

1 High-frequency variability in the  
2 North Icelandic Jet

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**ABSTRACT**

7 We describe the high-frequency variability in the North Icelandic Jet (NIJ) on the Iceland  
8 Slope using data from the densely instrumented Kögur mooring array deployed upstream  
9 of the Denmark Strait sill from September 2011 to July 2012. Significant sub-8-day vari-  
10 ability is ubiquitous in all moorings from the Iceland slope with a dominant period of  
11 3.6 days. We attribute this variability to topographic Rossby waves on the Iceland slope  
12 with a wavelength of  $62 \pm 3$  km and a phase velocity of  $17.3 \pm 0.8$  km day<sup>-1</sup> directed  
13 downslope ( $-9^\circ\text{T}$ ). We test the theoretical dispersion relation for these waves against our  
14 observations and find good agreement between the predicted and measured direction of  
15 phase propagation. We additionally calculate a theoretical group velocity of 36 km day<sup>-1</sup>  
16 directed almost directly up-slope ( $138^\circ\text{T}$ ) which agrees well with the propagation speed  
17 and direction of observed energy pulses. We use an inverse wave tracing model to show  
18 that this wave energy is generated locally, offshore of the array, and does not emanate  
19 from the upstream or downstream directions along the Iceland slope. It is hypothesized  
20 that either the meandering Separated East Greenland Current located seaward of the NIJ,  
21 or intermittent aspiration of dense water into the Denmark Strait Overflow, are the drivers  
22 of the topographic waves.

## 23 1. Introduction

24 The Denmark Strait Overflow is the major pathway of dense water out of the Nordic  
25 Seas. It transports 3.2 Sv, or approximately 50%, of the total outflow (Dickson and Brown,  
26 1994; Jochumsen *et al.*, 2017), and hence plays a crucial role in the Atlantic meridional  
27 overturning circulation (AMOC). While the existence of this overflow has been known  
28 for many decades, our understanding of the processes that govern it and the underlying  
29 dynamics remains incomplete. One important aspect that requires further study is deter-  
30 mining the upstream sources of the dense water and how it approaches the sill. If we are to  
31 determine how a changing climate might impact the AMOC, we need to understand bet-  
32 ter the connection between the water mass transformation process and the flux of newly  
33 ventilated water to Denmark Strait.

34 Most of the Denmark Strait Overflow water (approximately 70%) comes from the  
35 East Greenland Current by way of the Nordic Seas boundary current system (Våge *et al.*,  
36 2013; Harden *et al.*, 2016) (see Figure 1). Specifically, warm Atlantic inflow across the  
37 Greenland-Scotland Ridge is progressively cooled as it flows northward towards Fram  
38 Strait, much of it recirculating in the strait and subducting to mid-depth (Mauritzen, 1996).  
39 This is joined by Atlantic water exiting the strait that has circumnavigated the Arctic, and  
40 together the transformed Atlantic water flows southward in the East Greenland Current.  
41 As the current rounds Scoresby Sund, it splits into two branches (Figure 1). One continues  
42 towards the sill as a shelfbreak jet (Håvik *et al.*, 2017). The other carries approximately  
43 60% of the East Greenland Current water out into the central strait via eddies and/or gyre-  
44 like deflections of the shelfbreak jet (Våge *et al.*, 2013; Harden *et al.*, 2016). This interior  
45 pathway, known as the separated East Greenland current, then flows into the strait along  
46 the outer Iceland slope.

47 The remaining 30% of Denmark Strait Overflow water is supplied by the North Ice-  
48 landic Jet (NIJ), a more recently discovered branch of the upstream circulation (Jonsson  
49 and Valdimarsson, 2004; Våge *et al.*, 2011). This mid-depth intensified jet advects waters

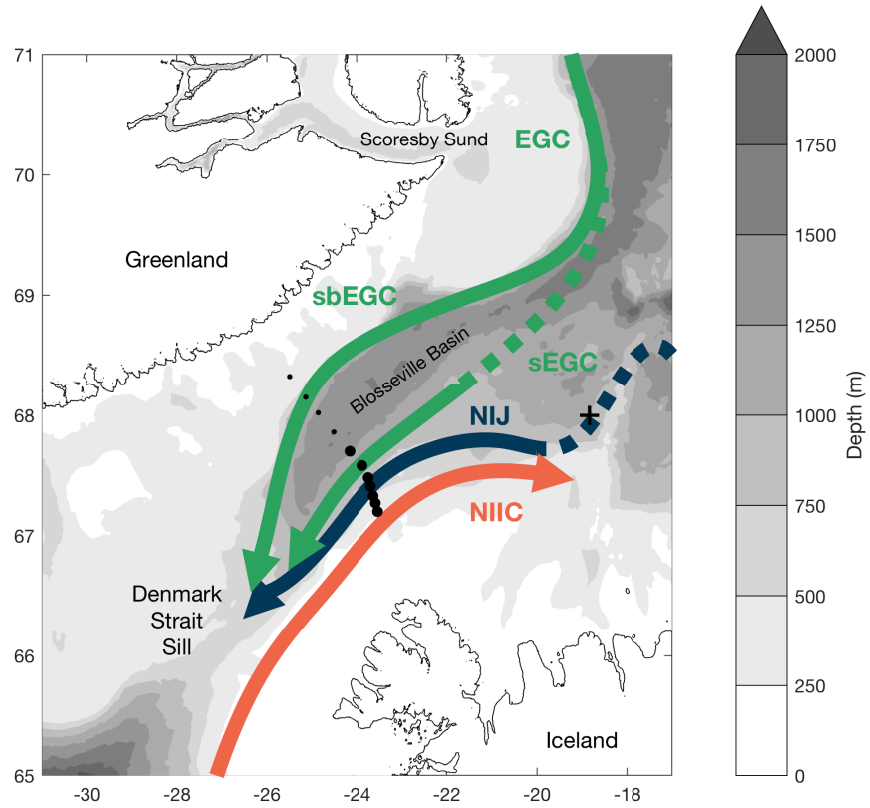


Figure 1: Map of the study region showing the overflow pathways approaching the Denmark Strait Sill: the North Icelandic Jet (NIJ) and the two East Greenland Current (EGC) pathways, one along the shelfbreak (sbEGC) and the other in a separated branch on the Iceland Slope (sEGC). Dashed portions show parts of pathways that still need further clarification. Also shown is the northward flowing surface-intensified current, the North Icelandic Irminger Current (NIIC). Black dots show the locations of the moorings in the Kögur array with larger dots indicating the subset of seven moorings used in this study. The upstream cross is the mooring to the west of the Kolbensey ridge referred to in the text. The bathymetry is from IBCAO v3.

50 distinct from those found in the East Greenland Current (colder and fresher) suggestive of  
51 a source in the central Iceland or Greenland seas (Våge *et al.*, 2011; 2015; Harden *et al.*,  
52 2016). The NIJ contains the densest water that feeds the overflow; its waters are found in  
53 the deepest part of the sill (Mastropole *et al.*, 2017) and subsequently sink to the deepest  
54 depths in the core of the overflow.

55 The leading hypothesis for the formation of the NIJ, supported by both models and  
56 observations, is that it represents the lower limb of a local overturning cell in the Iceland  
57 sea (Våge *et al.*, 2011; Behrens *et al.*, 2017). The upper limb of the cell is the NIIC, which  
58 sheds warm water into the Iceland Sea that is cooled by air-sea heat loss. The transformed  
59 water then returns southward towards the boundary where it sinks and forms the NIJ.  
60 However, many questions remain unanswered about this proposed system. For instance,  
61 the winter mixed-layers in the Iceland Sea don't appear to be dense enough to account for  
62 the deepest water in the NIJ (Våge *et al.*, 2015), whereas those in the Greenland Sea do  
63 (Strass *et al.*, 1993; Rudels *et al.*, 2002).

64 Regardless of the source of the NIJ, it clearly constitutes a vital component of the  
65 circulation upstream of the sill. Harden *et al.* (2016) investigated the jet's mean and sea-  
66 sonal contribution to the overflow, demonstrating that there is time-dependent partitioning  
67 of transport between the NIJ and the other two overflow branches on weekly to monthly  
68 timescales, likely driven by the wind. Pickart *et al.* (2017) noted that the NIJ appears to  
69 be coupled to the northward-flowing NIIC and that, on occasion, it consists of multiple  
70 branches. Using historical hydrographic data, Pickart *et al.* (2017) also revealed a clear  
71 link between the interannually varying properties of the NIJ and those of the densest water  
72 at the Denmark Strait sill, leaving little doubt that the NIJ is a major source of the overflow  
73 plume.

74 It has long been known that the Denmark Strait Overflow varies on short (order days)  
75 timescales (Smith, 1976; Bruce, 1995; Käse *et al.*, 2003). Some of this variability is  
76 associated with the passage of lenses of cold, dense, overflow water referred to as boluses

77 (Cooper, 1955). Recently, von Appen *et al.* (2017) identified a second type of mesoscale  
78 feature in the strait that was termed a pulse. In contrast to boluses, pulses correspond to  
79 a thinning of the overflow layer associated with a large increase in equatorward velocity.  
80 Both of these features have been identified in a high-resolution regional model as well  
81 (Almansi *et al.*, 2017). von Appen *et al.* (2017) showed that, taking into account both  
82 boluses and pulses, a mesoscale feature passes through Denmark Strait on average every  
83 2 days. Presently, however, it is unknown if these disturbances originate from upstream or  
84 if they are associated with local dynamics near the sill.

85 The goal of the present study is to shed light on some of the above processes by de-  
86 scribing the high frequency variability of the NIJ north of the Denmark Strait. We use  
87 timeseries data from a year-long mooring array that was maintained roughly 200 km up-  
88 stream of the sill (Figure 1). This is the same data set used by Harden *et al.* (2016) to inves-  
89 tigate the mean and seasonal attributes of the NIJ. While Harden *et al.* (2016) mentioned  
90 that the NIJ exhibits high-frequency variability, they did not elaborate on this. We begin  
91 with a brief description of the data, followed by a characterization of the high-frequency  
92 signal. We discuss how this signal is consistent with the existence of topographic Rossby  
93 waves on the Iceland slope, and then investigate the source region of the energy in these  
94 waves through inverse wave tracing.

## 95 **2. Data and Methods**

96 The data for this study come from the densely instrumented Kögur mooring array  
97 spanning the Denmark Strait approximately 200 km upstream of the sill. The array was  
98 deployed for 11 months from September 2011 to July 2012 and consisted of 12 moorings  
99 (named KGA 1-12) equipped with instrumentation to measure both the hydrography and  
100 velocity of the water column from 50 m to the bottom. Harden *et al.* (2016) present a  
101 detailed description of the mooring data, including the instrumentation, processing steps,  
102 and sensor accuracies. The array captured the majority of the overflow water (denser than

103 27.8 kg m<sup>-3</sup>) passing through the northern part of the strait towards the sill.

104 Here we use primarily the gridded product described in Harden *et al.* (2016), which  
105 has a lateral resolution of 8km and vertical resolution of 50 m. Because of our focus  
106 on the Iceland slope, we consider a subset of these data up to and including the location  
107 of mooring KGA 7, approximately 70 km offshore of the Iceland shelfbreak. The mean  
108 velocity sections demonstrate that this portion of the array captures both the NIJ and the  
109 majority of the Separated EGC (Figure 2). For parts of the analysis we also use the data  
110 on a mooring-by-mooring basis. All of the velocities have been de-tided using a 36-hour  
111 low-pass filter.

112 Additional data come from a mooring located approximately 200 km upstream of the  
113 Kögur Array on the west side of the Kolbesney Ridge (68°00'N, 18°50'W, see Figure  
114 1). This was deployed on the 1000 m isobath from September 2007 to mid-October 2008  
115 and consisted of a McLane Moored Profiler and acoustic current meter providing profiles  
116 between 100 m and the bottom at 8 hour intervals. As with the Kögur data, we low-passed  
117 the velocity timeseries using a 36-hr filter to remove the tidal components of the flow.  
118 These data are described in greater detail by Jónsson and Valdimarsson (2012).

119 The inverse wave tracing of topographic Rossby waves (TRWs) was done using the  
120 model described by Meinen *et al.* (1993) and implemented by Pickart (1995) for investigat-  
121 ing TRWs in the Deep Western Boundary Current off of Cape Hatteras, North Carolina.  
122 The method uses the TRW dispersion relation to calculate the group velocity and then  
123 backtracks the evolution of the wave with a time step of 30 minutes. The wave parameters  
124 are recalculated at each step for the local bottom depth, bottom slope, and water column  
125 stratification. A new group velocity is then found and used to further trace the wave. Most  
126 of the required input parameters for the inverse wave tracing model come directly from the  
127 moored data and are the same as those used for the theoretical TRW dispersion relation  
128 calculations (see Section 3.a.). For the bathymetry we used the International Bathymetric  
129 Chart of the Arctic Ocean 30-arcsec gridded product (Jakobsson *et al.*, 2012). To remove

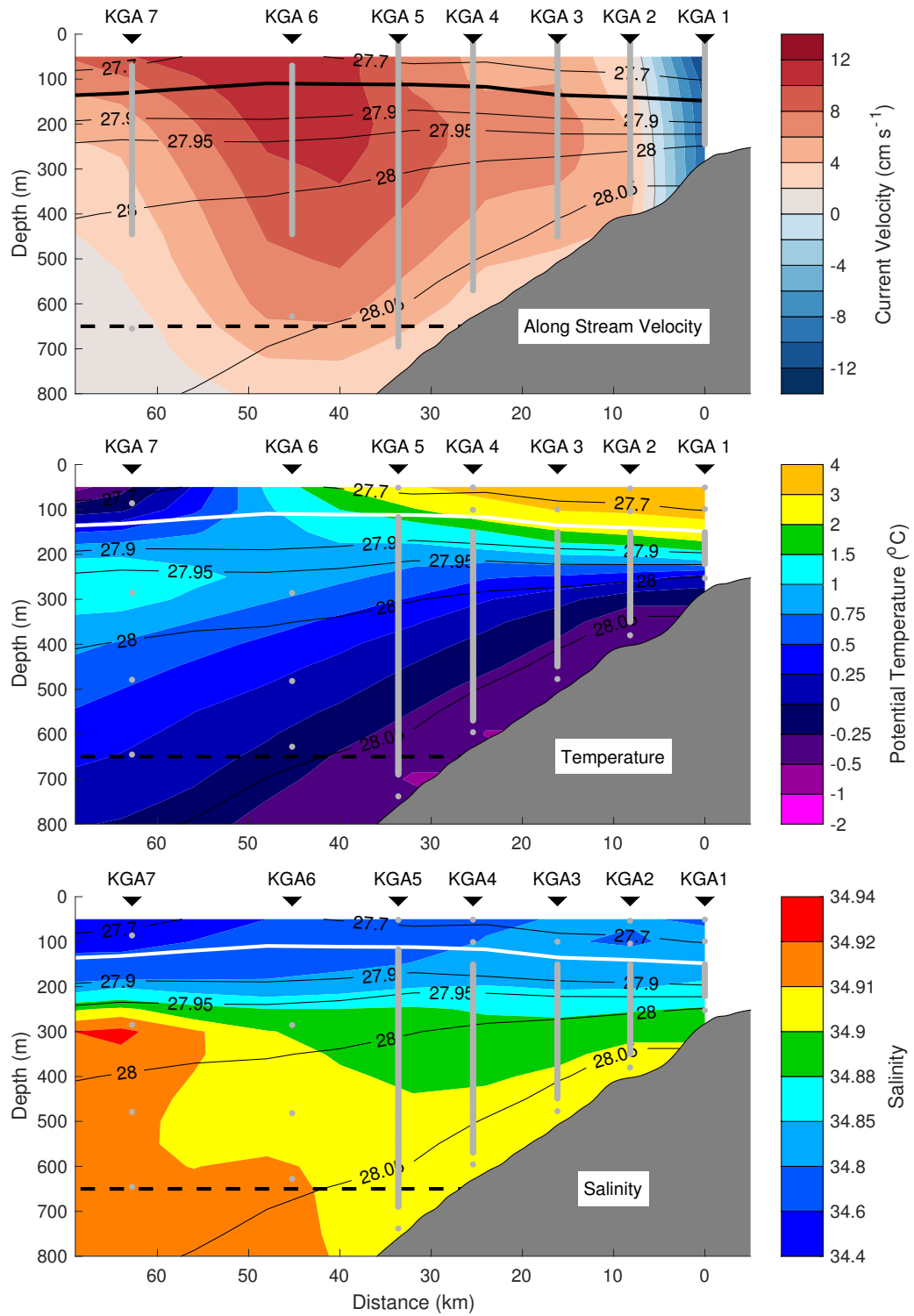


Figure 2: Mean vertical section of the along-stream (cross-transect) velocity (top), and median sections of potential temperature (middle) and salinity (bottom) for the 11-month period of the Kögur array. Overlaid in black contours on each panel is the mean density with the  $27.8 \text{ kg m}^{-3}$  isopycnal (the upper boundary of Denmark Strait Overflow Water) highlighted. The viewer is looking to the northeast with Iceland on the right. Positive velocities are equatorward. The horizontal black dashed line indicates the depth of the Denmark Strait sill. The moorings (black triangles) are labeled, and the average instrument locations are shown by the grey points. The bathymetry is from a shipboard echosounder.



130 seamounts and other sharp topographic features we smoothed the bathymetry using a filter  
131 of 60 km (comparable to our measured TRW wavelength). In contrast to Pickart (1995)  
132 who subsequently fit splines to the data to be able to find the bottom depth and gradients  
133 at any location, we deemed our resolution to be high enough (and our smoothing window  
134 great enough) to simply use linear interpolation. The total integration period for the wave  
135 tracing was 48 hours.

### 136 **3. Results**

137 As discussed in Harden *et al.* (2016), the vertical sections of velocity and hydrography  
138 at the Kögur site show the signatures of both the NIJ and the Separated EGC. However, the  
139 two features are merged to some degree in the mean (Figure 2). The NIJ is on the upper  
140 Iceland slope and is characterized by a mid-depth intensified flow carrying the coldest,  
141 densest overflow water banked up on the slope. The Separated EGC is farther offshore;  
142 its key features are a surface intensification and the transport of warmer, saltier overflow  
143 water at approximately 300 m. Inshore of both these currents, on the Iceland shelf, is the  
144 poleward flowing NIIC (see also Figure 1).

145 The two overflow currents are merged in the mean largely due to the high degree  
146 of variability on weekly timescales. The depth-integrated, along-stream velocity exhibits  
147 constant pulsing through this portion of the strait (Figure 3a). The period of the pulsing  
148 in the vicinity of the NIJ is concentrated at sub-8-day periods with a maximum average  
149 energy at 3.6 days (Figure 4). Farther offshore, near the Separated EGC, we also see  
150 such short-period pulses in addition to more consistent longer-period variability (Figure  
151 3a). The lower frequency signals were described by Harden *et al.* (2016) and attributed in  
152 part to the time-varying upstream bifurcation of the EGC. Here we focus on the higher-  
153 frequency, sub-8-day variability. To facilitate this we used an 8-day butterworth filter.<sup>1</sup>

154 The current ellipses of this high-frequency variability for each mooring are useful for

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<sup>1</sup>Different period filters were implemented, ranging in length from 4 days to 30 days, but the 8-day filter was most effective in isolating the peak high-frequency energy.

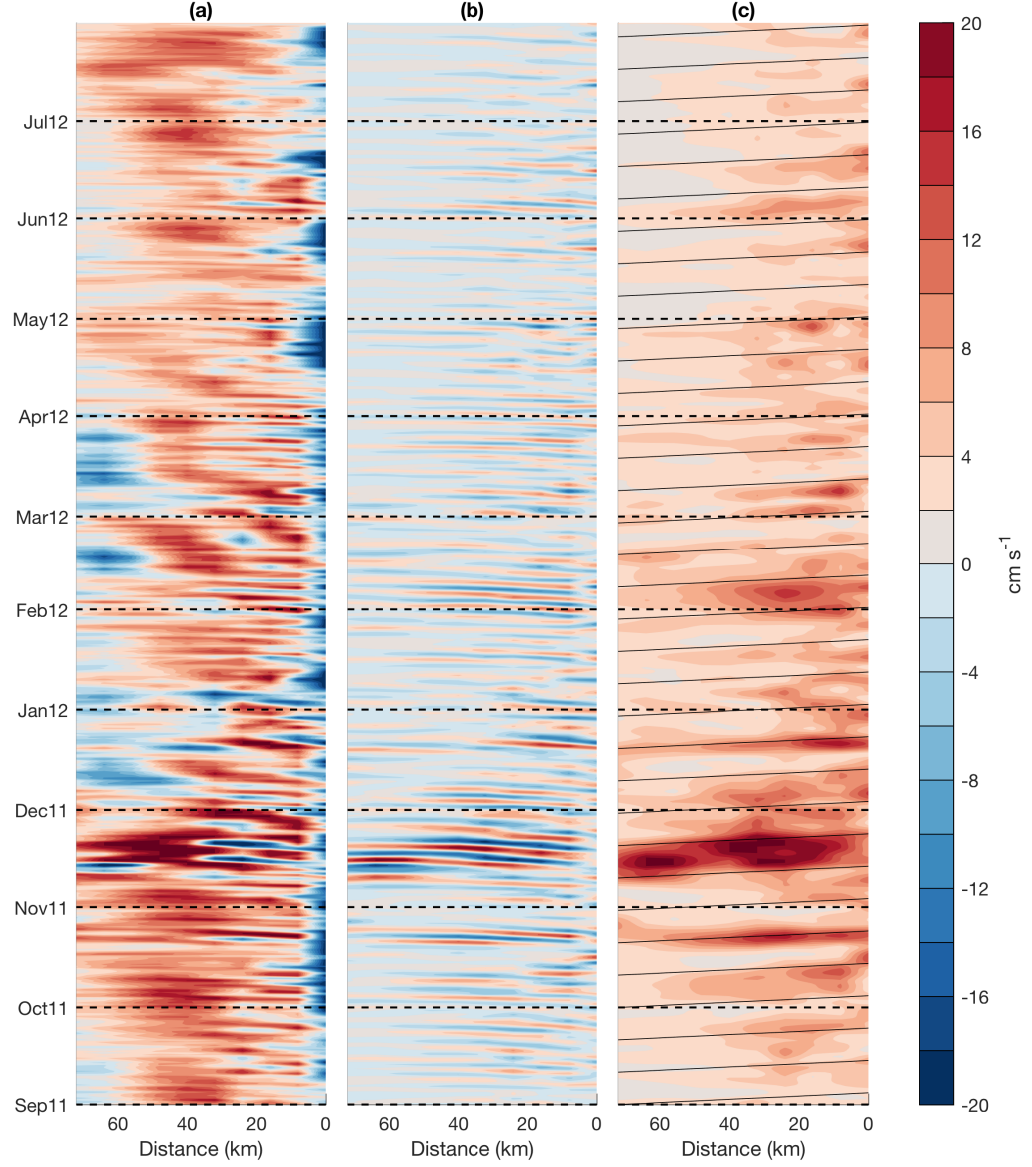


Figure 3: Hovmöller plots from the gridded mooring data of a) the depth-mean along-stream velocity (below 100 m, same for all plots); b) the 8-day high-passed, depth-mean component of velocity in the direction of the major axis of the local current ellipse; and c) the wavelet amplitude at a 4-day period for the depth-mean velocity. Iceland is to the right of each panel as in Figure 2. The wavelet analysis uses the jLab toolbox (Lilly, 2017) with standard Morlet wavelets with  $\gamma=3$  and  $\beta = 2$ . The sloped, black guidelines in panel c are angled at the theoretical group velocity for the measured topographic Rossby waves (see text for details).

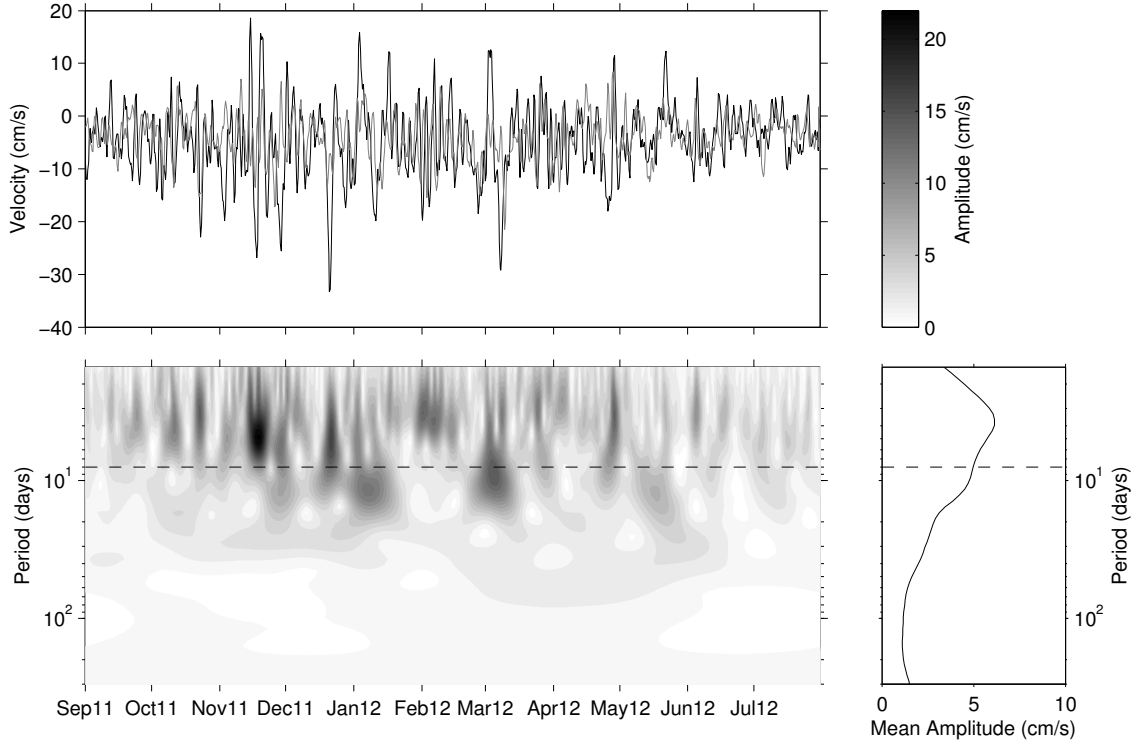


Figure 4: Top: Depth-averaged along-stream (black) and cross-stream (grey) components of velocity for the grid point closest to mooring KGA 3. Bottom left: Wavelet spectrum of the depth-averaged velocity using Morlet wavelets (Lilly, 2017). The color scale for this plot is at the top right. Bottom right: Mean wavelet amplitude for the length of the deployment. The dashed line in the bottom panels indicates the 8-day cut-off period for the high-pass filter used in this study.

155 characterizing different regimes across the array (Figure 5). In the NIIC (KGA 1), the  
 156 current ellipse is elongated in the direction of the mean flow indicative of a current pulsing  
 157 along its axis. By contrast, within the Separated EGC (KGA 6 and 7), the elongation of the  
 158 current ellipses is perpendicular to the mean flow demonstrating that this current meanders.  
 159 However, in the NIJ (KGA 2-4), the major axes of the current ellipses are aligned at an  
 160 oblique angle to both the mean flow and the underlying bathymetry. KGA 5 appears to be  
 161 in a transition region between conditions in the NIJ and those in the Separated EGC.

#### 162 *a. Topographic Rossby Waves*

163 We resolved the sub-8-day depth-averaged flow in the gridded product along the major  
 164 axis of the current ellipses at each offshore location. Particularly in the NIJ, the variability

165 along these axes have a sinusoidal form and are lagged between moorings such that the  
 166 pulses of current progress offshore in time (Figure 3b). This implies a downslope phase  
 167 propagation of this variability.

168 We argue that this is the signature of TRWs. These waves are supported by topo-  
 169 graphic  $\beta$  and result in transverse fluctuations that are often at an oblique angle to the  
 170 mean flow. TRWs are found in many slope regions of the worlds oceans (Garrett, 1979;  
 171 Louis *et al.*, 1982; Pickart and Watts, 1990). Key features of TRWs include wave vec-  
 172 tors (and hence phase velocities) that are perpendicular to the velocity variability, a group  
 173 velocity which is at an oblique angle to the phase velocity, and a tendency to be bottom-  
 174 trapped in regions of significant stratification.

175 Given that the phase propagation is perpendicular to the velocity variability, we de-  
 176 duce that the wave phase is progressing downslope at  $-9^\circ\text{T}$  (average from moorings KGA  
 177 2–4, see Figure 5). Following Pickart and Watts (1990), we then calculated the phase  
 178 speed over the range of moorings KGA 2–4 (where the wave signal is most pronounced)  
 179 using,

$$c_p = \frac{1}{T} \frac{360}{\bar{\phi}} \frac{\overline{\Delta S}}{\cos(\Delta)}$$

180 where  $T$  is the wave period ( $= 3.6$  days),  $\bar{\phi}$  is the average phase offset ( $= 48 \pm 3^\circ$ ),  
 181  $\overline{\Delta S}$  is the average instrument spacing ( $= 8.1 \pm 0.2$  km), and  $\Delta$  is the angle between the  
 182 mooring array and the direction of wave propagation ( $= 8 \pm 4^\circ$ ). The resulting phase  
 183 speed is  $17.3 \pm 0.8$  km day $^{-1}$  corresponding to a wavelength of  $62 \pm 3$  km. The error  
 184 estimates arise in equal contributions from uncertainties in  $\bar{\phi}$ ,  $\overline{\Delta S}$ , and  $\Delta$ .

185 As a consistency check that the observed fluctuations are in fact TRWs, we can em-  
 186 ploy the TRW dispersion relation for a uniformly stratified ocean neglecting planetary  $\beta$ .  
 187 Following Pedlosky (1979), this can be written as:

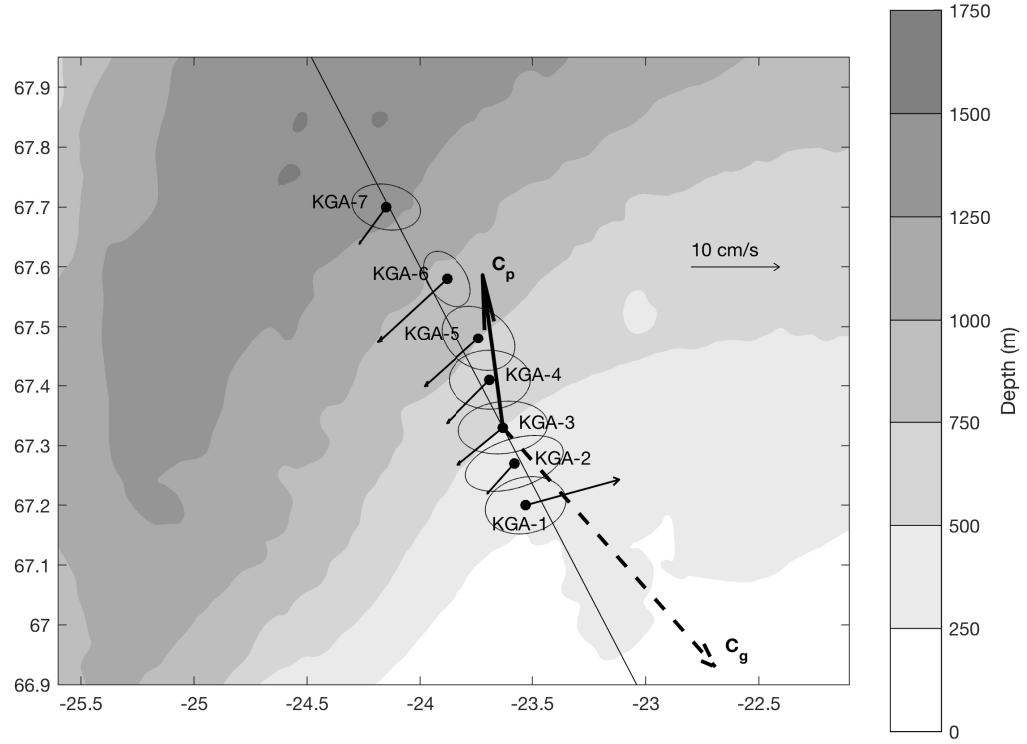


Figure 5: Aspects of the flow measured by the Kögur moorings (black circles). The thin vectors indicate the mean velocity averaged from 100 m to the depth of the ADCP at each mooring (see gray lines in Figure 2). Also shown are the 8-day high-passed current ellipses for the same depth range. The thick black arrow ( $C_p$ ) denotes the direction of TRW phase propagation averaged over KGA 2-4 (plotted at KGA 3). The dashed black arrow shows the direction of TRW group velocity ( $C_g$ ). All vectors and current ellipses are drawn to the same scale as indicated. The long black line is the mean downslope direction averaged between KGA 2-4. The bathymetry is from IBCAO v3.

$$T = \frac{2\pi \tanh(\frac{2\pi ND}{\lambda f})}{N\Gamma \sin(\theta)}$$

188 where  $T$  is the period of the wave,  $N$  is the average water column Brunt Väisälä  
 189 frequency ( $= 3.3 \times 10^{-5}$ , averaged using the gridded data below 100 m),  $D$  is the depth

190 ( $= 500$  m),  $\lambda$  is the wavelength,  $f$  is the Coriolis parameter ( $= 1.35 \times 10^{-4}$ ),  $\Gamma$  is the  
191 bottom slope ( $= 0.016$ , from IBCAO v3), and  $\theta$  is the phase velocity direction relative to  
192 downslope.

193 We can test the predicted value of  $\theta$  against the observed value using our knowledge  
194 of the other variables. The predicted angle of  $29^\circ$  compares well with the measured value  
195 of  $24^\circ$  (from the average downslope angle between moorings KGA 2–4). There is of  
196 course uncertainty in the measured downslope angle depending on the region selected  
197 for the averaging. For example, if we expand the calculation of the downslope direction  
198 to encompass KGA 1–5, the measured  $\theta$  becomes  $33^\circ$ , which still agrees well with the  
199 predicted value. In addition, the bottom-trapping scale ( $= f/Nk$ ) is much greater than  
200 1000 m, in agreement with the observed velocities which are largely barotropic.

201 All of this supports our assertion that the dominant high-frequency variability in the  
202 NIJ is due to TRWs. The obvious question is, where and how are these waves being  
203 generated? Using the dispersion relation we can calculate the group velocity. For the  
204 observed parameters, we find this to be  $36 \text{ km day}^{-1}$  directed almost directly up-slope at  
205 the array site ( $138^\circ\text{T}$ , see Figure 5). This implies that the energy source lies offshore.  
206 We can corroborate this onshore propagation of energy observationally by considering the  
207 wavelet amplitude for the 4-day signal at each mooring site. The Hovmöller plot of this  
208 shows clear occurrences of onshore energy propagation that are in line with the predicted  
209 group velocity (Figure 3c).

## 210 *b. Wave Tracing and TRW Formation Mechanisms*

211 In order to shed light on the source of the TRWs, we implemented the inverse wave  
212 tracing model described in Section 2. In particular, we calculated the wave paths back-  
213 wards in time from moorings KGA 2–5. For each mooring, the model was initialized  
214 with the local wavenumber (assuming constant phase velocity and wave period). Since  
215 KGA 5 only marginally displayed TRW behavior, the results from that mooring should  
216 be considered less robust. The calculated paths indicate that the waves originate offshore

217 of the moorings in the vicinity of the deep Blosseville Basin (Figure 6). While the traces  
218 diverge somewhat going offshore, it is clear that they do not deflect significantly upstream  
219 or downstream. In other words, the energy is not propagating along the Iceland continental  
220 slope.

221 TRWs are a ubiquitous feature in the middle Atlantic Bight between Cape Hatteras,  
222 NC and the Grand Banks (Louis *et al.*, 1982; Johns and Watts, 1986; Pickart and Watts,  
223 1990). The source of the waves appears to be the Gulf Stream. Both Hogg (1981) and  
224 Schultz (1987) argued that TRWs observed along the US continental slope emanated from  
225 large amplitude Gulf Stream meanders offshore. Louis *et al.* (1982) made the case that  
226 bursts of TRWs measured south of Nova Scotia resulted from Gulf Stream eddy formation.  
227 Pickart (1995) demonstrated that the TRWs observed near Cape Hatteras were forced by  
228 meanders of the Gulf Stream as it flowed over a bend in topography farther to the east.

229 In light of these studies, it is natural to suspect that the TRWs measured at the Kögur  
230 array site are generated by the Separated EGC. This current is energetic, and, as noted  
231 above, is subject to significant meandering (akin to the Gulf Stream). The wave tracing  
232 indicates that the energy emanates from the Blosseville Basin where the Separated EGC  
233 resides. Additionally, there is evidence that times of strong TRW activity on the upper  
234 slope are often preceded slightly by increases in meander energy offshore (Figure 3). The  
235 high-energy event in November is one example of this, but there are additional instances  
236 in late October, late December, and early March.

237 Another possible trigger for the waves is the intermittent aspiration of deeper waters  
238 towards the Denmark Strait Sill. Harden *et al.* (2016) demonstrated that 0.6 Sv of the  
239 overflow transport approaching the sill does so from below sill depth. Pulsing of this  
240 aspirated component of the flow across the isobaths could initiate topographic wave activ-  
241 ity. Regardless of the mechanism, the presence of TRWs raises the question of whether  
242 they are present along the entire Iceland slope or whether they are unique to our sampling  
243 region. To address this we examined the velocity data from a mooring deployed approxi-

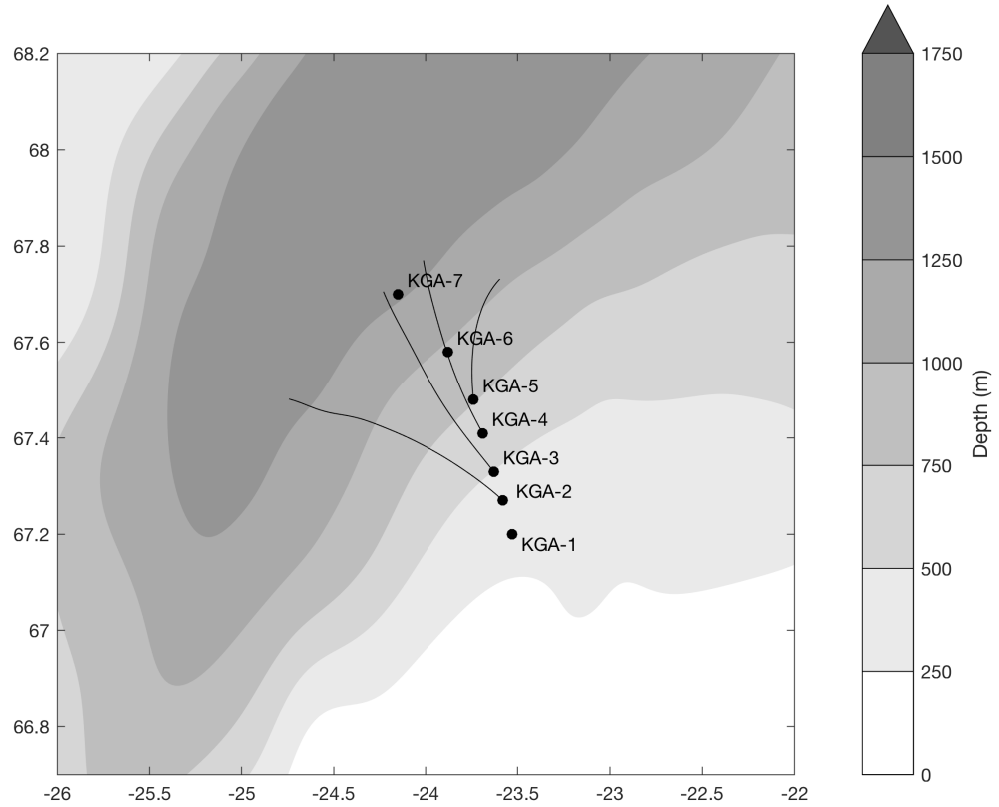


Figure 6: Paths of the Topographic Rossby Waves (thin lines) computed using the inverse wave tracing model for moorings KGA 2-5. Wave traces are truncated as they pass the 1500 m isobath. The bathymetry is from IBCAO v3 smoothed over 60 km (see text for details).



244 mately 200 km upstream on the Iceland slope near the Kolbeinsey ridge from 2007–2008  
245 (Jónsson and Valdimarsson, 2012). The depth-mean velocity showed very little energy  
246 in the 4-day period, at odds with the large TRW signal found at this period at the Kögur  
247 array. Notably, the upstream mooring site is quite far from the Separated EGC (Figure 1)  
248 and hence lacks that as an energy source.

## 249 **4. Summary and Discussion**

250 We have documented the existence of energetic topographic Rossby Waves (TRWs)  
251 within the North Icelandic Jet (NIJ) using observations from the densely-instrumented  
252 Kögur Array located approximately 200 km upstream of the Denmark Strait Sill. The  
253 mean period of the waves is 3.6 days, the wavelength is  $62 \pm 3$  km, and the phase velocity  
254 is  $17.3 \pm 0.8$  km day<sup>-1</sup> directed downslope ( $-9^\circ\text{T}$ ). Using the TRW dispersion relation,  
255 we corroborated our observed direction of phase propagation relative to the downslope  
256 direction ( $24^\circ$ ) with the theoretical value ( $29^\circ$ ). We further calculated that the wave energy  
257 is progressing up-slope ( $138^\circ\text{T}$ ) at  $36$  km day<sup>-1</sup>, in agreement with our observational  
258 data. It is likely that the energy in the TRWs emanates locally near the mooring site,  
259 either through the meandering of the offshore Separated East Greenland Current (EGC),  
260 or through pulses of cross-bathymetric flow due to the aspiration of deep overflow water  
261 as it approaches Denmark Strait.

262 Notably, our data imply that the dominant high-frequency variability at the Kögur site  
263 does not originate from the Denmark Strait, nor does it propagate towards the sill. It sug-  
264 gests that the mesoscale features at the sill (boluses and pulses), diagnosed observationally  
265 by von Appen *et al.* (2017) and in a model framework by Almansi *et al.* (2017), are not  
266 triggered by, nor excite, the TRWs on the Iceland Slope. However, the likelihood of a  
267 connection between the high frequency variability at the two locations is still high given  
268 the geographic proximity and the similarity in timescales, but is presumably mediated by  
269 another process. The Denmark Strait overflow is believed to be subject to hydraulic con-

270 trol (Whitehead, 1998; Nikolopoulos *et al.*, 2003), and, consequently, information should  
271 be transferred between the sill and the region to the north, likely as Kelvin waves. The ex-  
272 istence of any such connection and the impact on both the sill and NIJ variability requires  
273 further investigation and is the subject of an on-going study.

274 Finally, one also needs to consider where the energy in the TRWs ends up and what  
275 impact it might have on the dynamics of the circulation inshore of the Iceland slope. The  
276 energy likely cascades into the North Icelandic Irminger Current (NIIC) where it dissi-  
277 pates, leading to enhanced mixing. It might also alter the stability of NIIC, which brings  
278 warm subtropical water into the Nordic Seas. Våge *et al.* (2011) hypothesize that the off-  
279 shore flux of this warm water associated with the disintegration of the NIIC is tied to the  
280 overturning loop that forms the NIJ. Notably, eddies of NIIC water are found both in the  
281 Blossville Basin (Jónsson and Valdimarsson, 2012) and farther north in the Iceland Sea  
282 (Våge *et al.*, 2011). It is intriguing to think that the TRWs described here could play a role  
283 in this aspect of the NIIC.

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