current (EIC) is indicated as well.

624 from the separated EGC north of the sill or whether or not there is a permanent gyre within the
625 Blosseville Basin. A recently deployed mooring array along the Kögur transect should shed light
626 on this.

627 We note that the high-resolution numerical simulations of the flow north of the Denmark Strait
628 in Köhl et al. (2007) have aspects that are similar to the circulation scheme presented here. For
629 example, the EGC bifurcates upstream of the Blosseville Basin. However, the eastern branch of the
630 EGC in their model flows southward along the Kolbeinsey Ridge and feeds the NIJ far upstream
631 of the Denmark Strait, which is inconsistent with our results. Nonetheless, Köhl et al. (2007)'s
632 simulations indicate that most of the overflow water in the East Greenland Current switches from
633 the Greenland to the Iceland slope at the northern end of the Blosseville Basin and that the overflow
634 water primarily approaches the Denmark Strait sill along the eastern boundary.

635 Between the Kögur transect and the Denmark Strait sill the three branches advecting overflow
636 water (the shelf break EGC, separated EGC, and the NIJ) presumably merge to form the DSOW
637 plume. For the NIJ and separated EGC, which are governed by different dynamics and transport
638 different water masses (mid-depth intensified flow of Arctic-origin overflow water and surface-
639 intensified flow advecting Atlantic-origin overflow water, respectively), the merging process may
640 lead to the generation of instabilities upstream of the sill (e.g. Fristedt et al., 1999). Such pre-
641 existing instabilities may subsequently be amplified during the descent of the overflow plume (e.g.
642 Spall and Price, 1998).

643 At present it is unknown which, if any, of these flow branches feeding the DSOW plume exerts
644 dominant control on the variability of the overflow. Numerical simulations suggest that changes in
645 the EGC north of the strait impact the transport and composition of the overflow plume (Karcher
646 et al., 2011; Hall et al., 2011). However, comparable studies remain to be undertaken for the
647 recently discovered NIJ. The mooring array across the Kogur line includes instruments in all three
648 branches, which may provide some insights along these lines. More detailed tracer studies will
649 also add to our understanding of the dominant pathways and upstream sources of overflow water.
650 We stress that the circulation scheme presented here needs further confirmation, particularly to
651 elucidate the merging of the different overflow branches and the fate of the freshwater exported
652 into the interior.

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665 **Appendix A. Removal of barotropic tides from the synoptic sections using a tidal model**

666 The most recent version of the Oregon State University Atlantic Ocean tidal model (Egbert
667 et al., 1994; Egbert and Erofeeva, 2002) was used to remove the barotropic tidal component from
668 the ADCP profiles prior to the construction of the absolute geostrophic velocity sections. The
669 improvements relative to the previous version used by Våge et al. (2011b) are higher resolution
670 ($1/30^\circ$ vs. $1/12^\circ$) and more accurate bathymetry. Comparison between the two versions at the
671 Kögur transect shows that the greatest difference is found in the western end of the section, pri-
672 marily due to the improved representation of the Greenland continental shelf and slope. For the
673 Iceland slope the difference is less pronounced. The mean absolute difference in the NIJ trans-
674 port resulting from de-tiding using the two models was 0.1 Sv, which is well within the overall
675 transport error estimates. In particular, the value of NIJ transport for the October 2008 occupation
676 quoted by Våge et al. (2011b) differs by only 0.03 Sv when the new tidal model is applied. Hence
677 the results presented here using the new model are consistent with previously published results in
678 terms of NIJ transport. A reduction in the error arising from the de-tiding is, however, not justified
679 from an overall quantitative comparison between the tidal model and actual bottom depths along
680 the Kögur transect.

681 **Appendix B. Quality control and gridding of the historical hydrography**

682 While most of the data have been subject to preliminary quality control, additional checks were
683 performed following the general procedure of previous studies (e.g. Skagseth and Mork, 2012).
684 Temperature and salinity measurements outside the expected range of values in the Nordic seas (-2-
685 20°C and 20-36, respectively) were discarded. Each profile was subsequently inspected for density
686 inversions, and profiles containing an inversion exceeding 0.05 kg/m³ were excluded (Rossby et al.,
687 2009; Skagseth and Mork, 2012). (Profiles with a single data spike were included after the removal
688 of the spike.) Finally, each profile was checked for outliers as follows. All profiles within an
689 effective radius of 110 km around the station in question (approximately one degree of latitude)
690 were identified. The effective radius is increased along isobaths in regions of large topographic
691 gradients and takes into account the greater correlation length scales along the bottom topography,
692 which is appropriate given the close alignment between the circulation in the Nordic seas and the
693 bottom contours (e.g. Nøst and Isachsen, 2003). The radius was calculated following Davis (1998):

$$r^2 = |\mathbf{x}_g - \mathbf{x}_o|^2 + \left| 3\lambda \frac{H_g - H_o}{H_g + H_o} \right|^2. \quad (\text{B.1})$$

694 The first term on the right hand side of (B.1) is the geographical distance between the profile to
695 be checked (subscript g) and all other profiles (subscript o), and the second term is the increase
696 in distance determined by the difference in bottom depths (H). The topographic parameter λ was
697 set to 100 km (Lavender et al., 2005; Voet et al., 2010; Skagseth and Mork, 2012). A doubling in
698 water depth between two profiles would lead to an increase in r^2 by λ^2 . Bathymetric data were
699 obtained from the ETOPO2 2-minute elevation data base and smoothed by convolution with a
700 20-km Gaussian window. All of the profiles so identified were then vertically interpolated at 5 m
701 intervals, and the mean and standard deviation of temperature and salinity calculated at each depth.
702 If the profile in question contained data points at any depth that differed from the mean by more
703 than six standard deviations, it was discarded.

704 For the present analysis, observations from the historical hydrographic data set of a given
705 property at a given depth level were anisotropically interpolated onto a regular 0.2° longitude by
706 0.1° latitude grid using (B.1). For each grid point (subscript g) the effective distance to each data
707 point (subscript o) was calculated. The average value of all data points within an effective distance
708 of 50 km or less, weighted by the inverse of the distance (profiles closer to the grid point than 1 km
709 were weighted equally), was assigned to that grid point. Using an effective distance increased
710 by the difference in barotropic potential vorticity ($PV = f/H$, where f is the Coriolis parameter)

711 instead of difference in depth (e.g. Böhme and Send, 2005) yielded qualitatively similar results.
712 Finally, the gridded fields were smoothed by convolution with a 30 km Gaussian window.

713 **Bibliography**

714 **References**

715 Aagaard, K., Carmack, E. C., 1989. The role of sea ice and other fresh water in the Arctic circula-
716 tion. *Journal of Geophysical Research* 94, 14485–14498.

717 Adcroft, A., Hill, C., Marshall, J., 1997. Representation of topography by shaved cells in a height
718 coordinate ocean model. *Monthly Weather Review* 125, 2293–2315.

719 Böhme, L., Send, U., 2005. Objective analyses of hydrographic data for referencing profil-
720 ing float salinities in highly variable environments. *Deep Sea Research II* 52, 651–664,
721 doi:10.1016/j.dsr2.2004.12.014.

722 Bourke, R. H., Paquette, R. G., Blythe, R. F., 1992. The Jan Mayen Current of the Greenland Sea.
723 *Journal of Geophysical Research* 97, 7241–7250.

724 Cooper, L. H. N., 1955. Deep water movements in the North Atlantic as a link between climate
725 changes around Iceland and the biological productivity of the English Channel and the Celtic
726 Sea. *Journal of Marine Research* 14, 347–362.

727 Davis, R. E., 1998. Preliminary results from directly measuring middepth circulation in the tropical
728 and South Pacific. *Journal of Geophysical Research* 103, 24619–24639.

729 de Steur, L., Hansen, E., Gerdes, R., Karcher, M., Fahrbach, E., Holfort, J., 2009. Freshwater
730 fluxes in the East Greenland Current: A decade of observations. *Geophysical Research Letters*
731 36, L23611, doi:10.1029/2009GL041278.

732 Dee, D. P., Uppala, S. M., Simmons, A. J., Berrisford, P., Poli, P., Coauthors, 2011. The ERA-
733 Interim reanalysis: configuration and performance of the data assimilation system. *Quarterly*
734 *Journal of the Royal Meteorological Society* 137, 553–597, doi:10.1002/qj.828.

735 Dickson, R. R., Brown, J., 1994. The production of North Atlantic Deep Water: Sources, rates and
736 pathways. *Journal of Geophysical Research* 99, 12319–12341.

- 737 Dickson, R. R., Dye, S., Jónsson, S., Köhl, A., Macrander, A., Marnela, M., Meincke, J., Olsen,
738 S., Rudels, B., Valdimarsson, H., Voet, G., 2008. The overflow flux west of Iceland: Variability,
739 origins and forcing. In: Dickson, R. R., Meincke, J., Rhines, P. (Eds.), *Arctic-Subarctic Ocean*
740 *Fluxes: Defining the role of the northern seas in climate*. Springer, Dordrecht, The Netherlands,
741 pp. 443–474.
- 742 Dickson, R. R., Rudels, B., Dye, S., Karcher, M., Meincke, J., Yashayaev, I., 2007. Current es-
743 timates of freshwater flux through Arctic and subarctic seas. *Progress in Oceanography* 73,
744 210–230.
- 745 Dodd, P. A., Heywood, K. J., Meredith, M. P., Naveira-Garabato, A. C., Marca, A. D., Falkner,
746 K. K., 2009. Sources and fate of freshwater exported in the East Greenland Current. *Geophysical*
747 *Research Letters* 36, L19608, doi:10.1029/2009GL039663.
- 748 Dodd, P. A., Rabe, B., Hansen, E., Falck, E., Mackensen, A., Rohling, E., Stedmon, C., Kris-
749 tiansen, S., 2012. The freshwater composition of the Fram Strait outflow derived from a decade
750 of tracer measurements. *Journal of Geophysical Research* 117, doi:10.1029/2012JC008011.
- 751 Doyle, J. D., Shapiro, M. A., 1999. Flow response to large-scale topography: The Greenland tip
752 jet. *Tellus* 51, 728–748.
- 753 Egbert, G. D., Bennett, A. F., Foreman, M. G. G., 1994. TOPEX/Poseidon tides estimated using a
754 global inverse model. *Journal of Geophysical Research* 99, 24821–24852.
- 755 Egbert, G. D., Erofeeva, S. Y., 2002. Efficient inverse modeling of barotropic ocean tides. *Journal*
756 *of Atmospheric and Oceanic Technology* 19, 183–204.
- 757 Eldevik, T., Nilsen, J. E. Ø., Iovino, D., Olsson, K. A., Sandø, A. B., Drange, H., 2009.
758 Observed sources and variability of Nordic seas overflow. *Nature Geoscience* 2, 406–410,
759 doi:10.1038/NGEO518.
- 760 Fristedt, T., Hietala, R., Lundberg, P., 1999. Stability properties of a barotropic surface-water jet
761 observed in the Denmark Strait. *Tellus* 51, 979–989.
- 762 Hall, S., Dye, S. R., Heywood, K. J., Wadley, M. R., 2011. Wind forcing of salinity anomalies in
763 the Denmark Strait overflow. *Ocean Science* 7, 821–834, doi:10.5194/os-7-821-2011.

- 764 Hansen, B., Østerhus, S., 2000. North Atlantic–Nordic Seas exchanges. *Progress in Oceanography*
765 45, 109–208.
- 766 Hansen, B., Østerhus, S., 2007. Faroe Bank Channel overflow 1995–2001. *Progress in Oceanog-*
767 *raphy* 75, 817–856, doi:10.1016/j.pocean.2007.09.004.
- 768 Harden, B. E., Renfrew, I. A., 2012. On the spatial distribution of high winds off southeast Green-
769 land. *Geophysical Research Letters* 39, L14806, doi:10.1029/2012GL052245.
- 770 Harden, B. E., Renfrew, I. A., Petersen, G. N., 2011. A climatology of wintertime barrier winds
771 off southeast Greenland. *Journal of Climate* 24, 4701–4717, doi:10.1175/2011JCLI4113.1.
- 772 Holfort, J., Albrecht, T., 2007. Atmospheric forcing of salinity in the overflow of Denmark Strait.
773 *Ocean Science* 3, 411–416.
- 774 Holfort, J., Hansen, E., Østerhus, S., Dye, S., Jónsson, S., Meincke, J., Mortensen, J., Meredith,
775 M., 2008. Freshwater fluxes east of Greenland. In: Dickson, R. R., Meincke, J., Rhines, P. (Eds.),
776 *Arctic-Subarctic Ocean Fluxes: Defining the role of the northern seas in climate*. Springer, Dor-
777 drecht, The Netherlands, pp. 263–287.
- 778 Holfort, J., Meincke, J., 2005. Time series of freshwater-transport on the East Greenland Shelf at
779 74°N. *Meteorologische Zeitschrift* 14, 703–710, doi:10.1127/0941–2948/2005/0079.
- 780 Jeansson, E., Jutterström, S., Rudels, B., Anderson, L. G., Olsson, K. A., Jones, E. P., Jr., W. M. S.,
781 Swift, J. H., 2008. Sources to the East Greenland Current and its contribution to the Denmark
782 Strait Overflow. *Progress in Oceanography* 78, 12–28, doi:10.1016/j.pocean.2007.08.031.
- 783 Jochumsen, K., Quadfasel, D., Valdimarsson, H., Jónsson, S., 2012. Variability of the Denmark
784 Strait overflow: Moored time series from 1996–2011. *Journal of Geophysical Research*, in press.
- 785 Jones, E. P., Anderson, L. G., Jutterström, S., Swift, J. H., 2008. Sources and distribu-
786 tion of fresh water in the East Greenland Current. *Progress in Oceanography* 78, 37–44,
787 doi:10.1016/j.pocean.2007.06.003.
- 788 Jónsson, S., 1999. The circulation in the northern part of the Denmark Strait and its variability.
789 ICES report CM-1999/L:06, 9 pp.

- 790 Jónsson, S., 2007. Volume flux and fresh water transport associated with the East Icelandic Current.
791 *Progress in Oceanography* 73, 231–241, doi:10.1016/j.pocean.2006.11.003.
- 792 Jónsson, S., Valdimarsson, H., 2004. A new path for the Denmark Strait overflow wa-
793 ter from the Iceland Sea to Denmark Strait. *Geophysical Research Letters* 31, L03305,
794 doi:10.1029/2003GL019214.
- 795 Jónsson, S., Valdimarsson, H., 2005. Recent developments in oceanographic research in Icelandic
796 waters. In: Caseldine, C., Russell, A., Harðardóttir, J., Knudsen, O. (Eds.), *Iceland – Modern
797 processes and past environments*. Elsevier, pp. 79–92.
- 798 Jónsson, S., Valdimarsson, H., 2012a. Hydrography and circulation over the southern part of the
799 Kolbeinsey Ridge. *ICES Journal of Marine Science*, doi:10.1093/icesjms/fss101.
- 800 Jónsson, S., Valdimarsson, H., 2012b. Water mass transport variability to the north Icelandic shelf,
801 1994-2010. *ICES Journal of Marine Science*, doi:10.1093/icesjms/fss024.
- 802 Karcher, M., Beszczynska-Möller, A., Kauker, F., Gerdes, R., Heyen, S., Rudels, B., 2011. Arctic
803 Ocean warming and its consequences for the Denmark Strait overflow. *Journal of Geophysical
804 Research* 116, doi:10.1029/2010JC006265.
- 805 Käse, R. H., Serra, N., Köhl, A., Stammer, D. B., 2009. Mechanisms for the variability of
806 dense water pathways in the Nordic Seas. *Journal of Geophysical Research* 114, C01013,
807 doi:10.1029/2008JC004916.
- 808 Kinder, T. H., Parrilla, G., 1987. Yes, some of the Mediterranean outflow does come from great
809 depth. *Journal of Geophysical Research* 92, 2901–2906.
- 810 Köhl, A., 2010. Variable source regions of Denmark Strait and Faroe Bank Channel overflow
811 waters. *Tellus* 62A, 551–568, doi:10.1111/j.1600-0870.2010.00454.x.
- 812 Köhl, A., Käse, R. H., Stammer, D. B., 2007. Causes of changes in the Denmark Strait Overflow.
813 *Journal of Physical Oceanography* 37, 1678–1696.
- 814 Kwok, R., Cunningham, G. F., Pang, S. S., 2004. Fram Strait sea ice outflow. *Journal of Geophys-
815 ical Research* 109, C01009, doi:10.1029/2003JC001785.

- 816 Lavender, K. L., Owens, W. B., Davis, R. E., 2005. The mid-depth circulation of the subpolar
817 North Atlantic Ocean as measured by subsurface floats. *Deep Sea Research I* 52, 767–785.
- 818 Marshall, J., Hill, C., Perelman, L., Adcroft, A., 1997. Hydrostatic, quasi-hydrostatic, and nonhy-
819 drostatic ocean modeling. *Journal of Geophysical Research* 102, 5733–5752.
- 820 Mauritzen, C., 1996. Production of dense overflow waters feeding the North Atlantic across the
821 Greenland-Scotland Ridge. Part 1: Evidence for a revised circulation scheme. *Deep Sea Re-*
822 *search I* 43, 769–806.
- 823 Mesinger, F., DiMego, G., Kalnay, E., Mitchell, K., Shafran, P. C., Coauthors, 2006. North Amer-
824 ican regional reanalysis. *Bulletin of the American Meteorological Society* 87, 343–360.
- 825 Moore, G. W. K., 2003. Gale force winds over the Irminger Sea to the east of Cape Farewell,
826 Greenland. *Geophysical Research Letters* 30, 1894, doi:10.1029/2003GLO18012.
- 827 Moore, G. W. K., 2012. A new look at Greenland flow distortion and its impact on barrier flow, tip
828 jets and coastal oceanography. *Geophysical Research Letters*, doi:10.1029/2012GL054017.
- 829 Moore, G. W. K., Pickart, R. S., Renfrew, I. A., 2008. Buoy observations from the windiest loca-
830 tion in the world ocean, Cape Farewell, Greenland. *Geophysical Research Letters* 35, L18802,
831 doi:10.1029/2008GL034845.
- 832 Moore, G. W. K., Renfrew, I. A., 2005. Tip jets and barrier winds: A QuikSCAT climatology of
833 high wind speed events around Greenland. *Journal of Climate* 18, 3713–3725.
- 834 Moore, G. W. K., Renfrew, I. A., Pickart, R. S., 2012. Spatial distribution of air-sea heat
835 fluxes over the sub-polar North Atlantic Ocean. *Geophysical Research Letters* 39, L18806,
836 doi:10.1029/2012GL053097.
- 837 Nansen, F., 1912. Das Bodenwasser und die Abkühlung des Meeres. *Internationale Revue der*
838 *gesamten Hydrobiologie und Hydrographie* Band V, 1–42.
- 839 Nilsen, J. E. Ø., Hátún, H., Mork, K. A., Valdimarsson, H., 2008. The NISE data set. Tech. Rep.
840 08-01, Faroese Fisheries Laboratory, Box 3051, Tórshavn, Faroe Islands.
- 841 Nilsson, J., Björk, G., Rudels, B., Winsor, P., Torres, D. J., 2008. Liquid freshwater transport
842 and Polar Surface Water characteristics in the East Greenland Current during the AO-02 Oden
843 expedition. *Progress in Oceanography* 78, 45–57, doi:10.1016/j.pocean.2007.06.002.

- 844 Nøst, O. A., Isachsen, P. E., 2003. The large-scale time-mean ocean circulation in the Nordic Seas
845 and Arctic Ocean estimated from simplified dynamics. *Journal of Marine Research* 61, 175–210.
- 846 Olsson, K. A., Jeansson, E., Tanhua, T., Gascard, J.-C., 2005. The East Greenland Current studied
847 with CFCs and released sulphur hexafluoride. *Journal of Marine Systems* 55, 77–95.
- 848 Petersen, G. N., Renfrew, I. A., 2009. Aircraft-based observations of air–sea fluxes over Denmark
849 Strait and the Irminger Sea during high wind speed conditions. *Quarterly Journal of the Royal
850 Meteorological Society* 135, 2030–2045, doi:10.1002/qj.355.
- 851 Pickart, R. S., Torres, D. J., Fratantoni, P. S., 2005. The East Greenland Spill Jet. *Journal of
852 Physical Oceanography* 35, 1037–1053.
- 853 Pratt, L. J., 2004. Recent progress on understanding the effects of rotation in models of sea straits.
854 *Deep Sea Research II* 51, 351–369.
- 855 Renfrew, I. A., Petersen, G. N., Sproson, D. A. J., Moore, G. W. K., Adiwidjaja, H., Zhang, S.,
856 North, R., 2009. A comparison of aircraft-based surface-layer observations over Denmark Strait
857 and the Irminger Sea with meteorological analyses and QuikSCAT winds. *Quarterly Journal of
858 the Royal Meteorological Society* 135, 2046–2066, doi: 10.1002/qj.444.
- 859 Rossby, T., Ozhigin, V., Ivshin, V., Bacon, S., 2009. An isopycnal view of the Nordic Seas hy-
860 drography with focus on properties of the Lofoten Basin. *Deep Sea Research I* 56, 1955–1971,
861 doi:10.1016/j.dsr.2009.07.005.
- 862 Rudels, B., Björk, G., Nilsson, J., Winsor, P., Lake, I., Nohr, C., 2005. The interaction between
863 waters from the Arctic Ocean and the Nordic Seas north of Fram Strait and along the East Green-
864 land Current: Results from the Arctic Ocean-02 Oden expedition. *Journal of Marine Systems*
865 55, 1–30.
- 866 Rudels, B., Eriksson, P., Buch, E., Budéus, G., Fahrbach, E., Malmberg, S.-A., Meincke, J.,
867 Mälkki, P., 2003. Temporal switching between sources of the Denmark Strait overflow water.
868 *ICES Marine Science Symposia* 219, 319–325.
- 869 Rudels, B., Fahrbach, E., Meincke, J., Budéus, G., Eriksson, P., 2002. The East Greenland Current
870 and its contribution to the Denmark Strait overflow. *ICES Journal of Marine Science* 59, 1133–
871 1154.

- 872 Segtnan, O. H., Furevik, T., Jenkins, A. D., 2011. Heat and freshwater budgets of the Nordic
873 seas computed from atmospheric reanalysis and ocean observations. *Journal of Geophysical*
874 *Research* 116, C11003, doi:10.1029/2011JC006939.
- 875 Serreze, M. C., Carse, F., Barry, R. G., Rogers, J. C., 1997. Icelandic Low cyclone activity: Clima-
876 tological features, linkages with the NAO, and relationships with recent changes in the Northern
877 Hemisphere circulation. *Journal of Climate* 10, 453–464.
- 878 Skagseth, Ø., Mork, K. A., 2012. Heat content in the Norwegian Sea, 1995-2010. *ICES Journal of*
879 *Marine Science*, doi:10.1093/icesjms/fss026.
- 880 Spall, M. A., Pickart, R. S., 2003. Wind-driven recirculations and exchange in the Labrador and
881 Irminger Seas. *Journal of Physical Oceanography* 33, 1829–1845.
- 882 Spall, M. A., Price, J. F., 1998. Mesoscale variability in Denmark Strait: The PV outflow hypoth-
883 esis. *Journal of Physical Oceanography* 28, 1598–1623.
- 884 Stommel, H., Bryden, H., Mangelsdorf, P., 1973. Does some of the Mediterranean outflow come
885 from great depth? *Pure and Applied Geophysics* 105, 879–889.
- 886 Sutherland, D. A., Pickart, R. S., 2008. The East Greenland Coastal Current: Structure, variability
887 and forcing. *Progress in Oceanography* 78, 58–77, doi:10.1016/j.pocean.2007.09.006.
- 888 Swift, J. H., Aagaard, K., 1981. Seasonal transitions and water mass formation in the Iceland and
889 Greenland seas. *Deep Sea Research* 28A, 1107–1129.
- 890 Swift, J. H., Aagaard, K., Malmberg, S.-A., 1980. The contribution of the Denmark Strait overflow
891 to the deep North Atlantic. *Deep Sea Research* 27A, 29–42.
- 892 Tanhua, T., Olsson, K. A., Jeansson, E., 2005. Formation of Denmark Strait overflow water and its
893 hydro-chemical composition. *Journal of Marine Systems* 57, 264–288.
- 894 Tanhua, T., Olsson, K. A., Jeansson, E., 2008. Tracer evidence of the origin and variability of
895 Denmark Strait Overflow Water. In: Dickson, R. R., Meincke, J., Rhines, P. (Eds.), *Arctic-*
896 *Subarctic Ocean Fluxes: Defining the role of the northern seas in climate*. Springer, Dordrecht,
897 The Netherlands, pp. 475–503.

- 898 Våge, K., Pickart, R. S., Sarafanov, A., Knutsen, Ø., Mercier, H., Lherminier, P., Bersch,
899 M., van Aken, H. M., Meincke, J., Quadfasel, D., Bacon, S., 2011a. The Irminger Gyre:
900 Circulation, convection, and interannual variability. *Deep Sea Research I* 58, 590–614,
901 doi:10.1016/j.dsr.2011.03.001.
- 902 Våge, K., Pickart, R. S., Spall, M. A., Valdimarsson, H., Jónsson, S., Torres, D. J., Østerhus, S.,
903 Eldevik, T., 2011b. Significant role of the North Icelandic Jet in the formation of Denmark Strait
904 Overflow Water. *Nature Geoscience* 4, 723–727, doi:10.1038/NGEO1234.
- 905 Våge, K., Spengler, T., Davies, H. C., Pickart, R. S., 2009. Multi-event analysis of the westerly
906 Greenland tip jet based upon 45 winters in ERA-40. *Quarterly Journal of the Royal Meteorolog-
907 ical Society* 135, 1999–2011, doi: 10.1002/qj.488.
- 908 Voet, G., Quadfasel, D., Mork, K. A., Sjøiland, H., 2010. The mid-depth circulation of the
909 Nordic Seas derived from profiling float observations. *Tellus* 62, 516–529, doi:10.1111/j.1600–
910 0870.2010.00444.x.
- 911 Wadhams, P., 1981. The ice cover in the Greenland and Norwegian Seas. *Reviews of Geophysics
912 and Space Physics* 19, 345–393.
- 913 Wong, A. P. S., Johnson, G. C., Owens, W. B., 2003. Delayed-mode calibration of autonomous
914 CTD float profiling salinity data by θ -S climatology. *Journal of Atmospheric and Oceanic Tech-
915 nology* 20, 308–318.