

1 **Fluctuations of the Atlantic North Equatorial**
2 **Undercurrent and associated changes in oxygen**
3 **transports**

4 **K. Burmeister¹, J. F. Lübbecke^{2,3}, P. Brandt^{2,3}, M. Claus^{2,3}, and J. Hahn²**

5 ¹SAMS, Scottish Marine Institute, Oban, Argyll, PA37 1QA, United Kingdom

6 ²GEOMAR Helmholtz Centre for Ocean Research Kiel, Düsternbrooker Weg 20, 24105 Kiel, Germany

7 ³Christian-Albrechts-Universität zu Kiel, Christian-Albrechts-Platz 4, 24118 Kiel, Germany

8 **Key Points:**

- 9 • Transport time series of the Atlantic North Equatorial Undercurrent estimated
10 from moored observations
- 11 • North Equatorial Undercurrent dominated by intraseasonal variability, only weak
12 seasonal cycle
- 13 • Occasional increase of eastward oxygen supply by North Equatorial Undercurrent
14 due to short sporadic events

Corresponding author: Kristin Burmeister, kburmeister@geomar.de

Abstract

15 **Abstract**
16 Although the core velocity of the Atlantic North Equatorial Undercurrent (NEUC) is low
17 ($0.1 - 0.3 \text{ m s}^{-1}$) it has been suggested to act as an important oxygen supply route towards
18 the oxygen minimum zone in the eastern tropical North Atlantic. For the first time the
19 intraseasonal to interannual NEUC variability and its impact on oxygen is investigated based
20 on shipboard and moored velocity observations around 5°N , 23°W . In contrast to previous
21 studies that were mainly based on models or hydrographic data, we find hardly any seasonal
22 cycle of NEUC transports in the central Atlantic. The NEUC transport variability is instead
23 dominated by sporadic intraseasonal events. Only some of these events are associated with
24 high oxygen levels suggesting an occasional eastward oxygen supply by NEUC transport
25 events. Nevertheless, they likely contribute to the local oxygen maximum in the mean
26 shipboard section along 23°W at the NEUC core position.

Plain Language Summary

27 **Plain Language Summary**
28 In the eastern tropical North Atlantic a zone of low-oxygen waters exists between 100 m
29 and 700 m depth due to high oxygen consumption and weak exchange of water masses. Long-
30 term oxygen changes in this zone have been reported with possible impacts on, for example,
31 the ecosystem or the available habitat for fish. Typically, water masses in that region
32 are exchanged via weak eastward and westward currents. As the oxygen concentration
33 in the western Atlantic basin is high, an eastward current such as the North Equatorial
34 Undercurrent (NEUC) may transport oxygen-rich waters into the eastern low-oxygen zone.
35 Given the east-west difference in oxygen concentration, we assume that a stronger NEUC
36 is transporting more oxygen-rich water from the western towards the eastern basin. This is
37 the first study that investigates the variations in NEUC transport based on direct velocity
38 measurements at 5°N , 23°W . In contrast to previous studies based on model simulations or
39 hydrographic data, we do not find a seasonal cycle of the NEUC transport. Instead, changes
40 of the NEUC transport are dominated by bursts of eastward flow which persist for a few
41 months. These eastward flow bursts are only occasionally associated with higher oxygen
42 concentrations.

43 1 Introduction

44 The circulation of the upper tropical Atlantic Ocean is characterized by a complex
45 current system which takes part in the wind-driven equatorial gyre circulation, the shallow
46 subtropical and tropical overturning cells and the basin-wide Atlantic meridional overturn-
47 ing circulation (e.g. Hazeleger & Drijfhout, 2006; Schott et al., 2004). Zonal currents play a
48 key role in the basin wide distribution of water mass properties and affect the transport of
49 heat, salt and biogeochemical components such as oxygen (e.g. Brandt et al., 2015; Hazeleger
50 & Drijfhout, 2006; Schott et al., 2004). The North Equatorial Undercurrent (NEUC) is an
51 eastward flowing subsurface (here defined in the depth range 65-270 m) current centered at
52 5°N . Its upper limit is commonly defined as the 24.5 kg m^{-3} neutral density layer, which
53 separates the tropical surface water from the subtropical underwater (e.g. Bourlès et al. ,
54 1999; Goes et al., 2013; Schott et al., 1995). Although the NEUC core velocity (0.1 m s^{-1}
55 to 0.3 m s^{-1}) is one of the lowest among the wind-driven off-equatorial currents in the trop-
56 ical Atlantic the NEUC has been suggested to act as an important oxygen supply route
57 towards the oxygen minimum zone in the eastern tropical North Atlantic (Brandt et al.,
58 2010; Stramma et al., 2008).

59 Model studies generally agree that the NEUC is mainly in geostrophic balance but
60 its driving mechanism is under discussion. Potential mechanisms that were described for
61 the NEUC or similar subsurface currents in other tropical basins are the conservation of
62 angular momentum of the tropical overturning cells (Marin et al., 2000), the Eliassen-
63 Palm flux associated with the propagation of Tropical Instability Waves (TIWs; Jochum
64 & Malanotte-Rizzoli, 2004) or the pull of upwelling within domes in the eastern basin or at
65 the eastern boundary (Furue et al., 2007, 2009; McCreary et al., 2002).

66 Until now, the seasonal to long-term variability of the NEUC has been investigated
67 based on model output (Burmeister et al., 2019; Hüttl-Kabus & Böning, 2008) or geostrophic
68 velocities derived from a combination of hydrography and satellite data (Goes et al., 2013).
69 In these studies, a seasonal cycle of the NEUC with amplitudes of 1 Sv to 3.5 Sv was iden-
70 tified. In general, NEUC transport estimates derived from meridional ship sections are
71 obscured by mesoscale activity (Weisberg & Weingartner, 1988) and interannual variability
72 (Hüttl-Kabus & Böning, 2008). So far, studies based on shipboard velocity observations
73 have not been able to detect a seasonal cycle of the NEUC (Bourlès et al., 2002, 1999;
74 Burmeister et al., 2019; Schott et al., 2003, 1995; Urbano et al., 2008).

75 For the first time, we will investigate the NEUC variability using direct velocity ob-
76 servations. In this study we reconstruct the eastward transport associated with the NEUC
77 at 5°N, 23°W using moored velocity observations from June 2006 to February 2008 and
78 November 2009 to January 2018 in combination with 24 meridional ship sections taken
79 between 21°W and 26°W. This study aims to investigate the intraseasonal to interannual
80 variability of the NEUC and its impact on oxygen levels.

81 **2 Data**

82 Moored Acoustic Doppler Current Profiler (ADCP) velocity, hydrography and oxygen
83 data are available at 5°N, 23°W (Jun 2006-Feb 2008, Nov 2009-Jan 2018) as well as at 4.6°N,
84 22.4°W and 4.5°N, 23.4°W (Nov 2012-Apr 2014). All instruments were set to a sampling
85 rate of 2 h or higher. The upper limit of the ADCP observations varies between 65 m and
86 85 m, the lower limit is at least 755 m. The moored velocity data were linearly interpolated
87 onto a regular time-depth grid (12 h \times 10 m), and a 40-day low-pass Butterworth filter was
88 applied to remove tides from the time series. As the NEUC is located approximately between
89 3.5°N and 6°N and in a depth of about 65 m to 270 m at 23°W (Burmeister et al., 2019)
90 the moored velocity observations do not always cover the entire extent of the NEUC which
91 can result in an underestimation of its mean transport. Observations of dissolved oxygen,
92 temperature, conductivity and pressure were used from respective sensors that were installed
93 at the mooring at 5°N, 23°W in 100 m, 200 m, and 300 m depth. The moored hydrography
94 and oxygen data were interpolated onto a 12-h time grid. This data set is an extension of
95 the one used in Hahn et al. (2014, see Text S1 for more details).

96 In addition to the mooring time series we use data from 24 meridional ship sections
97 taken between 21°W and 26°W in the time period 2002 to 2018 (Table S1). Only shipboard
98 ADCP, hydrography and oxygen sections that cover at least the upper 350 m between 0°
99 and 10°N are used. The ship sections are an extension of the data set used in Burmeister
100 et al. (2019, see Text S1 for more details).

101 **3 Observed velocity variability at 5°N, 23°W**

102 Moored ADCP measurements at 5°N, 23°W show a weak mean eastward velocity with
103 maximum values of 9 cm s⁻¹ at the upper limit of the ADCP range (85 m), while the merid-
104 ional velocity varies around zero (Fig. 1 a and b). In the upper 300 m zonal and meridional
105 velocities exhibit anomalies of comparable magnitude. The periodogram of the horizon-

106 tal velocity components indicates variability over a range of frequencies, in particular in
 107 the intraseasonal band (Fig. 1 c). While the zonal velocity exhibits variability mainly for
 108 periods greater than 70 days, the meridional velocity is dominated by variability with peri-
 109 ods between 35 and 75 days which is associated with TIWs (Jochum & Malanotte-Rizzoli,
 110 2003). The zonal velocity, while eastward in the mean, occasionally changes to westward.
 111 Its variability is characterized by strong eastward anomalies with a duration of about one
 112 to five months occurring without a clear seasonal preference. Unexpectedly, the seasonal
 113 cycle of the zonal velocity is much weaker than found in previous studies (Burmeister et
 114 al., 2019; Goes et al., 2013; Hüttl-Kabus & Böning, 2008). The annual harmonic fit with a
 115 maximum amplitude of 3 cm s^{-1} only explains between 2% (85 m) and 9% (300 m) of the
 116 zonal velocity variability. For the semi-annual harmonic we derived a maximum amplitude
 117 of 5 cm s^{-1} and an explained variance between 11% (75 m) and 1% (300 m).

123 **4 NEUC transport estimates at 23°W**

124 In this study, the NEUC transport was estimated by four different methods using only
 125 eastward velocities. (i) We calculated the NEUC transport between the 24.5 kg m^{-3} and
 126 26.8 kg m^{-3} neutral density surfaces from 24 ship sections using a path following algorithm
 127 developed by Hsin and Qiu (2012) (Fig. 2, green diamonds). This method follows the current
 128 core, thereby avoiding artifacts in the transport calculation if the current is meridionally
 129 migrating (see Text S2 for more details). Using this method we estimate a mean NEUC
 130 transport of $2.7 \pm 0.4 \text{ Sv}$. Uncertainties are given in terms of the standard error. In the
 131 following, we consider this estimate as a reference NEUC transport keeping in mind that
 132 the NEUC transport estimate from ship sections can be obscured by mesoscale activity
 133 (Weisberg & Weingartner, 1988) and interannual variability (Hüttl-Kabus & Böning, 2008).

134 (ii) The second approach is also based on ship sections only, but with a fixed smaller
 135 integration box to be consistent with NEUC transports calculated from a combination of
 136 ship mounted and moored observations. We integrate the meridional sections of zonal
 137 velocity between 4.25°N and 5.25°N , 65 m and 270 m (Fig. 2, black circles). The mean fixed
 138 box integrated transport derived from ship sections is $1.4 \pm 0.2 \text{ Sv}$. The fixed box integrated
 139 method underestimates the NEUC strength by 1.3 Sv. However, it represents the variability
 140 of the reference NEUC transport well ($R=0.91$, Fig. S2a).

141 (iii) We reconstruct the NEUC transport combining moored velocity observations at
 142 5°N, 23°W as well as 4.6°N, 22.4°W and 4.5°N, 23.4°W and ship sections following the
 143 optimal width (OW) method described in Brandt et al. (2014). This method aims to find
 144 an optimal latitude range W_i for each mooring position i to reconstruct the latitudinally
 145 integrated zonal velocity $U(z, t)$ by:

$$U(z, t) = \sum_{i=1}^3 W_i u_i(z, t) \quad (1)$$

146 For this purpose, we latitudinally integrate the eastward velocities of each ship section
 147 from 4.25°N to 5.25°N. The latitudinally integrated velocity is then regressed onto the
 148 eastward velocities of the ship sections at the three mooring position between 65m and
 149 270m depth to obtain W_i . We estimated that the moorings at 5°N, 4.6°N and 4.5°N
 150 correspond to latitude ranges, W_i , of 0.46°, 0.18° and 0.37°, respectively. The NEUC
 151 transport is then reconstructed by integrating Equation 1 over the same depth range as for
 152 method (ii) (Fig. 2, blue line). Note that data from all three mooring positions are only
 153 available for the time period November 2012 to April 2014. The root mean square error of
 154 the reconstructed transport using method (iii) and the box integrated transport (method ii)
 155 using the shipboard data is 0.16 Sv. The mean reconstructed transport is 0.9 ± 0.3 Sv. For
 156 a validation of the method see Text S2 in the supplementary information.

157 (iv) This method is similar to method (iii) but uses moored velocity observations only
 158 at 5°N, 23°W to obtain a longer transport time series (Fig. 2 grey line). Here, the 5°N
 159 mooring corresponds to a latitude range of 0.88°. Although the root mean square error
 160 between the reconstructed transport and the box integrated transport increases (0.51 Sv)
 161 when using only one mooring, the reconstructed transport based on one mooring agrees well
 162 with the one reconstructed using three moorings ($R=0.89$). The mean NEUC transport is
 163 the same as estimated for method (iii), i.e. 0.9 ± 0.2 Sv.

164 In the following we will analyze the NEUC variability based on the reconstructed time
 165 series using only the mooring at 5°N, 23°W. Although the reconstructed transport time
 166 series from moored velocity observations tends to underestimate the mean current strength
 167 of the NEUC, we still consider the variability to be captured to a large extent. This is
 168 supported by transport estimates based on all three moorings combined with zonal velocity
 169 sections between 3.5°N and 6.0°N which agree reasonably well with the transport time series

170 reconstructed by method (iii) accounting for velocities between 4.25°N and 5.25°N only (Fig.
171 S2-S5, Text S2).

180 **5 NEUC variability**

181 Similar to the zonal velocity, the NEUC transport is dominated by strong eastward
182 anomalies that persist between one and five months. The transport time series do not
183 exhibit a seasonal cycle. The explained variance of the annual and semi-annual harmonic is
184 only 2% and 3%, respectively. We therefore focus the analysis on the intraseasonal eastward
185 flow events. Rather arbitrary, we define a strong eastward flow event if the NEUC transport
186 exceed 2.5 times the standard deviation of the complete monthly mean time series. We
187 find eight such events which take about one to five months to develop, peak and fade away
188 (duration from local minimum to minimum in the transport time series). They have a
189 maximum monthly mean transport between 2.3 Sv and 3.8 Sv and peak without a clear
190 seasonal preference (3 in January, 1 each in April, May, June, August and December).
191 During these events the 12-hourly transports reach maximum values from 3.8 Sv to 7.2 Sv.

192 The NEUC at 5°N , 23°W appears to be rather weak and it is likely that there are
193 different generation mechanisms for the strong eastward flow events. The short period
194 of intraseasonal events implies that the upwelling within the Guinea Dome in the eastern
195 basin as suggested by McCreary et al. (2002) and Furue et al. (2007, 2009) may not be
196 of first order in forcing them. To investigate the role of the wind forcing for the strong
197 eastward flow events we linearly regressed zonal wind stress anomalies (Bentamy & Fillon,
198 2012; Kobayashi et al., 2015; Large & Yeager, 2004) onto the NEUC transport time series
199 (Fig. S6, Text S3). Easterly wind stress anomalies in the eastern equatorial basin are
200 leading the NEUC transports by one to two months. We hypothesize that these zonal wind
201 stress anomalies may force equatorial Kelvin waves reflecting at the eastern boundary into
202 westward propagating Rossby waves and poleward propagating coastal trapped waves. The
203 northward propagating waves may shed Rossby waves when arriving at the exit of the Gulf
204 of Guinea where the coastline turns northward. Such remotely forced Rossby waves may
205 reach 5°N , 23°W , causing the observed eastward flow events. Rossby waves at 5°N , 23°W
206 can also be forced locally by wind stress curl anomalies of small meridional scale (Burmeister
207 et al., 2016; Foltz et al., 2010) or remotely generated by the radiation of Rossby waves from
208 coastal trapped waves generated by local wind anomalies in the Gulf of Guinea (Chu et
209 al., 2007). However, in a composite analysis of sea level anomalies we could not clearly

210 identify a reflection of an equatorial Kelvin wave at the eastern boundary or a Rossby wave
 211 propagating from the eastern basin to 5°N, 23°W prior to any eastward flow event.

212 **6 NEUC and oxygen**

213 The NEUC is thought to transport oxygen rich water from the western boundary to-
 214 wards the generally poorly ventilated eastern tropical North Atlantic (Brandt et al., 2010;
 215 Stramma et al., 2008). The mean ship section indicates a local oxygen maximum which is
 216 associated with the NEUC and single ship sections often show maxima in the area of the
 217 NEUC (Fig. 3). Here we will investigate if the strong eastward NEUC events are associated
 218 with an increased eastward oxygen transport using moored observations.

227 We calculated oxygen anomalies on isopycnals. First, a temporal mean oxygen profile
 228 as a function of density was calculated from the mooring time series. Next, for each time
 229 step, the oxygen anomaly was calculated with respect to the mean oxygen value of the
 230 respective density. Finally we applied a second-order 30-day low-pass Butterworth filter to
 231 both, the transport and oxygen time series (Fig. 3).

232 In 100 m and 200 m depth, we found a significant positive correlation between the
 233 NEUC transport and oxygen anomalies ($R_{\max,100m} = 0.35$, NEUC leads by 16 days;
 234 $R_{\max,200m} = 0.35$, NEUC leads by 2.5 days), while in 300 m depth, which is just below
 235 the NEUC, the correlation is not significant ($R_{300m} = 0.05$, zero lag). If data is available,
 236 positive oxygen anomalies can be found during most of the strong eastward flow events
 237 except for in 100 m depth during the 2012 and 2014 event. 2017 seems to be an exceptional
 238 year in the moored time series with the two strongest eastward flow events occurring con-
 239 secutively within six months. Additionally, these two events are associated with some of the
 240 highest oxygen anomalies observed. The correlation between oxygen and NEUC transport
 241 anomalies during these two events is much higher compared to the rest of the time series.
 242 The correlation between the time series for the period from September 2016 to February
 243 2018 is 0.63 (NEUC leads by 13.5 days) and 0.56 (NEUC leads by 1.5 days) in 100 m and
 244 200 m depth, respectively.

245 Positive oxygen anomalies also occur independently from strong eastward flow events
 246 and may be associated with changes in the meridional velocity. On intraseasonal time
 247 scales, zonal and meridional velocity anomalies are of similar strength and the dominant
 248 frequency band of the meridional velocity (30 to 70 days) overlaps with the frequency of

249 the strong eastward flow events (one to five months). It is thus not possible to clearly
 250 differentiate between the effect of meridional and zonal velocity on oxygen values at the
 251 mooring position. Yet, we find no significant correlation between the 30-day low-pass filtered
 252 meridional velocity and oxygen anomalies ($-0.12 \leq R \leq 0.06$ at zero lag) (Fig. S7).

253 7 Summary and Discussion

254 The NEUC is an eastward flowing current centered around 5°N in the tropical Atlantic.
 255 Although its core velocity is weak (below 0.3 ms^{-1}) it is thought to act as an important
 256 oxygen supply route towards the eastern tropical North Atlantic oxygen minimum zone
 257 (Brandt et al., 2010; Stramma et al., 2008). For the first time we reconstructed a time
 258 series of the NEUC transport based purely on direct velocity observations. By combining
 259 moored zonal velocities at 5°N , 23°W and meridional ship sections along $\sim 23^{\circ}\text{W}$ we obtained
 260 a NEUC transport time series from June 2006 to February 2008 and November 2009 to
 261 January 2018. In contrast to previous studies (Burmeister et al., 2019; Goes et al., 2013;
 262 Hüttl-Kabus & Böning, 2008), neither the moored zonal velocity at 5°N (Fig. 1) nor the
 263 reconstructed eastward NEUC transport (Fig. 2) exhibits a pronounced seasonal cycle.
 264 We find that both time series are dominated by strong intraseasonal eastward flow events
 265 which can peak throughout the year (Fig. 1 and 2). Although in the mean meridional
 266 ship sections the zonal velocity maximum associated with the NEUC coincides with a local
 267 oxygen maximum, we find that the eastward flow events are only occasionally associated
 268 with high oxygen levels (Fig. 3).

269 We reconstructed the NEUC transport at 23°W based on velocity observations from
 270 three moorings and 24 meridional ship sections (Fig. 2). To obtain a longer time series we
 271 then reconstructed the NEUC variability based on data from only one mooring position.
 272 The comparison with the three-mooring solution indicates that the variability is dominated
 273 by a meridionally homogeneous structure covering the complete integration box (Fig. S3).
 274 However, the method has some limitations. The used integration box does not cover the
 275 entire NEUC region (Fig. S1). Consequently, our method is underestimating the true NEUC
 276 transport. A comparison of the box integrated transport calculated from ship sections with
 277 a reference transport calculated from ship sections using a path following algorithm indicates
 278 that the box integrated transport represents the variability of the NEUC reasonably well
 279 ($R=0.9$, Fig. S2 and S5). An additional uncertainty of the reconstructed transport is
 280 due to the upper range of the moored velocity observations that varies between 65 m and

281 85 m, i.e. the moored observations do not always cover the entire NEUC depth range. In
282 summary, the used method is underestimating the true NEUC strength but it is still capable
283 of representing its variability, as this is dominated by a homogeneous structure in latitude
284 and depth.

285 Within the scope of this paper we could not identify a clear forcing mechanism for the
286 strong eastward flow events. We found that easterly wind stress anomalies in the eastern
287 equatorial basin are leading the NEUC transport by one to two months. Zonal wind stress
288 anomalies and associated wind stress curl anomalies could directly (Burmeister et al., 2016;
289 Foltz et al., 2010) or remotely (Chu et al., 2007) force Rossby waves which reach 5°N , 23°W
290 to cause the observed eastward flow events. However, we could not clearly identify a Rossby
291 wave propagating from the eastern basin to 5°N , 23°W prior to any event. Other potential
292 mechanisms might include nonlinear processes resulting e.g. in the development of multiple
293 cores that were discussed by Furue et al. (2007). These cores that were superimposed on the
294 otherwise linear arrested front dynamics (Dewar et al., 1991) were identified in simulations
295 with enhanced horizontal resolution.

296 Finally, we investigated the relationship between the NEUC transport events and oxy-
297 gen. In 2017, eastward transport and oxygen anomalies agree very well. For other events
298 the correlation is weaker or non-existent (Fig. 3). One possible explanation is that the local
299 oxygen maximum associated with the NEUC is not located directly at 5°N as visible in the
300 meridional ship sections of March 2017 (Fig. 3b,e). In that case the higher oxygen concen-
301 trations are simply not captured by the mooring. Another explanation may be that some
302 strong eastward flow events are not connected to the well-ventilated western boundary, but
303 are instead supplied out of the westward flow of low-oxygen waters north and south of the
304 NEUC. For example, in the meridional section of February 2018, we find strong eastward
305 velocities between 4°N and 5°N at 50 m and 100 m depth that are associated with low oxygen
306 concentrations, which might be due to a recirculation of low-oxygen waters typically present
307 south of the NEUC (Fig. 3c and f; Burmeister et al., 2019). As the zonal velocity events
308 with a duration between one and five months overlap with the dominant frequency band
309 of meridional velocity (30 to 70 days) it is not possible to clearly differentiate between the
310 effect of meridional and zonal velocity on oxygen. However, we did not find any correlation
311 between the meridional velocity and oxygen anomalies on intraseasonal time scales at 5°N ,
312 23°W . Furthermore, Hahn et al. (2014) found that the mean meridional eddy flux at the
313 mooring position is not significantly different from zero. Consistent with previous studies

314 (Burmeister et al., 2019; Hahn et al., 2014; Weisberg & Weingartner, 1988), we suggest
315 that a continuous flow of oxygen-rich water from the western boundary toward the eastern
316 Atlantic basin by the NEUC is regularly altered by TIWs or other mesoscale recirculations.
317 This can explain why not all NEUC transport events are associated with higher oxygen
318 levels. Nevertheless, the significant positive correlation between the 30-day low-pass filtered
319 NEUC transport and oxygen anomalies at 5°N, 23°N (Fig. 3) indicates that the NEUC
320 transport events likely result in an elevated mean eastward oxygen transport and presum-
321 ably contribute to sustain the oxygen maximum observed in ship sections at the NEUC core
322 position.

323 **Acknowledgments**

324 This study was funded by the Deutsche Forschungsgemeinschaft as part of the Sonder-
325 forschungsbereich 754 “Climate-Biogeochemistry Interactions in the Tropical Ocean,” through
326 several research cruises with RV L’Atalante, RV Maria S. Merian, RV Meteor, and RV Po-
327 larstern, by the project FOR1740 and by the Deutsche Bundesministerium für Bildung und
328 Forschung (BMBF) as part of projects RACE (03F0651B) and RACE-Synthesis (03F0824C).
329 We thank the captains, crews, scientists, and technical groups involved in the different na-
330 tional and international research cruises to the eastern tropical North Atlantic that con-
331 tributed to collecting CTD, velocity as well as mooring data, and making them freely avail-
332 able. We thank Rebecca Hummels for post-processing of the recent mooring and ship
333 section data. The shipboard data are accessible at [https://doi.pangaea.de/10.1594/](https://doi.pangaea.de/10.1594/PANGAEA.899052)
334 [PANGAEA.899052](https://doi.pangaea.de/10.1594/PANGAEA.899052). The mooring data are accessible at [https://doi.pangaea.de/10.1594/](https://doi.pangaea.de/10.1594/PANGAEA.903913)
335 [PANGAEA.903913](https://doi.pangaea.de/10.1594/PANGAEA.903913). We are grateful to NOAA/PMEL and NOAA/AOML for making the data
336 of the PIRATA Northeast Extension cruises freely available at [https://www.aoml.noaa](https://www.aoml.noaa.gov/phod/pne/cruises.php)
337 [.gov/phod/pne/cruises.php](https://www.aoml.noaa.gov/phod/pne/cruises.php). ASCAT data were obtained from the Centre de Recherche et
338 d’Exploitation Satellitaire (CERSAT), at IFREMER, Plouzané (France) and are available at
339 <ftp://ftp.ifremer.fr/ifremer/cersat/products/gridded/MWF/L3/ASCAT/Daily/>. Global
340 Ocean Gridded L4 sea surface heights and derived variables were made available by E.U.
341 Copernicus Marine Environment Monitoring Service (CMEMS). The data is available at
342 [http://marine.copernicus.eu/services-portfolio/access-to-products/?option=com](http://marine.copernicus.eu/services-portfolio/access-to-products/?option=com_csw&view=details&product_id=SEALEVEL_GLO_PHY_L4_REP_OBSERVATIONS_008_047)
343 [_csw&view=details&product_id=SEALEVEL_GLO_PHY_L4_REP_OBSERVATIONS_008_047](http://marine.copernicus.eu/services-portfolio/access-to-products/?option=com_csw&view=details&product_id=SEALEVEL_GLO_PHY_L4_REP_OBSERVATIONS_008_047). The
344 JRA-55 reanalysis surface winds used for this study are provided by the Japanese 55-year
345 Reanalysis (JRA-55) project carried out by the Japan Meteorological Agency (JMA) and

346 is available at https://jra.kishou.go.jp/JRA-55/index_en.html#jra-55 after registra-
347 tion.

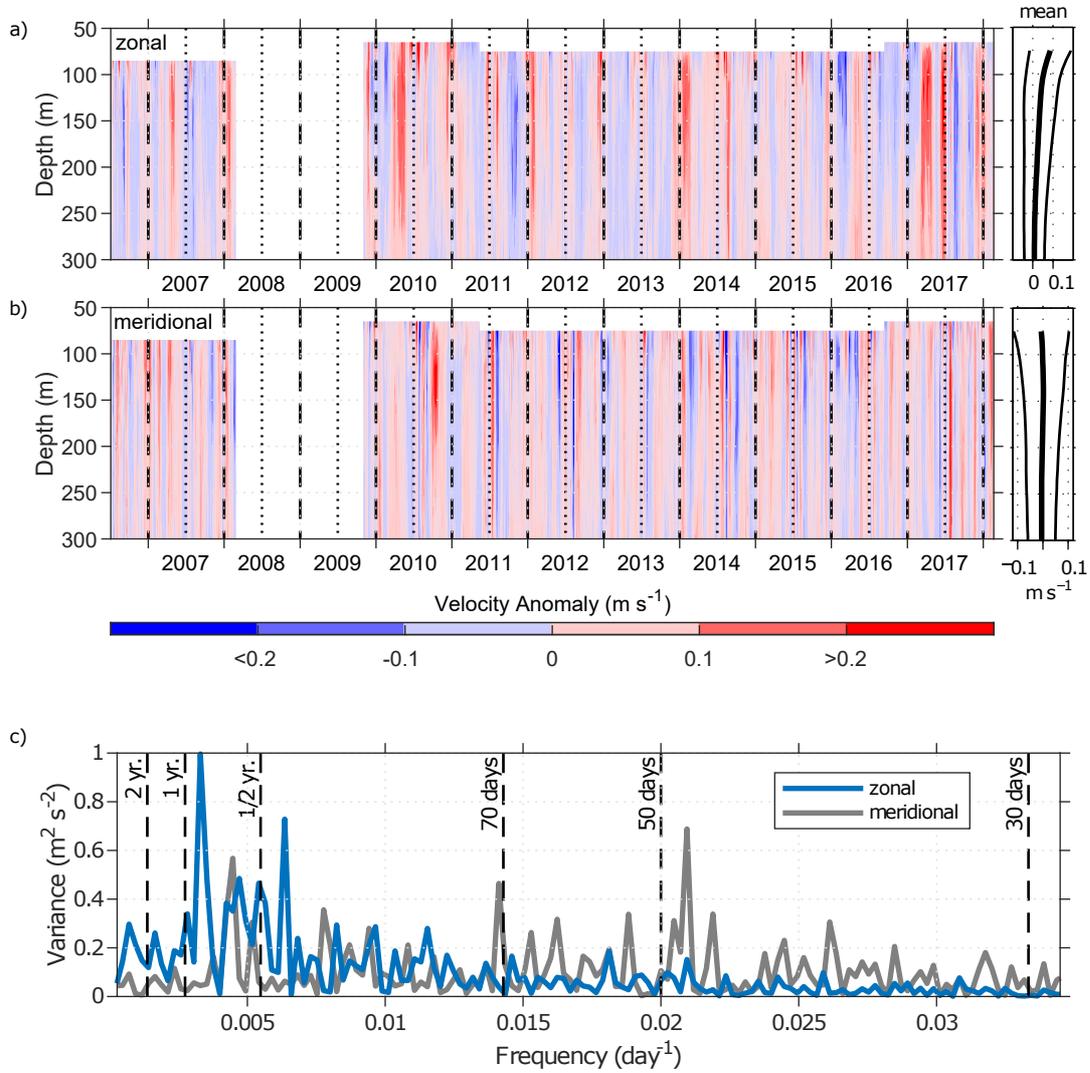
348 References

- 349 Bentamy, A., & Fillon, D. C. (2012). Gridded surface wind fields from Metop/ASCAT
350 measurements. *Int. J. Remote Sens.*, *33*(6), 1729–1754. doi:
351 10.1080/01431161.2011.600348
- 352 Bourlès, B., D’Orgeville, M., Eldin, G., Gouriou, Y., Chuchla, R., DuPenhoat, Y., &
353 Arnault, S. (2002). On the evolution of the thermocline and subthermocline
354 eastward currents in the Equatorial Atlantic. *Geophys. Res. Lett.*, *29*(16),
355 32–1–32–4. doi: 10.1029/2002GL015098
- 356 Bourlès, B., Gouriou, Y., & Chuchla, R. (1999). On the circulation in the upper layer of
357 the western equatorial Atlantic. *J. Geophys. Res. Ocean.*, *104*(C9), 21151–21170.
358 doi: 10.1029/1999JC900058
- 359 Brandt, P., Bange, H. W., Banyte, D., Dengler, M., Didwischus, S.-H., Fischer, T.,
360 Greatbatch, R. J., Hahn, J., Kanzow, T., Karstensen, J., Körtzinger, A., Krahnmann,
361 G., Schmidtko, S., Stramma, L., Tanhua, T., & Visbeck, M. (2015). On the role of
362 circulation and mixing in the ventilation of oxygen minimum zones with a focus on
363 the eastern tropical North Atlantic. *Biogeosciences*, *12*(2), 489–512. doi:
364 10.5194/bg-12-489-2015
- 365 Brandt, P., Funk, A., Tantet, A., Johns, W. E., & Fischer, J. (2014). The Equatorial
366 Undercurrent in the central Atlantic and its relation to tropical Atlantic variability.
367 *Clim. Dyn.*, *43*(11), 2985–2997. doi: 10.1007/s00382-014-2061-4
- 368 Brandt, P., Hormann, V., Körtzinger, A., Visbeck, M., Krahnmann, G., Stramma, L., . . .
369 Schmid, C. (2010). Changes in the Ventilation of the Oxygen Minimum Zone of the
370 Tropical North Atlantic. *J. Phys. Oceanogr.*, *40*(8), 1784–1801. doi:
371 10.1175/2010JPO4301.1
- 372 Burmeister, K., Lübbecke, J. F., & Brandt, P. (2016). Revisiting the cause of the eastern
373 equatorial Atlantic cold event in 2009. *J. Geophys. Res. Ocean.*, *7*(121), 4777–4789.
374 doi: 10.1002/2016JC011719
- 375 Burmeister, K., Lübbecke, J. F., Brandt, P., & Duteil, O. (2019). Interannual variability
376 of the Atlantic North Equatorial Undercurrent and its impact on oxygen. *J.*
377 *Geophys. Res. Ocean.*, *124*(4), 2348–2373. doi: 10.1029/2018JC014760

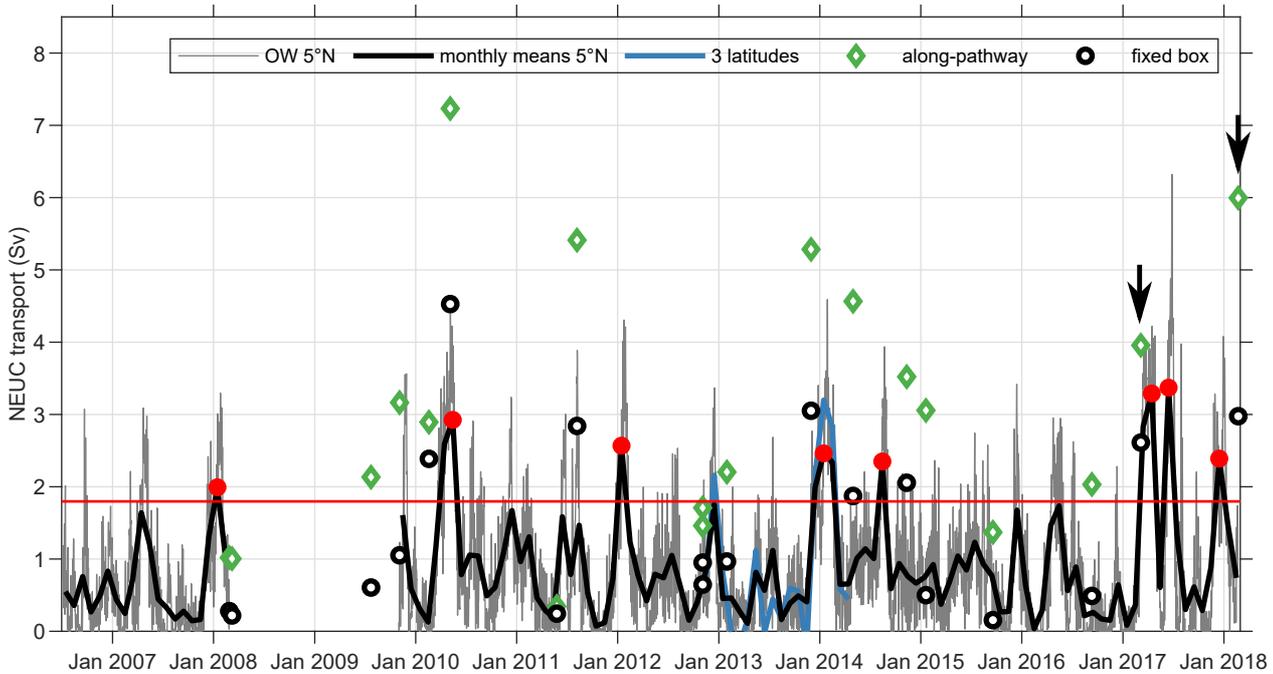
- 378 Chu, P. C., Ivanov, L. M., Melnichenko, O. V., & Wells, N. C. (2007). On long baroclinic
379 Rossby waves in the tropical North Atlantic observed from profiling floats. *J.*
380 *Geophys. Res.*, *112*(C5), C05032. doi: 10.1029/2006JC003698
- 381 Dewar, W. K. (1991). Arrested fronts. *J. Mar. Res.*, *49*, 21–52. doi:
382 10.1357/002224091784968576
- 383 Fischer, J., Brandt, P., Dengler, M., Müller, M., & Symonds, D. (2003). Surveying the
384 upper ocean with the ocean surveyor: A new phased array Doppler current profiler.
385 *J. Atmos. Ocean. Technol.*, *20*(5), 742–751. doi:
386 10.1175/1520-0426(2003)20<742:STUOWT>2.0.CO;2
- 387 Foltz, G. R., & McPhaden, M. J. (2010). Interaction between the Atlantic meridional and
388 Niño modes. *J. Geophys. Res.*, *37*(18), L18604. doi: 10.1029/2010GL044001
- 389 Furue, R., McCreary Jr., J. P., Yu, Z., & Wang, D. (2007). Dynamics of the Southern
390 Tsuchiya Jet*. *J. Phys. Ocean.*, *37*(3), 531–553. doi: 10.1175/JPO3024.1
- 391 Furue, R., McCreary Jr., J. P., & Yu, Z. (2009). Dynamics of the Northern Tsuchiya
392 Jet*. *J. Phys. Oceanogr.*, *39*(9), 2024–2051. doi: 10.1175/2009JPO4065.1
- 393 Goes, M., Goni, G., Hormann, V., & Perez, R. C. (2013). Variability of the Atlantic
394 off-equatorial eastward currents during 1993–2010 using a synthetic method. *J.*
395 *Geophys. Res. Ocean.*, *118*(6), 3026–3045. doi: 10.1002/jgrc.20186
- 396 Hahn, J., Brandt, P., Greatbatch, R. J., Krahlmann, G., & Körtzinger, A. (2014). Oxygen
397 variance and meridional oxygen supply in the Tropical North East Atlantic oxygen
398 minimum zone. *Clim. Dyn.*, *43*(11), 2999–3024. doi: 10.1007/s00382-014-2065-0
- 399 Hahn, J., Brandt, P., Schmidtko, S., & Krahlmann, G. (2017). Decadal oxygen change in
400 the eastern tropical North Atlantic. *Ocean Sci.*, *13*(4), 551–576. doi:
401 10.5194/os-13-551-2017
- 402 Hazeleger, W., & Drijfhout, S. (2006). Subtropical cells and meridional overturning
403 circulation pathways in the tropical Atlantic. *J. Geophys. Res. Ocean.*, *111*(3), 1–13.
404 doi: 10.1029/2005JC002942
- 405 Hsin, Y. C., & Qiu, B. (2012). Seasonal fluctuations of the surface North Equatorial
406 Countercurrent (NECC) across the Pacific basin. *J. Geophys. Res. Ocean.*, *117*(6),
407 1–17. doi: 10.1029/2011JC007794
- 408 Hüttel-Kabus, S., & Böning, C. W. (2008). Pathways and variability of the off-equatorial
409 undercurrents in the Atlantic Ocean. *J. Geophys. Res. Ocean.*, *113*(10), 1–14. doi:
410 10.1029/2007JC004700

- 411 Jochum, M., & Malanotte-Rizzoli, P. (2003). On the generation of North Brazil Current
412 rings. *J. Mar. Res.*, *61*(2), 147–173. doi: 10.1357/002224003322005050
- 413 Jochum, M., & Malanotte-Rizzoli, P. (2004). A New Theory for the Generation of the
414 Equatorial Subsurface Countercurrents. *J. Phys. Oceanogr.*, *34*(4), 755–771. doi:
415 10.1175/1520-0485(2004)034<0755:ANTFTG>2.0.CO;2
- 416 Kobayashi, S., Ota, Y., Harada, Y., Ebata, A., Ota, Y., Moriya, M., Onoda, H., Onogi, K.,
417 Kamahori, H., Kobayashi, C., Endo, H., Miyaoka, K., & Takahashi, K. (2015). The
418 JRA-55 Reanalysis: General Specifications and Basic Characteristics. *J. Met. Soc.
419 Jap.*, *93*(1), 5–48. doi: 10.2151/jmsj.2015-001
- 420 Large, W. G., & Yeager, S. (2004). Diurnal to decadal global forcing for ocean and sea-ice
421 models: The data sets and flux climatologies. *NCAR Technical Note
422 NCAR/TN-460+STR*. doi: 10.5065/D6KK98Q6
- 423 Marin, F., Hua, B. L., & Wacogne, S. (2000). The equatorial thermostat and subsurface
424 countercurrents in the light of the dynamics of atmospheric Hadley cells. *J. Mar.
425 Res.*, *58*(3), 405–437. doi: 10.1357/002224000321511098
- 426 McCreary Jr., J. P., Lu, P., & Yu, Z. (2002). Dynamics of the Pacific Subsurface
427 Countercurrents. *J. Phys. Oceanogr.*, *32*, 2379–2404. doi:
428 10.1175/1520-0485(2002)032<2379:DOTPSC>2.0.CO;2
- 429 Oschlies, A., Brandt, P., Stramma, L., & Schmidtko, S. (2018). Drivers and mechanisms
430 of ocean deoxygenation. *Nat. Geosci.*, *11*(7), 467–473. doi:
431 10.1038/s41561-018-0152-2
- 432 Schott, F. A., Dengler, M., Brandt, P., Affler, K., Fischer, J., Bourlès, B., . . . Rhein, M.
433 (2003). The zonal currents and transports at 35W in the tropical Atlantic. *Geophys.
434 Res. Lett.*, *30*(7), 35–38. doi: 10.1029/2002GL016849
- 435 Schott, F. A., McCreary Jr., J. P., & Johnson, G. C. (2004). Shallow Overturning
436 Circulations of the Tropical-Subtropical Oceans. *Earth's Clim.*, *147*, 261–304.
- 437 Schott, F. A., Stramma, L., & Fischer, J. (1995). The warm water inflow into the western
438 tropical Atlantic boundary regime , spring 1994. *J. Geophys. Res.*, *100*(C12), 24745
439 – 24760. doi: 10.1029/95JC02803
- 440 Stramma, L., Brandt, P., Schafstall, J., Schott, F., Fischer, J., & Körtzinger, A. (2008).
441 Oxygen minimum zone in the North Atlantic south and east of the Cape Verde
442 Islands. *J. Geophys. Res. Ocean.*, *113*(4), 1–15. doi: 10.1029/2007JC004369
- 443 Urbano, D. F., De Almeida, R. A., & Nobre, P. (2008). Equatorial Undercurrent and

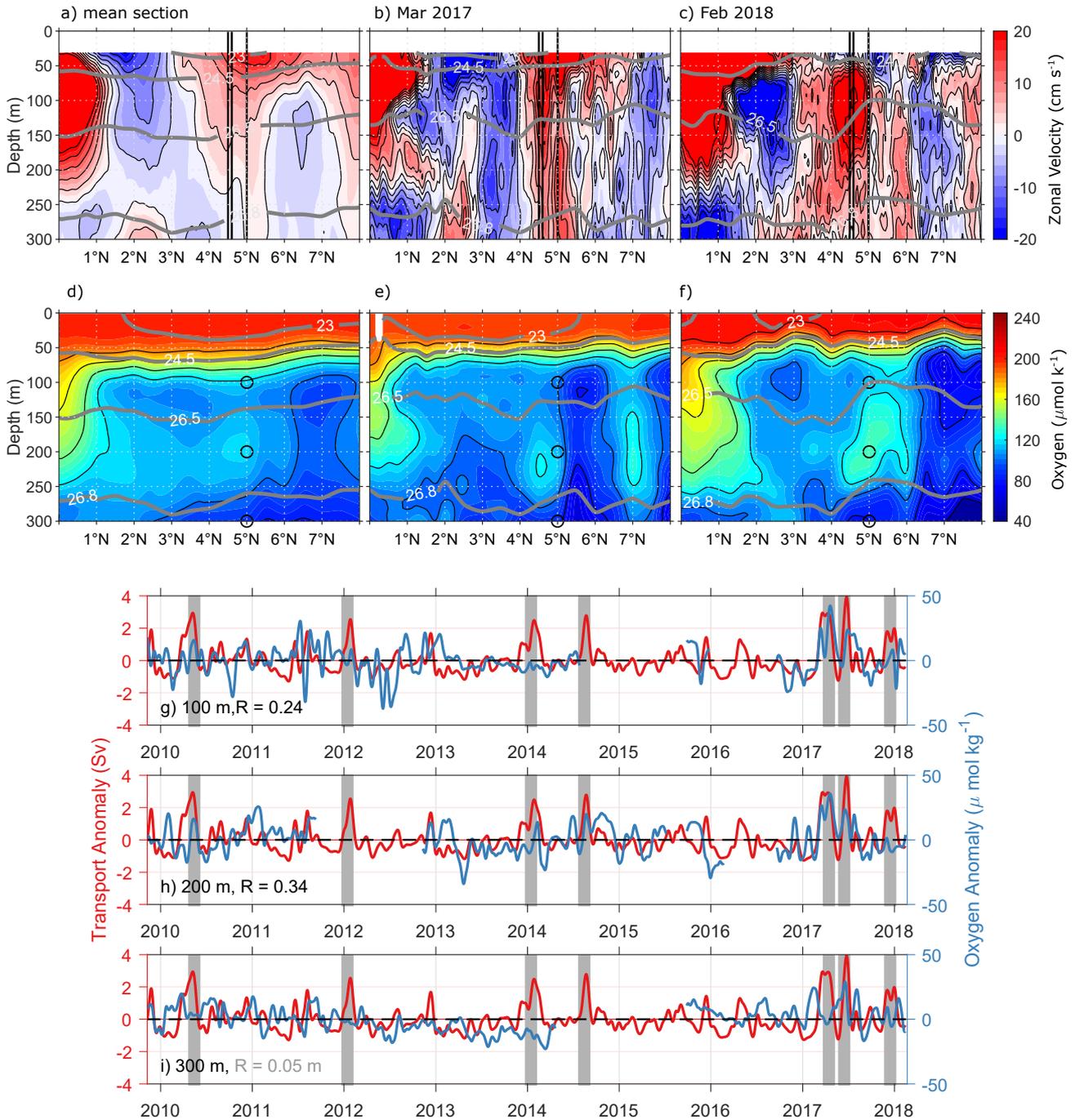
444 North Equatorial Countercurrent at 38°W: A new perspective from direct velocity
445 data. *J. Geophys. Res. Ocean.*, 113(4), 1–16. doi: 10.1029/2007JC004215
446 Weisberg, R. H., & Weingartner, T. J. (1988). Instability Waves in the Equatorial
447 Atlantic Ocean. *J. Phys. Oceanogr.*, 18(11), 1641–1657. doi:
448 10.1175/1520-0485(1988)018<1641:IWITEA>2.0.CO;2



118 **Figure 1.** Anomalous (a) zonal and (b) meridional velocity measurements from moored ADCPs
 119 at 5°N, 23°W. The temporal mean profiles (thick black line) are shown at the right side of the (a)
 120 zonal and (b) meridional velocity time series. The thin black line marks the mean profile ± one
 121 standard deviation. (c) Lomb-Scargle periodogram of zonal (blue line) and meridional (grey line)
 122 velocity averaged between 95 m and 275 m depth.



172 **Figure 2.** NEUC transport at 23°W calculated by four different methods: (i) from ship observa-
 173 tions using a path following algorithm (green diamonds); (ii) from ship sections by integrating the
 174 eastward velocities in a fixed box (black circles); (iii) by the OW method combining ship sections
 175 and moored zonal velocities at three mooring positions (monthly means, blue line); (iv) by the OW
 176 method combining ship sections with moored zonal velocities at 5°N , 23°W (gray line). The black
 177 thick line shows the monthly mean values of the NEUC transport reconstructed at 5°N . The red
 178 line marks 2.5 times the standard deviation of the monthly mean transports used to define strong
 179 transport events (red dots). The black arrows mark the month of the cruises in 2017 and 2018.



219 **Figure 3.** (a-f) Zonal velocity (a-c) and oxygen (d-f) observations along 23°W with mean sections
 220 of all 24 cruises (a,d) and sections taken during Ronald H. Brown cruise PNE 2017 (b,e) and during
 221 Meteor cruise M145 (c,f). Grey lines mark neutral density surfaces (kg m^{-3}), black vertical lines
 222 (a-c) mark the position of moorings, black circles (d-f) mark single point oxygen measurements.
 223 (g-i) 30-day low-pass filtered NEUC transport (red lines) and oxygen anomalies (blue lines) at 5°N,
 224 23°W at a depth of (g) 100 m, (h) 200 m, and (i) 300 m. Grey bars mark strong NEUC events. R
 225 is the correlation coefficient of zonal velocity and oxygen anomalies at zero lag (black/grey colour
 226 indicate that R is significant/not significant on a 95% confidence level).