

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 on the Iceland Slope
8 using data from the densely instrumented Kögur mooring array deployed upstream of the
9 Denmark Strait sill from September 2011 to July 2012. Significant sub-8-day variability
10 is ubiquitous in all moorings from the Iceland slope with a dominant period of 3.6 days.
11 We attribute this variability to Topographic Rossby Waves on the Iceland slope with a
12 wavelength of 62 ± 3 km and a phase velocity of 17.3 ± 0.8 km day⁻¹ directed downslope
13 (-9°T). We test the theoretic dispersion relation for these waves against our observations
14 and find good agreement between the direction of phase propagation. We additionally
15 calculate a theoretical group velocity of 36 km day⁻¹ directed almost directly up-slope
16 (138°T) which agrees well with the propagation speed of observed energy pulses headed
17 upslope. We use a wave tracing model to show that this wave energy is generated locally,
18 offshore of the array, and not in the upstream or downstream directions. We hypothesize
19 that either the meandering Separated East Greenland Current at the foot of the Iceland
20 slope or intermittent aspiration into the Denmark Strait Overflow are the drivers of the
21 Topographic Rossby Waves. Regardless of the formation mechanism, the waves appear to
22 be a local phenomena, not found in an instrumented record upstream.

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. As
41 the current rounds Scoresby Sund, it splits into two branches (Figure 1). One continues to-
42 wards the sill as a shelfbreak jet (Håvik *et al.*, 2017). The other carries approximately 60%
43 of the East Greenland Current water out into the central strait via eddies and/or gyre-like
44 deflections of the shelfbreak jet (Våge *et al.*, 2013; Harden *et al.*, 2016). This separated
45 pathway then flows into the strait along the outer Iceland slope.

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

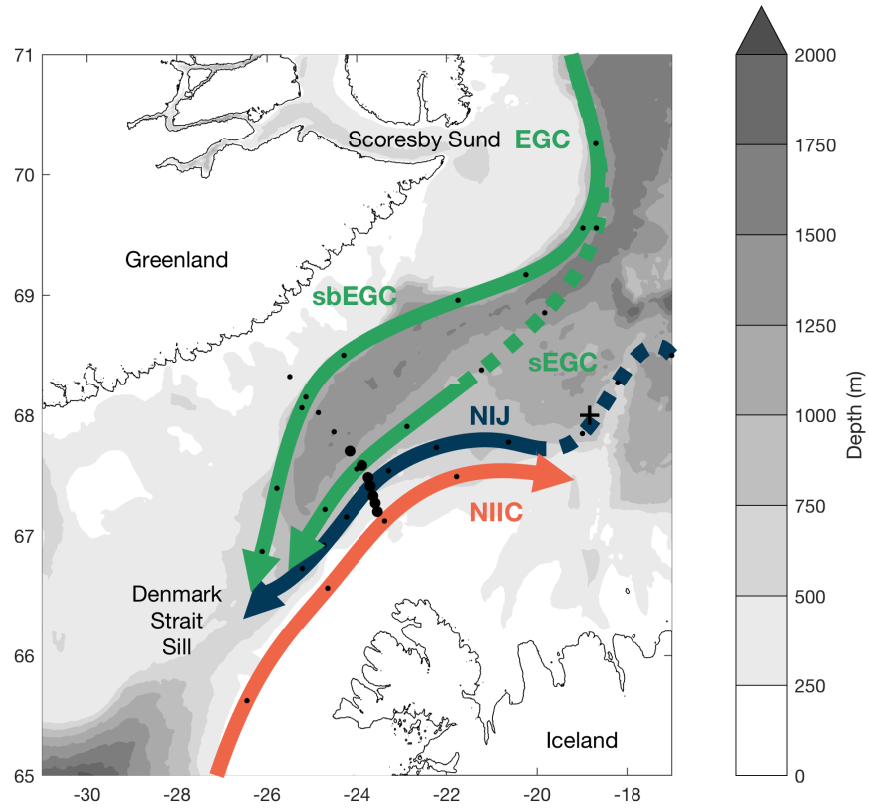


Figure 1: Map of the study region showing the overflow pathways approaching the Denmark Strait Sill: the North Icelandic Jet (NIJ) and the two East Greenland Current (EGC) pathways, one along the shelfbreak (sbEGC) and the other in a separated branch on the Iceland Slope (sEGC). Dashed portions show parts of pathways that still need further clarification. Also shown is the northward flowing surface-intensified current, the North Icelandic Irmingier Current (NIIC). Block 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 a source in the central Iceland or Greenland seas (Våge *et al.*, 2011; 2015; Harden *et al.*,
51 2016). The NIJ contains the densest water that feeds the overflow; its waters are found in
52 the deepest part of the sill (Mastropole *et al.*, 2017) and subsequently sink to the deepest
53 depths in the core of the overflow.

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

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

73 It has long been known that the Denmark Strait Overflow varies on short (order days)
74 timescales (Smith, 1976; Bruce, 1995; Käse *et al.*, 2003). Some of this variability is as-
75 sociated with the passage of lenses of cold, dense, overflow water referred to as boluses
76 (Cooper, 1955). Recently, Appen *et al.* (2017) identified a second type of mesoscale fea-

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

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

94 **2. Data and Methods**

95 The data for this study come from the densely instrumented Kögur mooring array
96 spanning the Denmark Strait approximately 200 km upstream of the sill. The array was
97 deployed for 11 months from September 2011 to July 2012 and consisted of 12 moorings
98 (named KGA 1-12) equipped with instrumentation to measure both the hydrography and
99 velocity of the water column from 50 m to the bottom. Harden *et al.* (2016) present a
100 detailed description of the mooring data, including the instrumentation, processing steps,
101 and sensor accuracies. The array captured the majority of overflow water (denser than
102 27.8 kg m^{-3}) passing through the northern part of the strait towards the sill.

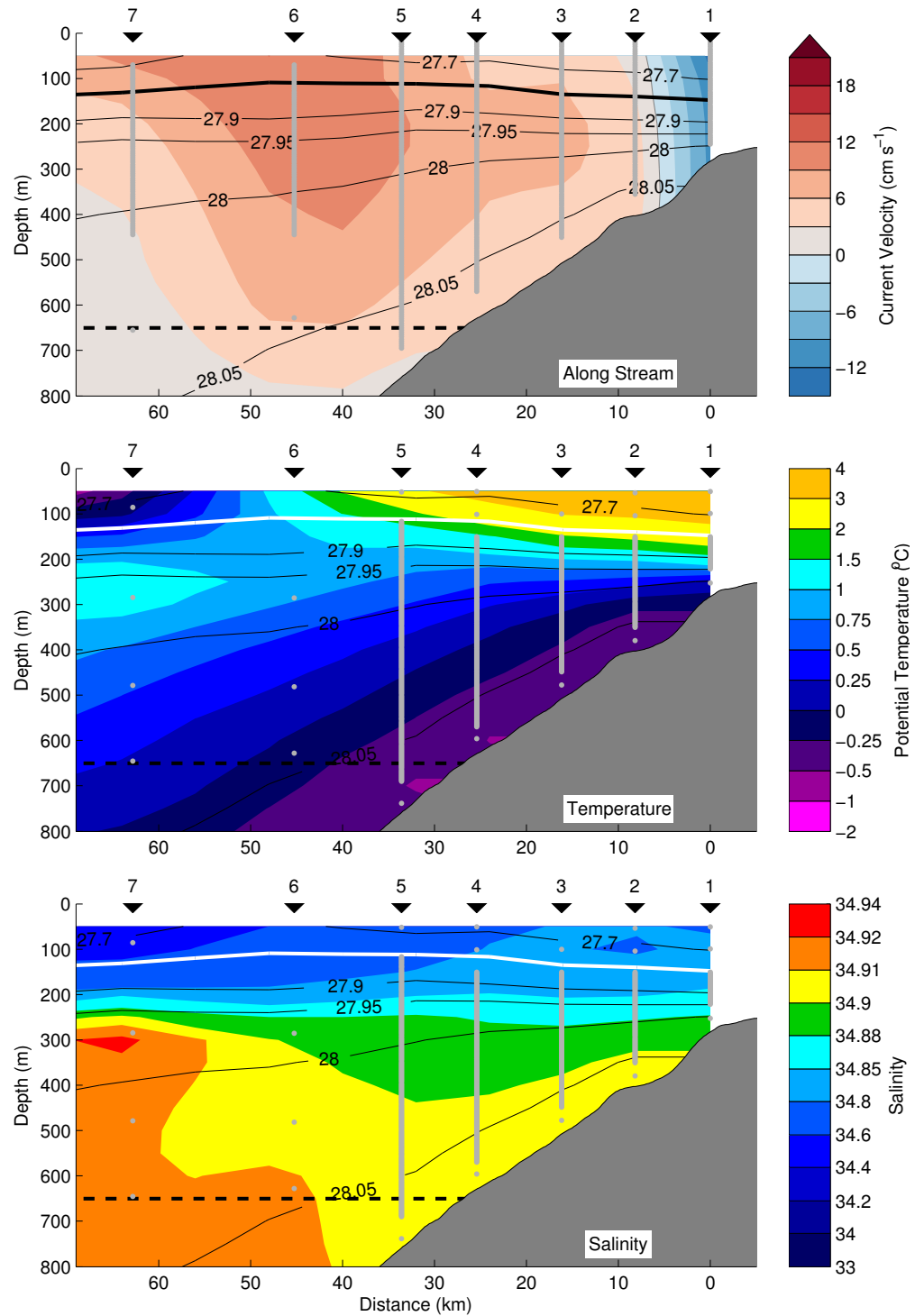


Figure 2: Mean vertical section of the along-stream (cross-transect) velocity (top), and median sections of potential temperature (middle) and salinity (bottom) for the 11-month period of the Kögur array. Overlaid in black contours on each panel is the mean density with the 27.8 kg m^{-3} isopycnal (the 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.

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

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

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

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

135 **3. Results**

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

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

153 The variance ellipses of this high-frequency variability for each mooring are useful
154 for characterizing different regimes across the array (Figure 5). In the NIIC (KGA 1), the

¹Different 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.

./figures/hoff_vars.pdf

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 major axis of the local variance ellipse; and c) the wavelet amplitude at a 4-day period for the depth-mean velocity. The wavelet analysis uses the jLab toolbox (Lilly, 2017) with standard Morlet wavelets with $\gamma=3$ and $\beta = 2$. The sloped, black guidelines are angled at the theoretical group velocity for the measured topographic Rossby waves (see text for details).

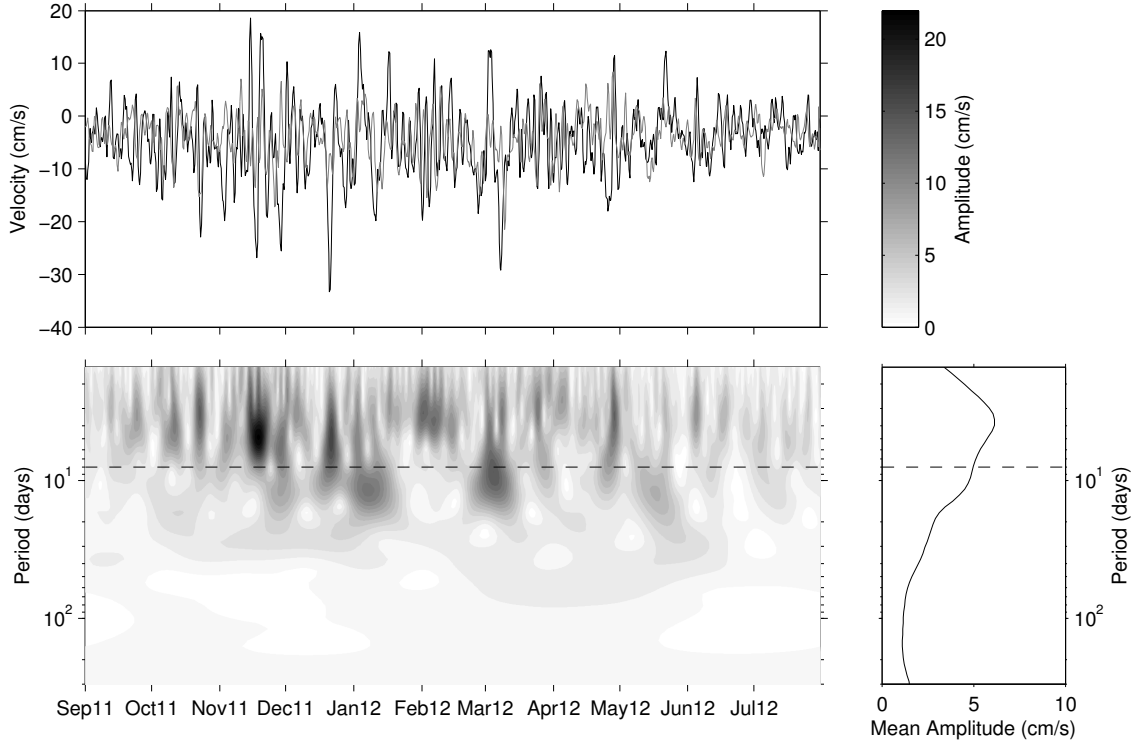


Figure 4: Top: Depth-averaged along-stream (black) and cross-stream (grey) components of velocity for the grid point closest to mooring KGA 3. Bottom left: Wavelet spectrum of the depth-averaged speed using Morlet wavelets. The color scale for this plot is at the top right. Bottom right: Mean wavelet 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.

155 variance ellipse is elongated in the direction of the mean flow indicative of a current puls-
 156 ing along its axis. By contrast, within the Separated EGC (KGA 6 and 7), the elongation of
 157 the variance ellipses is perpendicular to the mean flow demonstrating that this current me-
 158 anders. However, in the NIJ (KGA 2-4), the major axes of the variance ellipses are aligned
 159 at an oblique angle to both the mean flow and the underlying bathymetry. KGA 5 appears
 160 to be in a transition region between conditions in the NIJ and those in the Separated EGC.

161 *a. Topographic Rossby Waves*

162 We resolved the sub-8-day depth-averaged flow in the gridded product along the major
 163 axis of the variance ellipses at each offshore location. Particularly in the NIJ, the variability
 164 along these axes have a sinusoidal form and are lagged between moorings such that the

165 pulses of current progress offshore in time (Figure 3b). This implies a downslope phase
 166 propagation of this variability.

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

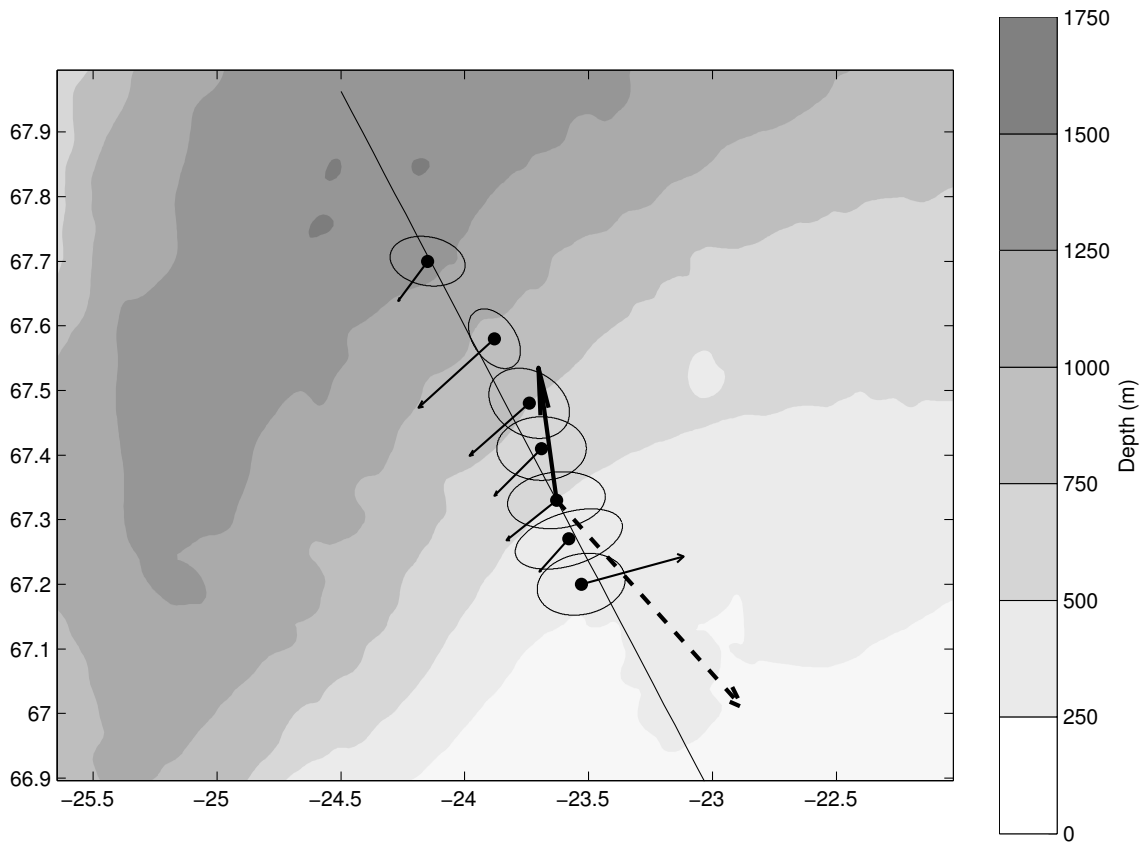


Figure 5: Aspects of the flow measured by the Kögur moorings (black circles). The thin vectors indicate the mean velocity averaged from 100 m to the depth of the ADCP at each mooring (see gray lines in Figure 2). Also shown are the 8-day high-passed variance ellipses for the same depth range. The thick black arrow 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. The long black line is the mean downslope direction averaged between KGA 2-4. Bathymetry is from IBCAO.

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

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

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

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

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

187 where T is the period of the wave, N is the average water column Brunt Väisälä
 188 frequency ($= 3.3 \times 10^{-5}$, averaged using the gridded data below 100 m), D is the depth
 189 ($= 500$ m), λ is the wavelength, f is the Coriolis parameter ($= 1.35 \times 10^{-4}$), Γ is the
 190 bottom slope ($= 0.016$, from IBCAO v3), and θ is the phase velocity direction relative to
 191 downslope.

192 We can test the predicted value of θ against the observed value using our knowledge

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

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

209 *b. Wave Tracing and TRW Formation Mechanisms*

210 In order to shed light on the source of the TRWs, we implemented the inverse wave
211 tracing model described in Section 2. In particular, we calculated the wave paths back-
212 wards in time from moorings KGA 2–5. For each mooring, the model was initialized
213 with the local wavenumber (assuming constant phase velocity and wave period). Since
214 KGA 5 only marginally displayed TRW behavior, the results from that mooring should
215 be considered less robust. The calculated paths indicate that the waves originate offshore
216 of the moorings in the vicinity of the deep Blosseville Basin (Figure 6). While the traces
217 diverge somewhat going offshore, it is clear that they do not deflect significantly upstream
218 or downstream. In other words, the energy is not propagating along the Iceland continental
219 slope.

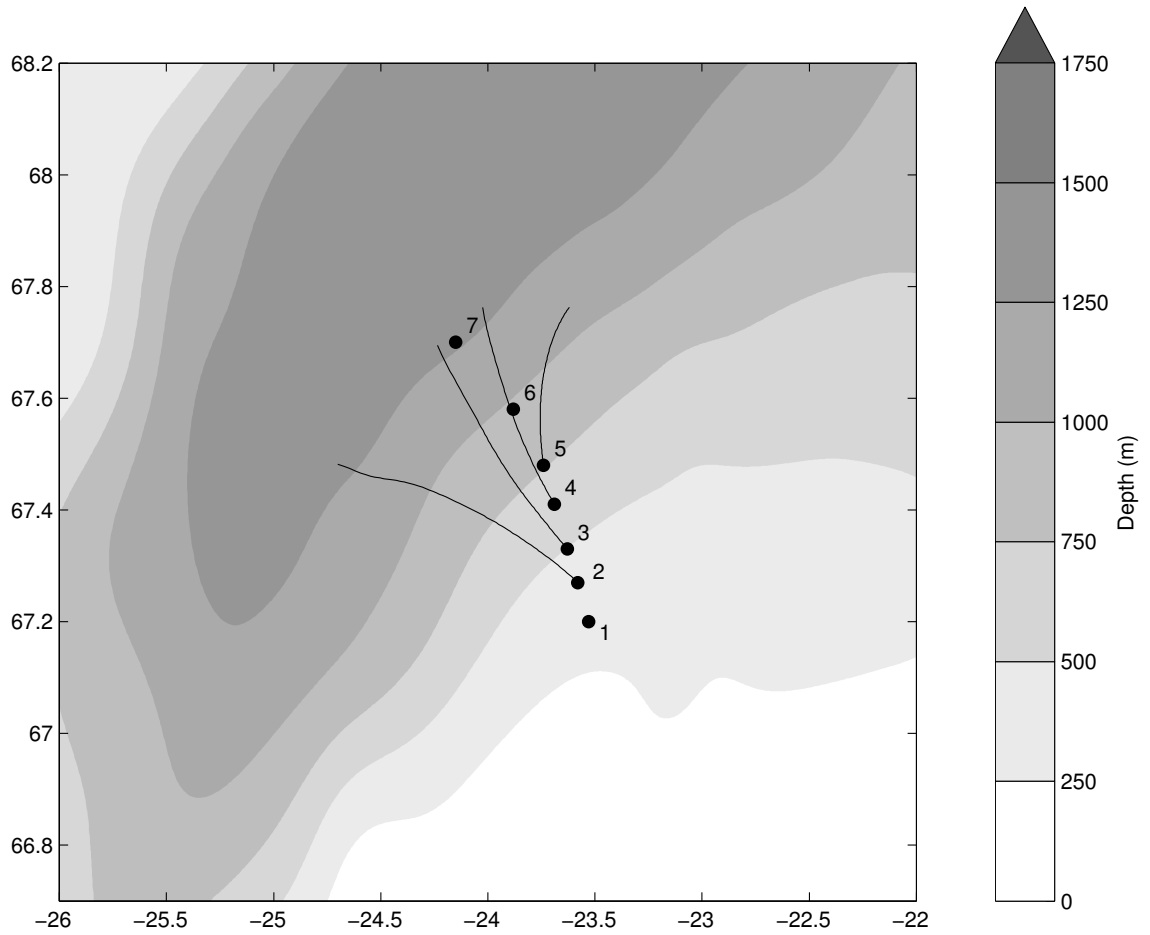


Figure 6: Paths of the topographic Rossby waves (thin lines) computed using the inverse wave tracing model for moorings KGA 2-5. The bathymetry is from IBCAO v3 smoothed over 60 km (see text for details).

220 TRWs are a ubiquitous feature in the middle Atlantic Bight between Cape Hatteras,
221 NC and the Grand Banks (Louis *et al.*, 1982; Johns and Watts, 1986-05-01T00:00:00;
222 Pickart and Watts, 1990). The source of the waves appears to be the Gulf Stream. Both
223 Hogg (1981) and Schultz (1987) argued that TRWs observed along the US continental
224 slope emanated from large amplitude Gulf Stream meanders offshore. Louis *et al.* (1982)
225 made the case that bursts of TRWs measured south of Nova Scotia resulted from Gulf
226 Stream eddy formation. Pickart (1995) demonstrated that the TRWs observed near Cape
227 Hatteras were forced by meanders of the Gulf Stream as it flowed over a bend in topogra-
228 phy 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-
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

array. Notably, the upstream mooring site is quite far from the Separated EGC (Figure 1) and hence lacks that as an energy source.

4. Summary and Discussion

We have documented the existence of energetic Topographic Rossby Waves (TRWs) within the North Icelandic Jet (NIJ) using observations from the densely-instrumented Kögur Array located approximately 200 km upstream of the Denmark Strait Sill. The mean period of the waves is 3.6 days, the wavelength is 64 ± 3 km, and the phase velocity is 17.7 ± 0.8 km day⁻¹ directed downslope (-9°T). Using the TRW dispersion relation, we corroborated our observed direction of phase propagation relative to the downslope direction (25°) with the theoretical value (26°). We further calculated that the wave energy is progressing up-slope (137°T) at 39 km day⁻¹, in agreement with our observational data. It is likely that the energy in the TRWs emanates locally near the mooring site, either through the meandering of the offshore Separated East Greenland Current (EGC), or through pulses of cross-bathymetric flow due to the aspiration of deep overflow water as it approaches Denmark Strait.

Notably, our data imply that the dominant high-frequency variability at the Kögur site does not originate from the Denmark Strait, nor does it propagate towards the sill. This is perhaps surprising in light of the fact that fluctuations of the dense overflow water at the sill occur on similar short timescales (Jochumsen *et al.*, 2017). It suggests that the mesoscale features (boluses and pulses) diagnosed observationally by Appen *et al.* (2017) and in a model framework by Almansi *et al.* (2017) are generated locally in the vicinity of the sill. On the other hand, the Denmark Strait overflow is believed to be subject to hydraulic control (Whitehead, 1998; Nikolopoulos *et al.*, 2003). Consequently, when there are variations at the sill the information should be transferred upstream towards the Iceland Sea, likely as Kelvin waves. Our results indicate that any such signals are dominated by the locally-generated TRW energy. However, this deserves more careful investigation,

273 and the question of how signals are transferred between the sill and the Iceland slope is
274 the subject of an on-going study.

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

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