



HAL
open science

A dynamical interpretation of the intensification of the winter North Atlantic jet stream in reanalysis

Alejandro Hermoso, Gwendal Rivière, Ben Harvey, John Methven, Sebastian Schemm

► **To cite this version:**

Alejandro Hermoso, Gwendal Rivière, Ben Harvey, John Methven, Sebastian Schemm. A dynamical interpretation of the intensification of the winter North Atlantic jet stream in reanalysis. *Journal of Climate*, 2024, 10.1175/JCLI-D-23-0757.1 . hal-04740036

HAL Id: hal-04740036

<https://hal.science/hal-04740036v1>

Submitted on 16 Oct 2024

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **A dynamical interpretation of the intensification of the winter North**
2 **Atlantic jet stream in reanalysis**

3 Alejandro Hermoso,^a Gwendal Rivière,^b Ben Harvey,^c ^d John Methven,^c and Sebastian
4 Schemm,^a

5 ^a *Institute for Atmospheric and Climate Science, ETH Zurich, Zurich, Switzerland*

6 ^b *LMD/IPSL, École Normale Supérieure, PSL Research University, Sorbonne Université, École*
7 *Polytechnique, CNRS, Paris, France*

8 ^c *Department of Meteorology, University of Reading, Reading, UK*

9 ^d *National Centre for Atmospheric Science, University of Reading, Reading, UK*

11 ABSTRACT: Jet streams play an important role in determining weather variability and extremes.
12 A better understanding of the mechanisms driving long-term changes in the jet is essential to
13 successfully anticipate extreme meteorological events. This study analyzes the intensification trend
14 of the North Atlantic jet using the ERA5 reanalysis and investigates the dynamical mechanisms
15 involved. The results highlight the importance of an increase in diabatic heating in the free
16 troposphere below the jet entrance over the Gulf Stream sector. This change in diabatic heating
17 modifies the jet directly and produces a local intensification and a slight poleward shift. A two-
18 dimensional frontal-geostrophic model illustrates this mechanism by considering the enhanced
19 diabatic heating associated with the baroclinic growth of extratropical cyclones. The change in
20 diabatic heating also affects the jet indirectly by increasing the mean baroclinicity and subsequent
21 eddy momentum flux convergence. This indirect mechanism has also an effect downstream, where
22 there is an acceleration of the jet core and reduced westerlies along the flanks, reducing the width
23 of the jet. An idealized warming experiment confirms this mechanism by determining the jet
24 response downstream of an idealized land-sea contrast. Finally, using a single-model ensemble
25 of fully-coupled climate simulations, we show that the differences in the evolution of the North
26 Atlantic jet are related to the latitude of the increase in baroclinicity, which has a large spread.
27 What emerges from the model hierarchy is a consistent dynamical chain of mechanisms associated
28 with the intensification trend of the North Atlantic jet stream.

29 **1. Introduction**

30 The processes that determine the position and intensity of jet streams, which drive daily weather
31 variability and contribute to the development of extreme weather events, are an ongoing area of
32 research (Shaw et al. 2016), particularly with regard to open questions about the future evolution of
33 the North Atlantic jet (Woollings et al. 2012). The high degree of uncertainty regarding the future
34 intensity and position of the North Atlantic jet translates directly into a high degree of uncertainty
35 with regard to weather variability. For example, there is not yet a significant consensus on the sign
36 of the projected annual precipitation change for 1.5° C and 2.0° C of warming (e.g., Fig. 4.32 in
37 Chapter 4 of the IPCC AR6 (Lee et al. 2021)), which is likely attributable to a low signal to noise
38 ratio. In particular, over the North Atlantic, Central, and Northern Europe, the sign of the annual
39 precipitation changes remains, in contrast to the Mediterranean region, inconclusive because of a
40 large inter-model spread (e.g., Fig. 4.42 e and Fig. 4.42 f in Chapter 4 of the IPCC AR6 (Lee et al.
41 2021)).

42 Baroclinicity measures the baroclinic growth potential of transient eddies (Charney 1947; Eady
43 1949). It is proportional to the horizontal temperature gradient - and by virtue of thermal wind
44 balance thus to the vertical wind shear - and inversely proportional to the vertical temperature
45 gradient, which is a measure of atmospheric stability (Lindzen and Farrell 1980). Baroclinic
46 growth of transient eddies in turn maintains deep westerly tropospheric jets against surface friction
47 through associated eddy momentum flux convergence (Hoskins et al. 1983; Vallis 2017). Due
48 to this inherent link between mean baroclinicity, baroclinic growth, and eddy momentum flux
49 convergence, the analysis of the life cycle of eddy-driven jets alongside baroclinicity and its forcing
50 processes (e.g. latent heat release) becomes crucial for understanding regional climate changes
51 in areas affected by storm tracks. Ambaum and Novak (2014) proposed a nonlinear oscillator
52 model to describe the joint evolution of diabatic heating, mean baroclinicity and baroclinic growth.
53 The proposed life cycle displays qualitative agreement to observations with peaks of intense
54 heat flux indicative for intense storm activity that acts to reduce the mean baroclinicity. When
55 baroclinicity becomes too low to maintain the baroclinic eddy generation, storm activity is reduced
56 and baroclinicity is replenished by diabatic heating influenced for instance by the orography or a
57 land-sea contrast (Brayshaw et al. 2011), allowing the cycle to be repeated. Increased diabatic heat
58 release (e.g. as expected in a warmer atmosphere) and/or modified temperature gradients (e.g. as

59 expected at low levels due to Arctic amplification and at upper levels due to tropical warming) thus
60 could both influence the life cycle of the storm track by changing the mean baroclinicity, which
61 would also affect the jet via eddy momentum feedback.

62 In the Northern Hemisphere, two competing trends due to anthropogenic climate change act on
63 the equator-to-pole temperature gradient. Near the surface, the meridional temperature gradient is
64 decreased due to amplified polar warming, a phenomenon known as the Arctic Amplification (AA)
65 (Screen and Simmonds 2010). In contrast, in the upper troposphere tropical warming increases
66 the meridional temperature gradient and reduces tropospheric stability. This larger temperature
67 increase in the tropics is produced because the atmosphere follows a moist adiabatic lapse rate
68 (Held 1993). In a warmer climate, the amount of water vapor in the atmosphere is higher and
69 consequently there is more latent heat release, which reduces the lapse rate, since the adiabatic
70 cooling resulting from the ascent is partially offset by higher latent heating. Jet streams are by virtue
71 of the thermal wind balance related to meridional temperature gradients. Therefore, the upper and
72 lower-level temperature trends are thus engaged in a tug-of-war around the future evolution of the
73 jet stream and its associated storm track (Shaw et al. 2016).

74 Besides forced trends, storm tracks and eddy-driven jets exhibit substantial temporal and spatial
75 variability (Hartmann 2007). Their activity is larger during winter when the equator-to-pole
76 temperature gradient is higher and consequently the energy transport is more intense compared
77 to summer. A remarkable exception to this behaviour is found in the North Pacific, where storm
78 activity is lower in mid-winter (Nakamura 1992; Schemm and Rivière 2019; Schemm et al. 2021).
79 The position of the storm tracks has also a seasonal cycle whereby storm tracks and jet streams
80 are located at lower latitudes during winter. Furthermore, they display high frequency variations
81 in intensity and position from shorter time scales of the order of weeks, related to oscillatory
82 behaviour of the regime life cycle produced by consumption and replenishment of baroclinicity,
83 to multidecadal variations. Indeed, storm tracks can be affected by tropical variability across
84 different scales, from sub-seasonal scales produced by the Madden-Julian Oscillation (Yadav and
85 Straus 2017) to decadal variations generated by El Niño-Southern Oscillation (Eichler and Higgins
86 2006; Schemm et al. 2018) and multidecadal variability (Bracegirdle et al. 2018).

87 Jet stream trends are therefore affected by both the impacts of anthropogenic warming on
88 circulation and natural variability. Previous studies based on reanalysis data have found strong

89 seasonal and regional variations in jet changes over the past decades with limited robust trends
90 in wind speed or position (Manney and Hegglin 2018) and in particular for the subtropical jet no
91 robust poleward trends have been identified in reanalyses (Maher et al. 2020). Simmons (2022)
92 reported an intensification and equatorward shift of the jet stream over the eastern North Atlantic
93 related to a warming minimum between Greenland and Europe in the ERA5. Martin (2021) found
94 increased waviness of the eddy-driven and subtropical jets and a poleward shift of the eddy-driven
95 jet while Woollings et al. (2023) highlighted the role of upper tropospheric warming to explain the
96 poleward shift of jet streams in both hemispheres, but did not look at the North Atlantic region in
97 particular.

98 Climate model projections exhibit large uncertainty concerning the future of the wintertime
99 eddy-driven jet on the North Atlantic compared to the jets over other ocean basins (Simpson et al.
100 2014; Lorenz and DeWeaver 2007; Woollings 2010), which is potentially linked to the nature of
101 the North Atlantic as a mostly eddy-driven jet stream (Lee and Kyung Kim 2003; Li and Wettstein
102 2012). These uncertainties have a strong influence on the low confidence in future changes in the
103 hydrological cycle and the occurrence of extreme wind events, since the vast majority of heavy
104 precipitation and intense wind events during the cold season are associated with extratropical
105 cyclones (Pfahl and Wernli 2012; Owen et al. 2021). For Europe, which lies at the end of the
106 North Atlantic storm track, changes in the jet stream and storm tracks are thus particularly relevant.
107 Therefore, future European rainfall and wind will be significantly affected by the direction and
108 magnitude of any meridional shift in the storm track and/or position of the jet stream (Priestley and
109 Catto 2022). Currently, it is only settled that sea surface temperatures (SST) and their influence
110 on low-level baroclinicity account for a fraction of the change (Woollings et al. 2012) but the
111 interplay between changes in baroclinicity, diabatic heating, storm tracks and the jet stream and the
112 chain of processes that explain past changes in the North Atlantic jet stream have not been clearly
113 established.

114 The atmospheric response to increased diabatic heating is relevant to understand changes in the
115 North Atlantic jet. In this regard, Peng and Whitaker (1999) investigate the response to diabatic
116 heating by using a linear baroclinic model, which is a primitive equation model linearized about a
117 basic state including heating and eddy terms as forcing. The response consists of a low-level trough,
118 whose extension depends on the heating distribution, and an upper-level ridge downstream. The

119 imposed heating is interpreted as the initial heating before the atmosphere has adjusted to it. The
120 resulting anomalous flow is inserted into to a linear storm track model, which is a quasigeostrophic
121 model linearized about a time-mean flow¹ to deduce the anomalous eddy forcing, which in turn
122 is re-inserted into the linearized baroclinic model to obtain the eddy-driven anomalous flow. By
123 doing so, the baroclinic response to diabatic heating evolves to an eddy-driven response with a
124 barotropic structure and finally to a net flow response (initial heating response plus eddy feedback)
125 of the North Pacific storm track. For a background state inspired by the January climatology,
126 they show that the eddy forcing acts to shift the heating-induced anomalous ridge downstream and
127 toward the northeast.

128 More generally, several mechanisms can affect the mean jet position and intensity in the presence
129 of diabatic heating acting on different time and spatial scales:

130 First, a large fraction of the diabatic heating on the scale of individual eddies is generated along
131 ascending airstreams in the warm sector of extratropical cyclones, known as warm conveyor belts
132 (Harrold 1973; Browning et al. 1973). As shown in Sheldon et al. (2017), the warm conveyor
133 belt activity at the entrance of the Atlantic storm tracks is anchored above the warm tongue of
134 the Gulf Stream and could explain the co-location between time-mean ascents, precipitation and
135 the warm tongue (Minobe et al. 2008). Warm conveyor belts can have a local influence on the
136 jet. Diabatic heating produces a positive potential vorticity (PV) anomaly at low levels and a
137 negative PV anomaly near the tropopause (Stoelinga 1996; Pomroy and Thorpe 2000), which is
138 schematically represented in Fig. 4a from Wernli and Davies (1997). Such PV anomaly can locally
139 intensify the jet (Grams et al. 2011; Schemm et al. 2013; Weijenborg and Spengler 2020; Rivière
140 et al. 2021; Wimmer et al. 2022).

141 Second, the North Atlantic jet stream is driven by eddy momentum flux convergence associ-
142 ated with the propagation of large-scale Rossby waves that originate from regions of enhanced
143 baroclinicity (Hoskins et al. 1983; Vallis 2017), suggesting a strengthening of the eddy-driven jet
144 if baroclinicity increases. In this regard, the presence of oceanic frontal areas, such as the Gulf
145 Stream, has been shown to be crucial to maintain the baroclinicity through sensible heat fluxes
146 (Sampe et al. 2010; Hotta and Nakamura 2011). Further, it has been established that not only is
147 an increase in baroclinicity important, but its the exact location relative to the mean jet position

¹The authors note that the time-mean flow was taken from the eastern North Pacific, which during winter is dominated by the subtropical jet stream, while the model appears to be too sensitive to the forcing when using the time mean conditions from the North Atlantic, which has more the nature of an eddy-driven jet.

148 matters (Rivière 2009). An increase of baroclinic eddy activity can thus lead to an acceleration or
149 shift of the jet but also entails an enhanced diabatic heating associated with the growing baroclinic
150 eddies that maintains the storm track (Hoskins and Valdes 1990a).

151 Quantifying the relevance of each of these mechanisms and their effects on the jet is therefore
152 necessary to understand changes in the North Atlantic jet stream. We aim at investigating the role of
153 changes diabatic heating in recent changes of the jet stream through the analysis of ERA5 data for the
154 winter season following the above outlined two mechanisms. To this end, next to ERA5, a hierarchy
155 of idealized and fully-coupled climate simulations is used to better understand the relevance of
156 diabatic heating and mean baroclinicity on the trends in the winter North Atlantic circulation.
157 This analysis also allows us to pinpoint to potential sources of uncertainty in climate projections,
158 and provide some guidance on relevant aspects required to adequately simulate jet stream changes
159 that most models do not adequately capture (Blackport and Fyfe 2022). Admittedly, additional
160 processes, such as stratospheric (Kidston et al. 2015) or tropical influences (Yu and Lin 2016),
161 may play a role in jet trends, but this study mainly focuses on the impact of changes in diabatic
162 heating. It should also be noted that this study does not attempt to attribute trends to either natural
163 variability or anthropogenic climate change.

164 Specifically, this study addresses the following research questions:

- 165 • What is the local effect on the jet of an increase in transient diabatic heating rate pulses
166 over the storm track entrance region resulting from enhanced diabatic heating on the scale of
167 individual cyclones (as, for example in Fig. 4a, in Wernli and Davies 1997)?
- 168 • What is the downstream impact of a change in mean baroclinicity, potentially resulting from
169 enhanced diabatic heating over the Gulf Stream sector, on the downstream evolution of the
170 North Atlantic jet via eddy momentum flux convergence (as, for example, in Hoskins et al.
171 1983)?
- 172 • Which of these processes are adequately represented in fully-coupled climate simulations and
173 what are the factors that contribute to uncertainty in climate projections of the North Atlantic
174 jet stream?

175 The remaining of the paper is structured as follows. Section 2 presents the winter North
176 Atlantic jet trends in the reanalysis and illuminates the main mechanisms driving the trends,

177 Section 3 discusses the local effect of diabatic heating over the Gulf Stream through an idealized
178 frontal-geostrophic model experiment, Section 4 explores the role of feedback of eddy horizontal
179 momentum fluxes onto the mean flow through aquaplanet simulations and Section 5 analyzes
180 jet trends in fully-coupled climate simulations. Main conclusions are provided in section 6. To
181 improve readability of the individual sections, the main data and methods are introduced in the
182 corresponding sections.

183 **2. Observed jet stream trends over the North Atlantic**

184 The ERA5 reanalysis (Hersbach et al. 2020) is used to evaluate recent trends in the winter
185 (December–February, DJF) North Atlantic jet stream. The dataset consists of global data with a
186 spatial resolution of 0.28° and 137 vertical levels up to 1hPa, spanning from 1979 until present with
187 hourly temporal resolution. For the analysis presented in this study horizontal wind components
188 and temperature were interpolated into an horizontal resolution of 0.5° and 11 pressure levels
189 between 900 and 100 hPa². Linear trends are computed from daily averages in the period 1979-
190 2022. This data frequency is considered to be representative enough for the subsequent trend
191 analysis. Although ERA5 reanalysis data are available from 1940, the period is restricted to the
192 satellite era. This is because the main region of interest is the North Atlantic and thus the reanalysis
193 may not be well constrained over the ocean in the earlier period. However, zonal wind speed trends
194 for the period before the satellite era are provided in Appendix A.

195 *a. Methods to analyze trends*

196 1) BAROCLINICITY: ISENTROPIC SLOPE

197 To study the impact of changes in baroclinicity in the modification of the jet stream, we use
198 the slope of the isentropic surfaces to quantify the growth potential of baroclinic waves, such as
199 extratropical cyclones (Van Delden 1999; Thompson and Birner 2012; Igel and van der Heever
200 2014; Papritz and Spengler 2015). A necessary condition for baroclinic growth of a disturbance is
201 that on average the motion of air parcels has a slope lower than the slope of the isentropic surfaces

²The following pressure levels in hPa are used in this study: 100, 200, 250, 300, 400, 500, 600, 700, 800, 850 and 900.

202 (Green 1960). The slope is computed as

$$S = \frac{\frac{\partial \theta}{\partial y}}{\frac{\partial \theta}{\partial p}}, \quad (1)$$

203 where θ is the potential temperature, y represents the meridional direction and p is pressure. The
204 slope is computed from hourly data averaged to obtain daily means, which are used to compute
205 linear trends.

206 2) EDDY MOMENTUM FLUX CONVERGENCE: E-VECTOR

207 Eddy momentum convergence is explored by means of the divergence of the E vector (Hoskins
208 et al. 1983), whose direction indicates eddy propagation, which is opposite from the eddy momen-
209 tum transfer. The three-dimensional E vector is defined as

$$\mathbf{E} = \left(0.5 \overline{(v'^2 - u'^2)}, -\overline{u'v'}, \frac{f}{\partial \theta / \partial p} \overline{v'\theta'} \right), \quad (2)$$

210 where u and v are the zonal and meridional wind components, respectively, f is the Coriolis
211 parameter and p is pressure. The overbars indicate time means and the primes denote anomalies,
212 which are computed by subtracting the 10-day high pass filtered component from the total fields.
213 The vertical component of the E vector is thus proportional to the heat flux.

214 The orientation of the E vector is related to the eddy shape and the orientation of Rossby
215 wave breaking (RWB) (Orlanski 1998). Equatorward E vectors indicate anticyclonic RWB, while
216 cyclonic RWB is associated with poleward E vectors (Drouard et al. 2015). Additionally, the
217 zonal component of the E vector provides an indication of the eddy shape, so meridional elongated
218 eddies are represented by eastward E vectors and zonally elongated eddies by westward E vectors.
219 6-hourly data are used to compute the E vector, which is subsequently daily averaged.

220 3) STATISTICAL SIGNIFICANCE

221 We apply the false-discovery rate (Wilks 2016) to compute p-values for the trends computed
222 throughout the study. With this approach, a global p-value, p^* defined by: $p^* = \max[p_i \leq (i/N)\alpha]$
223 is computed, where i is the index identifying the sorted p-values, N is the total number of grid

224 points and α is a threshold that controls the fraction of null hypotheses erroneously rejected, which
225 is fixed to 0.1 in this study. For each grid point, local p-values are compared to the global p-value.

226 *b. Observed jet stream trends over the North Atlantic*

227 Analysis of the wintertime (December-January-February) zonal wind speed trend in the ERA5
228 dataset from 1979 until 2022 shows a strengthening and mild poleward shift of the North Atlantic
229 jet stream between 40°N and 50°N (Fig. 1a) over the entrance of the storm track on the US East
230 Coast and the Gulf Stream sector (Fig. 1b). This is not the case further downstream over the eastern
231 North Atlantic, where the trend is characterized by a slight equatorward shift and a southwest-to-
232 northeast extension towards the UK and western Europe. There is also a decrease in zonal wind
233 speed between 20–30°N, extending towards the Iberian Peninsula and the western Mediterranean
234 (Fig. 1b), and an additional decrease north of the jet between 65–80°N over the Nordic Seas.

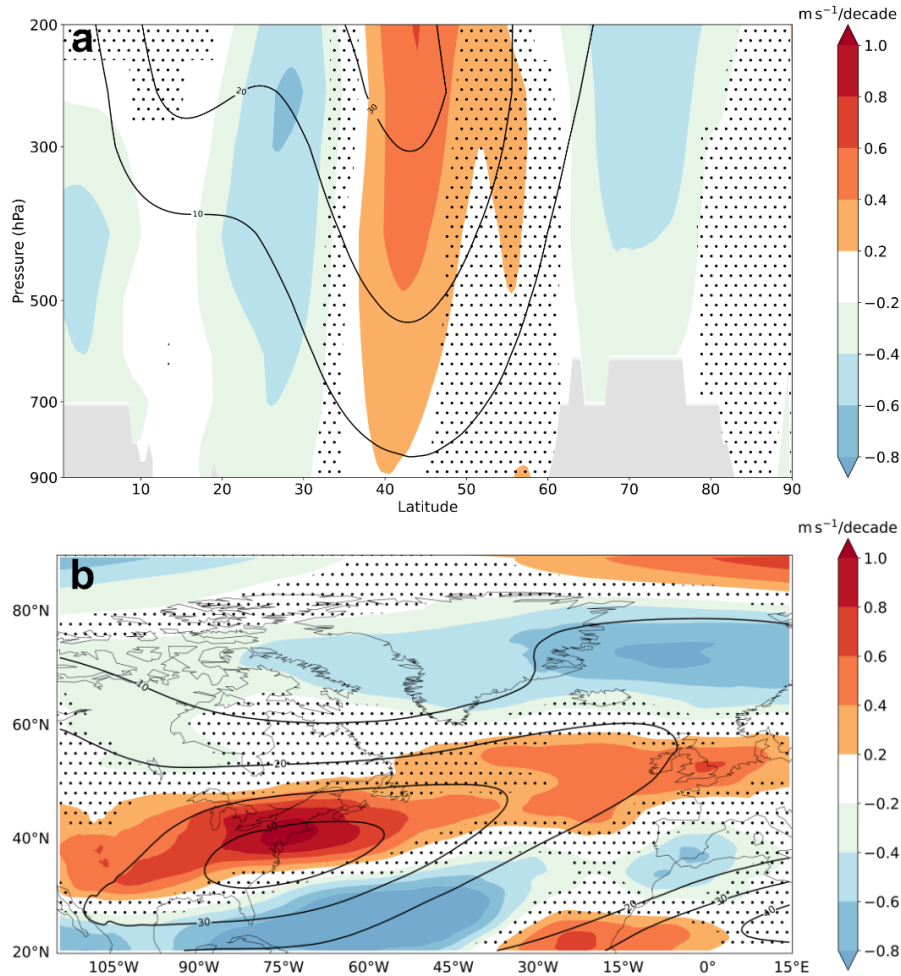
235 Upstream over eastern North America, the trend in the 500-hPa geopotential anomalies shows an
236 intensification of the climatological trough over North America (dashed green contours in Fig. 1b).
237 At lower levels, the negative anomaly extends to the east, indicating the presence of a westward
238 tilted trough with height. The implications of this pattern are discussed in the next subsection.

243 1) TRENDS IN POTENTIAL TEMPERATURE, DIABATIC HEATING AND TEMPERATURE ADVECTION

244 We begin the detailed analysis of the jet trends by investigating the mechanisms that affect it
245 over the Gulf Stream before consideration is given to the downstream sector over the eastern North
246 Atlantic. We first perform a trend analysis of the potential temperature and changes in its meridional
247 structure related to diabatic heating and advection. Figure 2 presents trends in potential temperature,
248 diabatic heating³ and three-dimensional potential temperature advection tendency. The general
249 trend in potential temperature depicts enhanced potential temperature between 30–40°N, slightly
250 equatorward of the region of increased zonal wind speed and 80–90°N (red shading in Fig. 2a,b).
251 The increase located near the pole is seemingly associated with the Arctic Amplification (AA). It
252 reaches into the stratosphere (Fig. 2a).

253 At midlatitudes between 30-50°N potential temperature increases in the troposphere but decrease
254 in the stratosphere, leading to a decrease in static stability (Fig. 2a) in the layer between 350–200 hPa,

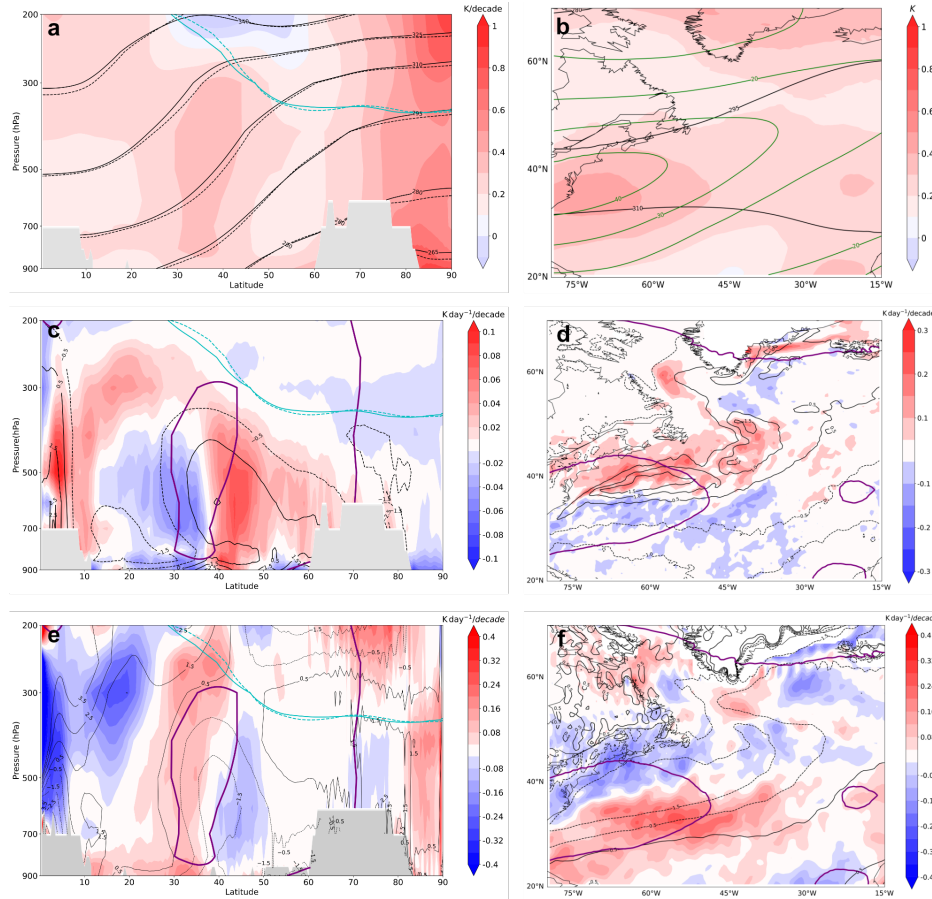
³Diabatic heating refers to the sum of the potential temperature tendencies due to radiation, turbulence, convection, the microphysics of clouds and the drag of gravity waves from the physical model underlying the reanalysis data.



239 **FIG. 1. Wintertime zonal wind speed trend over the North Atlantic in ERA5:** a) Mean of the trend
 240 (shading) and climatological average in zonal wind speed (black contours) over the North Atlantic storm track
 241 region (80°W – 15°W) for DJF in the period 1979-2022. b) Zonal wind trend (shading) and climatological mean
 242 (black contours) at 250 hPa. The stippling represents areas with p-values higher than p^* (see text for details).

255 exactly where the tropopause height increases (light blue contours in Fig. 2a indicate the dynamical
 256 tropopause during the first and last decades).

271 To better understand the trend in potential temperature, it is illustrative to analyze changes
 272 in diabatic heating, which are described above, and advection, as both processes locally affect
 273 potential temperature. Although attributing trends to either anthropogenic climate change or natural
 274 variability is beyond the scope of this study, it is appropriate to analyze trends in diabatic heating
 275 and advection in light of changes in a warmer climate and how these may affect baroclinicity. Moist

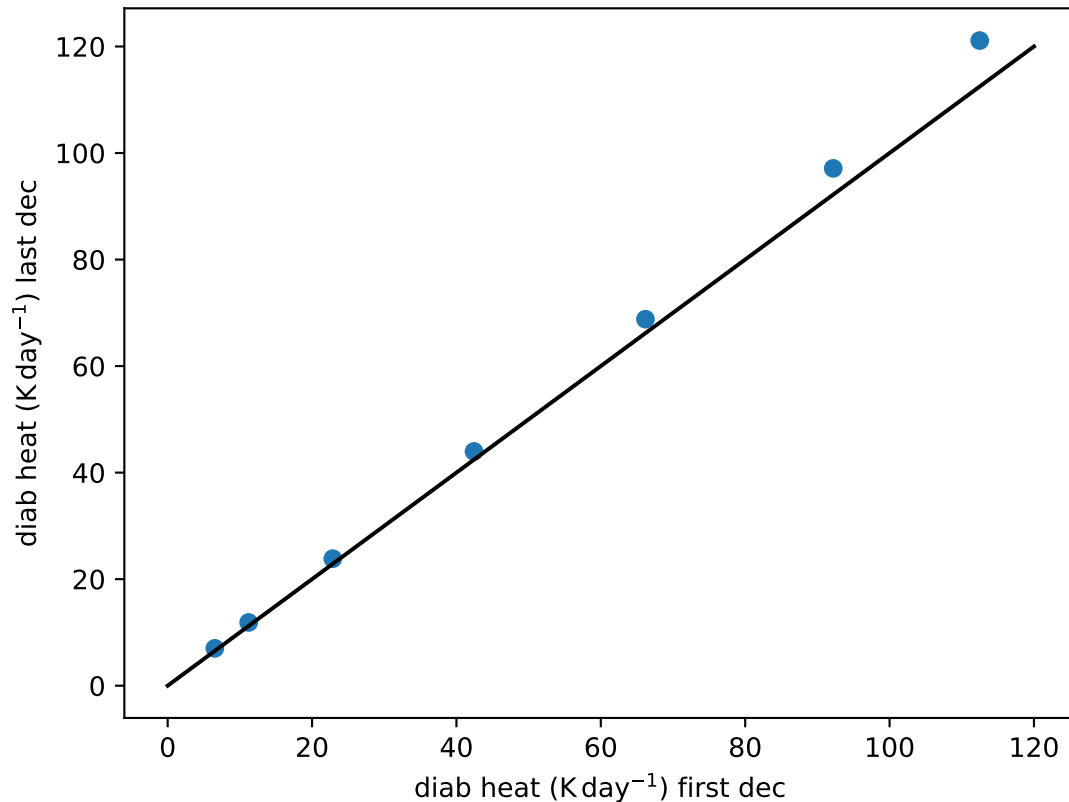


257 **FIG. 2. DJF Potential temperature, diabatic heating and advection trends over the North Atlantic:**
 258 a) Zonal mean of the potential temperature trend between 80° – 15° W (shading), zonally averaged potential
 259 temperature mean in the same sector for the first (solid black lines) and the last decade (dashed black lines) in the
 260 selected period for ERA5 (1979-2022). b) Trends in potential temperature (shading) and climatological mean
 261 (black contours) averaged between 300 and 850 hPa. Green contours represent zonal wind speed climatology in
 262 the ERA5 period at 250 hPa. c) Zonal mean of the trend in potential temperature tendency due to parameterizations
 263 (diabatic heating) between 80° – 15° W (shading). Black contours depict the climatological mean. d) Trends in
 264 diabatic heating (shading) and climatological mean (black contours) averaged between 300 and 850 hPa. e)
 265 Zonal mean of the trend in potential temperature tendency due to three-dimensional advection in the same North
 266 Atlantic sector. Black contours show the climatological mean over the same region. f) Trends in potential
 267 temperature tendency due to advection over the North Atlantic averaged between 300 and 850 hPa (shading) and
 268 climatological mean (black contours). In panels a), c) and e) solid and dashed light blue contours depict the
 269 2-PVU potential vorticity contour for the first and last decades in the ERA5 data, respectively. In panels c)-f) the
 270 contour corresponding to a potential temperature trend of 0.3 K/decade is represented by purple lines.

276 diabatic processes are expected to increase in a warmer climate, as the saturation vapor pressure of
277 the atmosphere increases with rising temperatures – as a result of the Clausius-Clapeyron relation
278 (Held and Soden 2006) –, thus increasing potential temperature locally, while a reduction in the
279 land-sea contrast has the potential to promote a reduction in cold-air advection into the region of
280 interest from upstream (Wallace and Joshi 2018), thus also increasing the potential temperature
281 over the Gulf Stream area.

282 In the Gulf Stream sector, ocean fronts have been shown to intensify the storm track (Small et al.
283 2014). In addition, this is an area of enhanced upward motion and cloud formation (Minobe et al.
284 2008; Kuwano-Yoshida et al. 2010), it is thus expected that diabatic heating plays climatologically
285 an important role. Indeed, the climatological mean diabatic heating (black contours in Fig. 2c,d)
286 is co-located with the mean position of the Gulf Stream. The trend pattern displays a dipolar
287 structure in a meridional band between 20–60°N (shading in Fig. 2c,d) with an increase near its
288 climatological mean position (black solid contour in Fig. 2d) and a slight poleward shift plus a
289 reduction south of 40°N, where it is climatologically negative (dashed black contour in Figs. 2c,d).
290 The trend pattern suggests a mild poleward shift paired with an intensification of diabatic heat
291 release close to its climatological maximum value.

292 Ambaum and Novak (2014) and Novak et al. (2015) show that the time series of diabatic heating
293 over the Gulf Stream region is characterized by spikes that occur on synoptic time scales and
294 are indicative of the baroclinic development growth of extratropical cyclones. These results and
295 the linear upward trend in diabatic heating suggest that the diabatic heat release for each, or at
296 least some of these events, has increased over recent decades. The trend in the jet over the Gulf
297 Stream likely reflects the cumulative influence of many of these diabatic heating peaks of increasing
298 intensity, the net effect of which is a local strengthening and regionally limited poleward shift of
299 the jet. To quantify this increase in strong peaks of diabatic heating, Fig. 3 compares the values of
300 the percentiles in the distribution of diabatic heating in an area near the Gulf Stream for the first
301 and last decades of the ERA5 period. In the last decade, the values of the highest percentiles are
302 larger than in the first decades, which supports the argument of increased intensity in the strongest
303 diabatic heating pulses. To illuminate this diabatic heating-jet stream interactions, an idealized
304 two-dimensional frontal-geostrophic simulation is used in Section 3 to estimate the effect a single



307 **FIG. 3. Quantile-quantile plot for the diabatic heating:** Values of diabatic heating percentiles in the area
 308 between 80–30°W, 35–45°N and 300–850 hPa for the first decade (horizontal axis) and last decade (vertical axis)
 309 in the considered ERA5 period (1979-2022). The lowest value corresponds to percentile 90 and the highest to
 310 percentile 99.9999. The black line represents the reference corresponding to equal values of the percentiles for
 311 both decades.

305 diabatic heating event has on the position and strength of an idealized jet stream and whether it
 306 produces an impact comparable to the observed jet feedback over the Gulf Stream sector.

312 Fig. 2e,f display advection trends in ERA5 data. In this case, there is a positive trend between
 313 20–35°N with a low vertical tilt over the US East Coast – Gulf Stream region (red shading in
 314 Fig. 2e,f). This positive trend in the potential temperature advection is dominated by an upward
 315 trend in zonal advection. Because advection north of 25°N is climatologically negative (dashed
 316 black contours in Fig. 2b,d) the upward trend indicates a reduction in cold-air advection. This
 317 effect may be related to a reduction in the winter land-sea temperature contrast, as warming is more
 318 pronounced over North America than over the adjacent ocean (see Fig. 3a in Simmons (2022)).

319 Although the positive advection trend is also strong at upper levels, the positive signal extends
320 towards the surface supporting the role of reduced land-sea contrast as a driver of this trend. The
321 maximum in potential temperature trend between 30–35°N is well aligned with the reduction in
322 cold advection. This suggests that advection is the main contributor to the potential temperature
323 trend in this region. Between 35 and 40 °N, the maximum in potential temperature trend coincides
324 with the positive trend in diabatic heating, especially at mid and upper levels (Fig. 2c).

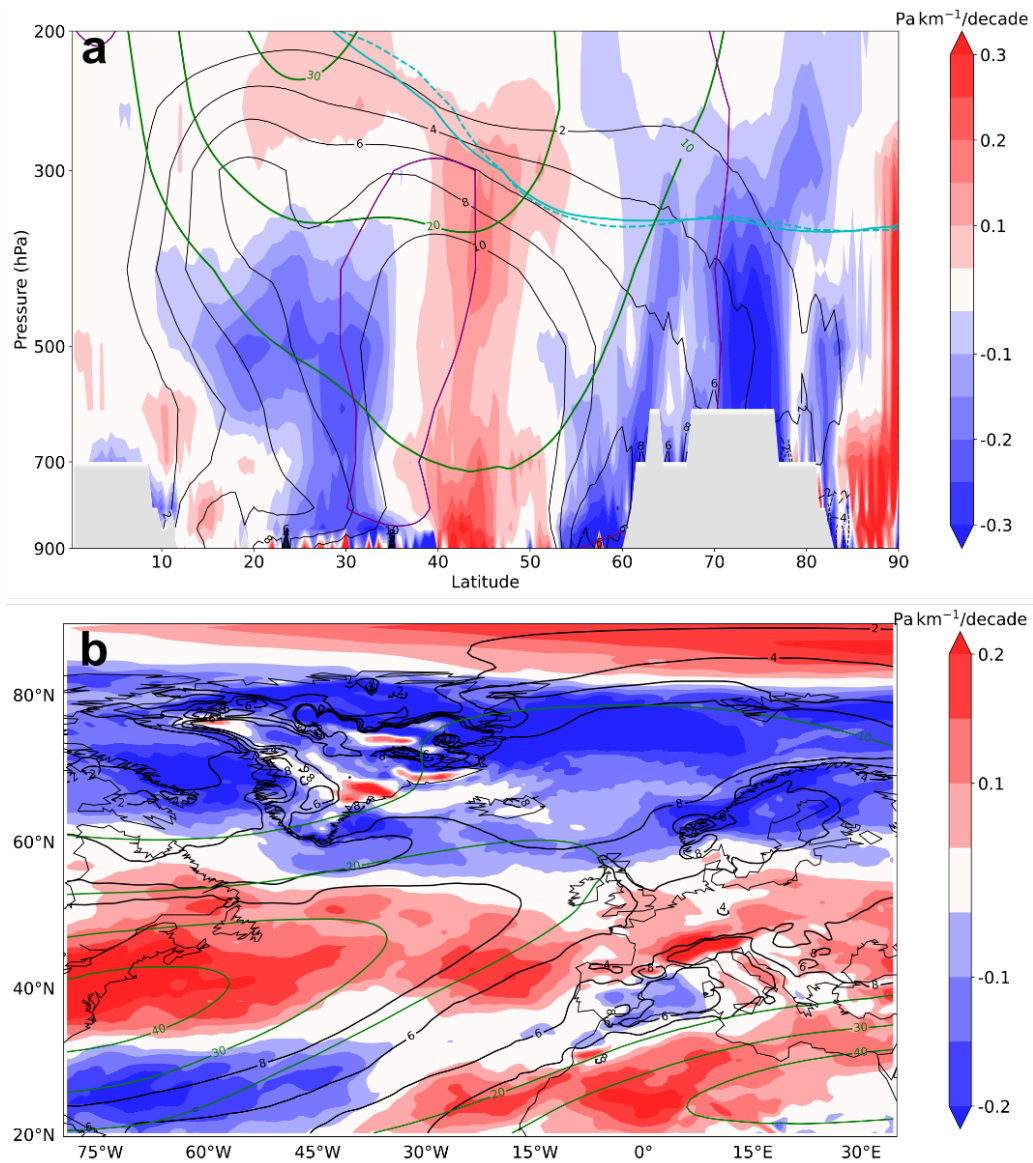
325 2) TRENDS IN MEAN BAROCLINICITY

326 Next, we analyze changes in baroclinicity, especially over the downstream sector. By thermal
327 wind balance, the baroclinic component of the jet increases where the horizontal temperature
328 gradient is stronger. Fig. 4 reveals a tripolar trend pattern in the isentropic slope, which is used as
329 a proxy for the mean baroclinicity, over the North Atlantic storm track (see subsection 1). There is
330 a positive trend in midlatitudes, between 40–50°N (red shading in Fig. 4a) flanked by two bands of
331 negative trends. The upper-level maximum of the increase is located at 300 hPa around the main
332 jet axis, close to where an increase in the zonal wind speed (Fig. 1a) is identified. The combination
333 of a reduction in upper-level static stability, coupled with an increase in the meridional temperature
334 gradient, leads to this increase in mean baroclinicity in the 40–50°N band, which appears to be
335 critical for the evolution of the North Atlantic jet in this sector. This is accompanied by a lifting
336 of the dynamical tropopause near 40°N (light blue solid and dashed contours in Fig. 4a). In the
337 middle and lower troposphere, the slope trend shows a decrease equatorward of the jet axis at 30°N,
338 and an increase around the mean jet axis and again a decrease in a latitude band between 50 °N
339 and 80°N (Fig. 4a). Trends in slope, potential temperature and diabatic heating are consistent. The
340 region of strongest increase in potential temperature is related to a combination of reduced cold-air
341 advection in the equatorward flank of this area and a positive trend trend in diabatic heating, which
342 is located polewards of the increase in advection. In a baroclinic atmosphere a diabatic heating
343 maximum tends to lower the height of a tilted isentropic surface on its equatorward side and a
344 lowered isentrope corresponds to an increase in potential temperature (Van Delden 1999; Papritz
345 and Spengler 2015). This pattern in combination with the reduced cold air advection results in the
346 increase in potential temperature (Fig. 2a,c), which is found slightly equatorward of the increase in
347 the slope (Fig. 4) and diabatic heating release (Fig. 2c,d). At high latitudes, the warming pattern

348 (Fig. 2a,c) causes the isentropic surfaces to bend downward causing a lowered slope at equatorward
349 latitudes and an increased slope at even higher latitudes, which is in agreement with the dipole
350 trend pattern of the slope between 70–90°N (Fig. 4a). Overall, the slope trend is strongest between
351 40–50°N, which is close to its climatological mean position, and it extends throughout the entire
352 troposphere (Fig. 4a).

360 3) TRENDS IN EDDY MOMENTUM FLUX

361 Finally, we analyze trends in eddy momentum convergence by means of the E vector. The
362 direction of horizontal momentum flux is opposite to the direction of the meridional E vector.
363 Therefore, increased eddy momentum convergence is represented by a larger E vector divergence.
364 The axis of zero E vector divergence is located around 45°N and eddy momentum convergence
365 peaks slightly north of the jet core over eastern North America (vectors and shading in Fig. 5),
366 which is in agreement with the intensification and northward shift of the jet in this sector (Fig.
367 1b) and the area of positive slope trend (Fig. 4b). Downstream over the eastern North Atlantic,
368 momentum convergence is enhanced south of the climatological jet position, where an intensifi-
369 cation, a mild equatorward shift and an extension of the jet are identified. In general, the area of
370 increased momentum convergence coincides well with the region where the jet intensifies (Fig. 1b),
371 consistent with the general understanding that the North Atlantic jet is an eddy-driven jet. The
372 enhanced convergence of the eddy momentum flux near the jet axis and a reduction poleward of
373 the climatological jet flank produces a reduction of the jet width, which agrees with analyses of the
374 projected trends of the North Atlantic jet (Peings et al. 2018). Overall, the momentum trend pattern
375 is remarkably similar to the archetypal steering of Rossby waves by baroclinic eddies, in particular
376 over the western North Atlantic and the storm track entrance region. Momentum convergence is
377 enhanced in the area of steering and waves propagate meridionally away from this region (Vallis
378 2017). In addition, over the western North Atlantic, upper-level changes in the eddy momentum
379 flux convergence are related to a trend towards more poleward-oriented E vectors. This change in
380 E vector is indicative of enhanced cyclonic RWB, which contributes to the equatorward shift of the
381 downstream where waves typically break. The intensification of the climatological trough provides
382 a potential explanation for the asymmetry of the E vector trend. Downstream of a trough anomaly,



353 **FIG. 4. DJF isentropic slope trends in ERA5 over the North Atlantic:** a) Zonal mean of the trend in the slope
 354 of the isentropic surfaces between $80^{\circ} - 15^{\circ}\text{W}$ (shading), climatological mean slope (black contours, in Pa km^{-1})
 355 and zonal mean zonal wind (green contours) in ERA5 (1979-2022). Blue contours represent the tropopause
 356 height (2 PVU contour) for the first (solid) and last decade (dashed) in the considered period. For reference, the
 357 contour corresponding to a potential temperature trend of $0.3 \text{ K}/\text{decade}$ is represented by purple lines. b) Trend
 358 of the slope of the isentropic surfaces averaged between 250 and 850 hPa (shading), climatological mean over
 359 the same vertical layer (black contours) and climatological mean zonal wind speed at 250 hPa (green contours).

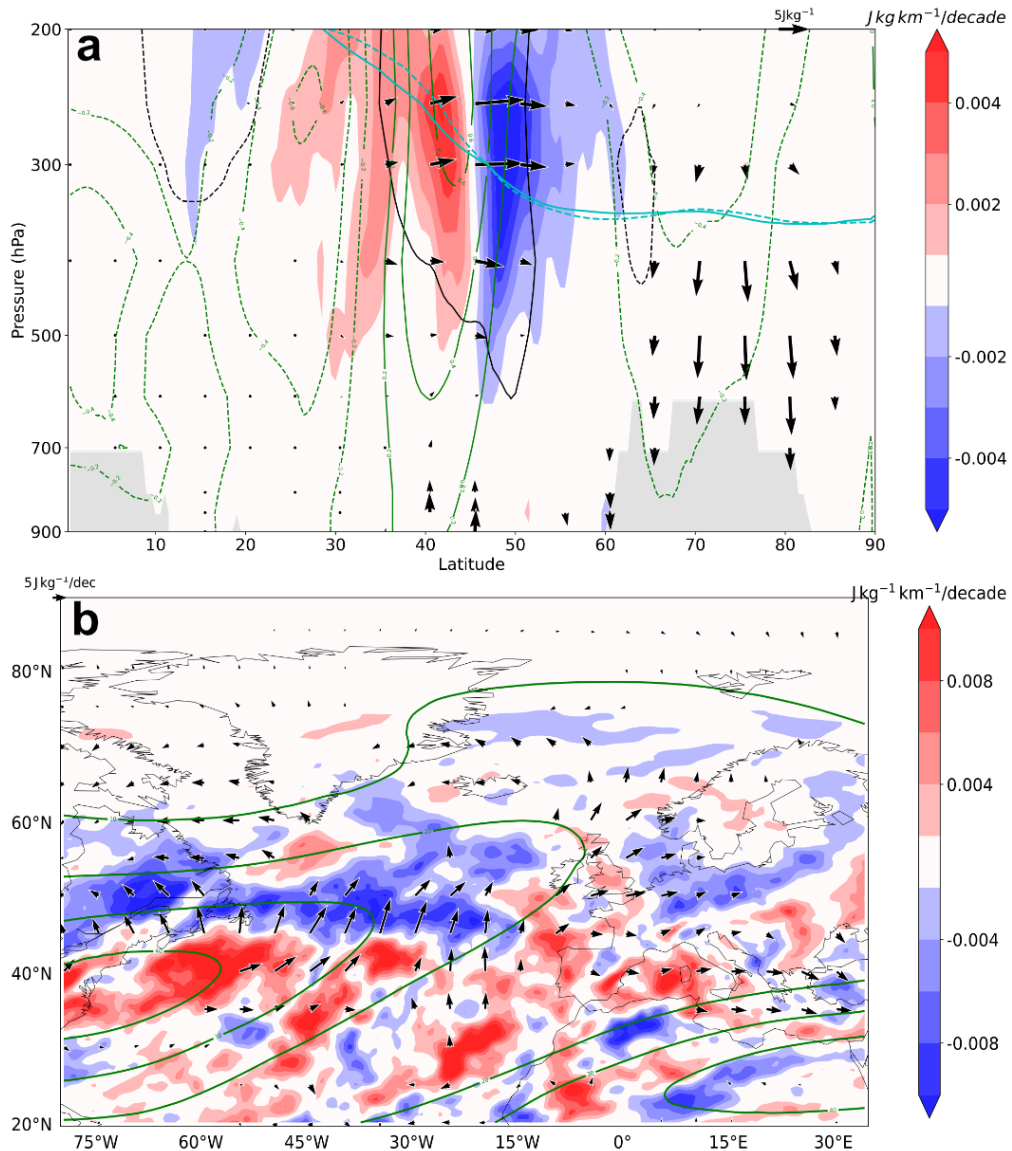
383 upper-level transient eddies tend to propagate with a cyclonic orientation, which is associated with
384 poleward pointing E vectors (Drouard et al. 2013).

394 At the same time, the vertical component of the E vector, which is proportional to the meridional
395 eddy heat flux, displays an upward trend between 40–50°N below 700 hPa (Fig. 5a). The upward
396 E vector in the lower troposphere indicates an increase in the initial growth of eddy activity at
397 the entrance region of the storm track. The anomalous E vector diverges in the upper troposphere
398 from the same latitudes where eddy activity increases, consistently with well-known storm-track
399 properties (see Fig. 13 in Hoskins et al. (1983)). Additionally, eddy heat flux is also reduced
400 below 400 hPa at high latitudes. This is associated with a decrease of the meridional near-surface
401 temperature gradient in this sector due to the Arctic Amplification. This also suggests that the
402 influence of the AA on the activity of the transient waves is limited to high latitudes north of
403 approximately 70°N.

404 The zonal component of the E vector is also reduced over the western North Atlantic between
405 75–60°W, 45–50°N (Fig. 5b). The zonal E vector component tends to cause a southwest–northeast
406 tilting of the storm track (see Fig. 3 in Orlanski (1998)), as a result of the quadruple vorticity
407 pattern associated with the zonal E vector component. Therefore, this reduction implies a reduced
408 northward deflection of the jet (Orlanski 1998) and thus, a more zonal orientation of the storm
409 track. The reduction in the zonal component of the E vector could explain the zonal extension over
410 the eastern North Atlantic and indicates changes in eddy shape, that is, a tendency towards less
411 meridionally and more zonally elongated eddies (Orlanski 1998).

412 In summary, over the western North Atlantic and the storm track entrance region, where diabatic
413 heating intensifies, the jet strengthens and shifts slightly poleward, while over the eastern North
414 Atlantic it intensifies, extends downstream, shifts slightly equatorward and tends to become more
415 zonal. Established theoretical considerations based on a linear wave model suggest that an increase
416 in diabatic heating is inherently connected to the formation of a westward tilted trough, which at
417 lower levels is located downstream of the maximum in heating (Hoskins and Karoly 1981). This
418 effect is not reproduced in the reanalysis (not shown) and this discrepancy could be related to the
419 role of nonlinearities in the response.

420 An increase in the mean baroclinicity over the Gulf Stream sector affects the entire depth of the
421 troposphere and an increase of the convergence of eddy momentum fluxes in a zonal band across the



385 **FIG. 5. DJF E vector trends in ERA5 over the North Atlantic:** a) Zonal mean of E vector divergence trend
 386 over the North Atlantic sector, between 80°–15°W (shading), climatological mean over the ERA5 period in DJF
 387 (black contours, $\pm 10^{-5} \text{ J kg}^{-1} \text{ m}^{-1}$) and trend in zonal wind speed (green contours). Black vectors represent E
 388 vector trend (per decade). The size of the vertical component is four times the size of the horizontal component.
 389 Blue contours represent the tropopause height (2 PVU contour) for the first (solid) and last decade (dashed) in
 390 the considered period. b) E vector trend (black vectors) and divergence trend (shading) in ERA5 (1979–2022)
 391 at 250 hPa over the North Atlantic. Green contours represent the climatological mean of the zonal wind at the
 392 same pressure level. In both panels, vectors are shown only if the p-value in at least one of the components is
 393 lower than the global p^* .

422 entire North Atlantic is suggestive of an acceleration of the jet through the excitation of baroclinic
423 waves. In addition, a northward trend in the E vector suggests an increase in cyclonic wave
424 breaking, which contributes to the equatorward shift of the jet downstream. In the next sections,
425 the local influence of diabatic heating over the Gulf Stream as well as the downstream mechanism
426 related to the eddy-mean flow interaction are further explored through idealized simulations.

427 **3. Local diabatic influence on the jet stream**

428 *a. Two-dimensional frontal-geostrophic simulation*

429 At the jet entrance above the east coast of North America and the Gulf Stream sector, the jet
430 latitude is less variable compared to the central and eastern North Atlantic (Brayshaw et al. 2011;
431 Small et al. 2014; O'Reilly et al. 2017). In this sector, rapid growth of baroclinic waves is occurring
432 due to high baroclinicity. Additionally, diabatic heating in the lower to mid troposphere, which
433 occurs episodically within baroclinic wave development (Ambaum and Novak 2014), is known to
434 have a local impact on the jet stream through the modification of PV (Stoelinga 1996; Pomroy
435 and Thorpe 2000) in an environment characterized by strong vertical wind shear beneath the jet
436 stream core. The purpose of this section is to quantify this direct and transient effect of a repeated
437 increase in diabatic heating during baroclinic wave development by using a two-dimensional
438 frontal-geostrophic model (Harvey et al. 2020) including an idealized diabatic heating perturbation
439 inspired by the trends obtained in ERA5 (Figs. 6 and 2c). It is expected a PV reduction above the
440 heating and an increase below, which are associated with circulation changes that locally affect the
441 jet (Wernli and Davies 1997).

442 The two-dimensional frontal-geostrophic model is based on the elliptic Sawyer-Eliassen equation,
443 including diabatic heating, under the Boussinesq approximation (Harvey et al. 2020). The model
444 is initialized with zonal wind and potential temperature profiles similar to the ERA5 climatology
445 in the Gulf Stream sector (Fig. 6) and a Gaussian diabatic heating perturbation, which remains
446 constant during the whole simulation period and has a maximum of approximately 0.4 Kday^{-1} ,
447 which matches the maximum in the long-term trend in ERA5. This perturbation is designed to
448 mimic the average rate of increased local diabatic heating in the Gulf Stream sector, as identified
449 in the long-term trend in ERA5. This experiment aims to establish whether the magnitude of the
450 increase in zonal wind in this area detected in the reanalysis trend is consistent with the effect of

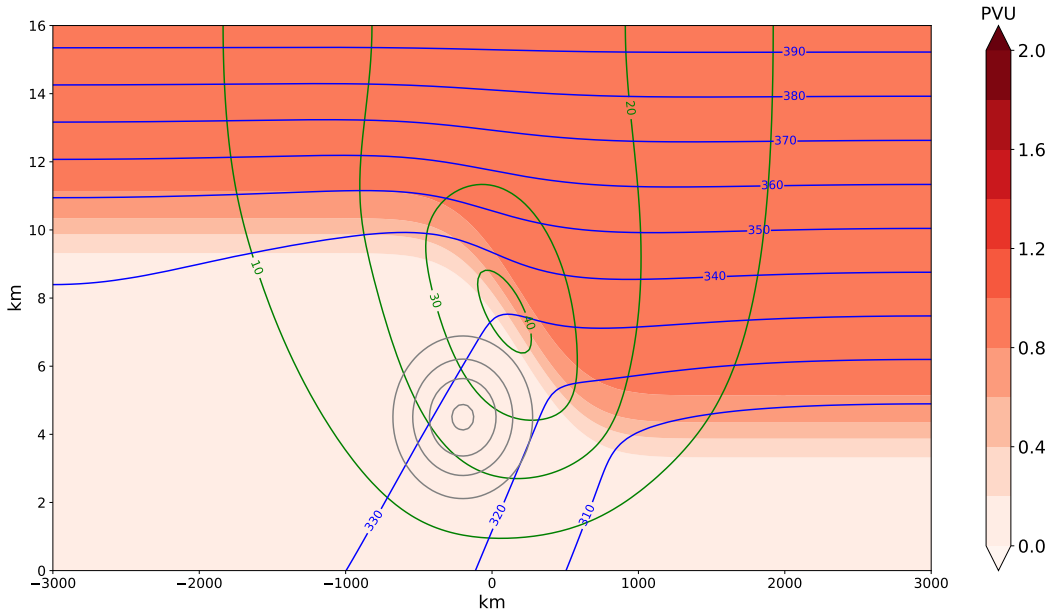
451 a succession of intensified diabatic heating pulses during the development of individual cyclones.
452 For heights between approximately 2 and 7 km the heating is greater than 0.1 K day^{-1} (Fig. 6).
453 Additional experiments with an extended region of heating reaching lower levels as obtained for
454 the trends in ERA5 (Fig. 2c) produce qualitatively similar results.

455 In more detail, the model is integrated forward in time while PV is positive over the full domain,
456 which is a necessary condition for the Sawyer-Eliassen equation to be elliptic. It typically takes
457 four to five days, given the magnitude of the diabatic heating rate imposed, before the forcing drives
458 the PV negative above the heating maximum. This is longer than typical synoptic timescales and
459 so we would expect in practice that the heating in each system would stop earlier as a baroclinic
460 wave life cycle follows its course. The model evolves by advecting the full Ertel PV field, which
461 under the two-dimensional Boussinesq assumptions reduces to $P(y, z) = (m_z \theta_y - m_y \theta_z) / \bar{\rho}$ where
462 zonal angular momentum $m = -fy + u$ and $\bar{\rho}$ is the background density, in the meridional (y, z)
463 plane. The (v, w) wind are obtained from the Sawyer-Eliassen equation assuming no normal flow
464 at the domain boundaries. The resulting (u, θ) fields are obtained by inverting P using a two-
465 dimensional Monge-Ampère solver, together with boundary conditions $u = 0$ on the meridional
466 boundaries ($\pm 3000 \text{ km}$), $\theta' = 0$ on a rigid lid (16 km) and a time-varying lower-boundary condition
467 $\lambda(y) = -\theta m_y$. This form of the lower boundary condition is analogous to advecting boundary
468 θ in the quasi-geostrophic equations, whilst maintaining the compatibility condition required for
469 solution of the two-dimensional Monge-Ampère equation.

474 This experiment has some similarities with the first step in the study of Peng and Whitaker
475 (1999), where a linear three-dimensional baroclinic model is used to investigate the response to
476 an initial anomalous heating (before adjustment and eddy feedback). However, the approach used
477 here is based on a two-dimensional nonlinear model and our interpretation is that it is the local
478 response of the jet structure to transient heating beneath the level of the jet maximum over the Gulf
479 Stream. Our 2D model cannot develop troughs and ridges, a mechanism which we explore with a
480 non-hydrostatic global atmospheric model in an aquaplanet configuration in another section.

481 *b. Experiment results*

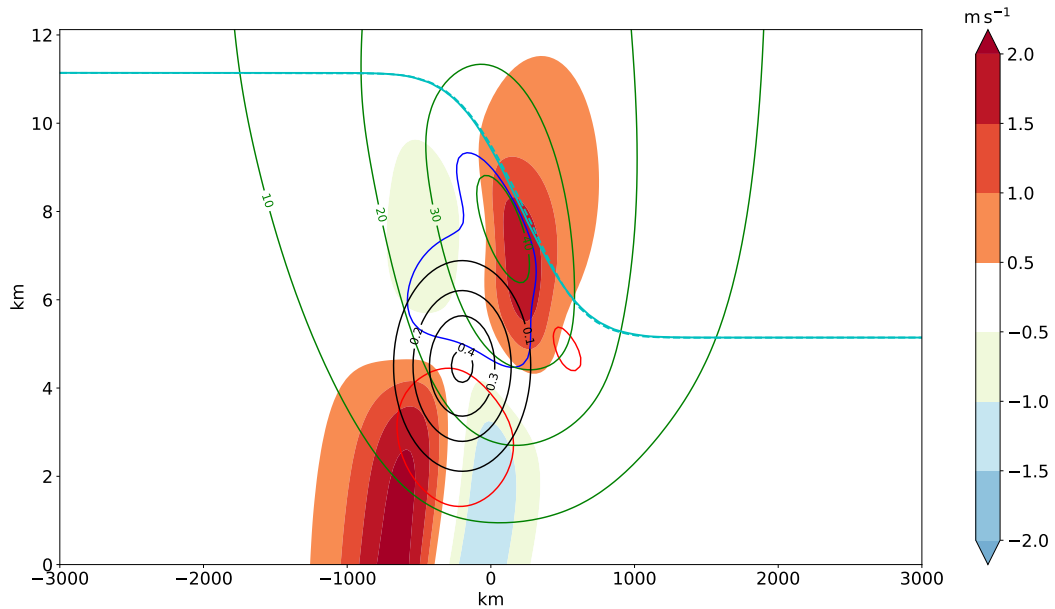
482 The results of the simulation show an intensification of the zonal wind, whose maximum exceeds
483 2 ms^{-1} and a slight poleward shift of the jet above the area of increased diabatic heating (Fig. 7).



470 **FIG. 6. Two-dimensional frontal-geostrophic simulation setup:** Initial conditions for the two-dimensional
 471 semi-geostrophic simulation for PV (shading), zonal wind speed (green contours), potential temperature (blue
 472 contours) and diabatic heating anomaly (gray contours, from 0.1 to 0.4 K day⁻¹). The horizontal axis represents
 473 the meridional coordinate.

484 This is expected from the formation of a negative PV anomaly (blue contour in Fig. 7) leading
 485 to anticyclonic circulation above the region of maximum heat release on the flank of the jet core
 486 and poleward advection of the tropopause by the secondary circulation induced by the heating.
 487 In contrast to reanalysis trends, which extent throughout the troposphere, the increase in zonal
 488 wind speed is confined to the area above the heating where the negative PV anomaly is formed.
 489 Assuming linearity in the response and that the jet trend is composed of an average of the responses
 490 to multiple heating events, with increasing intensity on average as climate warms, the results of the
 491 idealized frontal-geostrophic experiment suggest an intensification of the jet and slight poleward
 492 shift seen over the Gulf Stream produced by this mechanism, which is in agreement with the
 493 long-term ERA5 trend. It indicates that the mechanism has indeed sufficient amplitude to explain
 494 the trend in the jet entrance region. However, the baroclinic component of the response dominates
 495 in the frontal-geostrophic simulation in contrast to the reanalysis trend.

501 As outlined in the introduction, two potential mechanisms, one direct and one indirect are
 502 proposed to explain the effect of diabatic heating on the jet. The direct mechanism, which has



496 **FIG. 7. Impact of a local diabatic heating source in an idealized two-dimensional frontal-geostrophic**
 497 **experiment on the jet location and speed:** Shading represents the difference in zonal wind speed between final
 498 and initial states. Initial zonal wind speed is represented by green contours, diabatic heating by black lines. Red
 499 (blue) contour represents an increase (decrease) in potential vorticity of 0.025 PVU, while the light blue solid
 500 (dashed) contour indicates 0.8 PVU in the initial (final) state.

503 been analyzed by means of this idealized two-dimensional experiment, produces a modification of
 504 the PV via diabatic heating. This process is likely to be important at the entrance of the North
 505 Atlantic, as the heating is anchored by the land-sea contrast and the Gulf Stream SST anomaly
 506 (Minobe et al. 2008; Woollings et al. 2016). This direct mechanism is more likely to create a
 507 baroclinic structure. If diabatic heating is increased on average over many individual cyclogenesis
 508 events, a change in the mean state and an increase in the time mean baroclinicity over the Gulf
 509 Stream sector can be expected (Hoskins and Valdes 1990b; Parfitt and Czaja 2016), which is also
 510 in agreement with the enhanced baroclinicity at all levels identified in the ERA5 trends. In the
 511 indirect mechanism, diabatic heating modifies baroclinicity and consequently baroclinic growth
 512 and, finally, momentum deposit by eddies. Since baroclinic waves are typically not breaking in
 513 the western North Atlantic, the indirect mechanism is expected to become more relevant for the
 514 jet trend more downstream, despite the increase in eddy momentum convergence is found over
 515 the whole North Atlantic in midlatitudes. The increase in the mean baroclinicity will alter the

516 eddy-zonal flow feedback (Hoskins et al. 1983; Lorenz and Hartmann 2003), as seen by a positive
517 trend in the E vector divergence (Fig. 5). This feedback is not limited to the Gulf Stream sector,
518 but also affects the jet downstream into the North Atlantic, where the trend pattern indicates a push
519 of the jet equatorward because of an increased cyclonic wave breaking tendency, indicated by the
520 poleward E vector trend (Fig. 5b). The indirect mechanism is further explored next with idealized
521 aquaplanet simulations.

522 **4. Eddy-mean flow feedback**

523 The two-dimensional frontal-geostrophic model discussed in the previous section lacks the impor-
524 tant contribution by synoptic-scale baroclinic eddy growth and Rossby wave breaking. Therefore,
525 in order to substantiate the changes in eddy-zonal mean flow feedback resulting from enhanced
526 mean baroclinicity, we perform a set of simulations using a fully-fledged Atmospheric Global
527 Circulation Model (AGCM), but in a semi-realistic configuration with typical climate model res-
528 olution. However, given the reduced complexity of these simulations, the main goal of these
529 experiments is to identify which of the mechanisms analyzed in the reanalysis can be reproduced
530 in this simplified and controlled setup and are therefore easier to interpret, and which ones require
531 higher complexity or model resolution to be reproduced. The main focus is to analyze the effect of
532 baroclinic wave eddy fluxes on the jet downstream of an imposed local SST maximum intended to
533 mimic the land-sea contrast and the Gulf Stream and introduce a localized maximum in diabatic
534 heating.

535 *a. Idealized aquaplanet simulation setup*

536 The simulations use the Icosahedral Nonhydrostatic Weather and Climate (ICON) model version
537 2.6.5 (Zängl et al. 2015) in an aquaplanet setup with a zonal asymmetry mimicking the land-sea
538 contrast between North America and the Gulf Stream region. These simulations help to better
539 understand the potential effects of change in eddy-zonal flow feedback, which is absent in the
540 two-dimensional model. The simulations have a horizontal resolution of approximately 80 km
541 and 70 vertical levels up to 65 km height. The physical parameterizations used include a one-
542 moment two-category microphysics (Doms et al. 2011), Tiedke convection scheme (Tiedke 1989),
543 a prognostic turbulent kinetic energy scheme for sub-grid turbulent transfer (Doms et al. 2011),

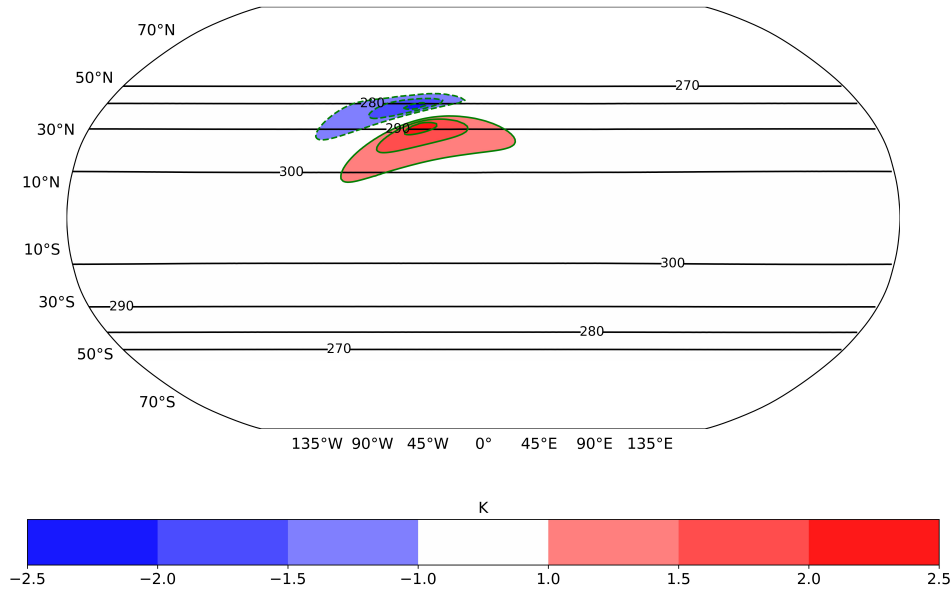
544 non-orographic gravity wave drag (Orr et al. 2010) and the ecRad radiation scheme (Hogan and
545 Bozzo 2018).

546 The atmosphere initialization follows the Jablonowski-Williamson baroclinic wave test case
547 (Jablonowski and Williamson 2006) and the SST is based on the “Qobs” distribution (Neale and
548 Hoskins 2001) with a superposed idealized SST anomaly in the Northern Hemisphere with an
549 amplitude of 10 K and rotated positive and negative ellipsoidal anomalies to represent the effects
550 of both the Gulf Stream and the land-sea contrast in the North Atlantic (Fig. 8). The central point
551 of the perturbation is located at 30°W, while three different latitudinal positions (38, 39 and 42°N
552 are considered to analyze the sensitivity of the jet response to the position of the zonal asymmetry.
553 Further details on the model configuration for these experiments are provided in Schemm et al.
554 (2022).

555 For each experiment with different positions of the SST perturbation, two simulations are run:
556 a control simulation with the above setup including the baseline SST distribution and the SST
557 anomaly shown in Fig. 8 and an additional simulation that includes an SST uniform warming of
558 4 K on top of the setup of the control simulation. The difference between these two simulations
559 constitutes the response to uniform warming. The SST anomaly, which represents the land-sea
560 contrast in this idealized framework, is the same in both simulations. This constitutes an additional
561 simplification since warming has also an impact on land-sea contrast. The response is compared
562 to the trends identified in the reanalysis. This allows us to estimate to which degree the physical
563 mechanisms contained in this idealized setup can explain the trends obtained in the reanalysis. The
564 simulations are run for a period of ten years in a perpetual winter configuration. Note, however,
565 that the trends in ERA5 contain the effects of both natural variability and anthropogenic forcing,
566 while in the aquaplanet simulations the response reflects the impacts of the imposed warming given
567 that the simulation period is long enough to limit the effect of internal variability.

571 *b. Simulation results*

572 The vertical cross section of the change in zonal wind speed, which is shown in Fig. 9a, indicates
573 an equatorward shift at upper-levels while at lower levels the response is characterized by a slight
574 poleward shift of the jet. Therefore, the vertical profile of the positive response in the zonal wind
575 has a stronger meridional tilt than that detected in ERA5 (Fig. 9a). The response to uniform surface



568 **FIG. 8. Aquaplanet simulation setup:** Setup used in the ICON aquaplanet simulations including a zonally
 569 symmetric SST profile represented by black contours from 270 to 310 K, known as “Qobs” and an SST
 570 perturbation (shading and green contours). The anomaly is centered at 30°W, 39°N.

576 warming by 4 K in the unperturbed Southern Hemisphere (SH) shows that in the absence of the
 577 idealized land-sea contrast, a poleward shift of the storm track is obtained (Fig. 9b), similar to
 578 what is projected by CMIP models in the SH. However, in the presence of the SST perturbation
 579 centered at 39°N, the well-marked poleward shift of the jet is suppressed (Fig. 9b,c). Instead
 580 a southwest-to-northeast oriented increase in wind speed near the SST asymmetry is observed
 581 (Fig. 9c). Further poleward, a local decrease in the wind speed is simulated (blue shading in Fig.
 582 9c at 15°W and 70°N).

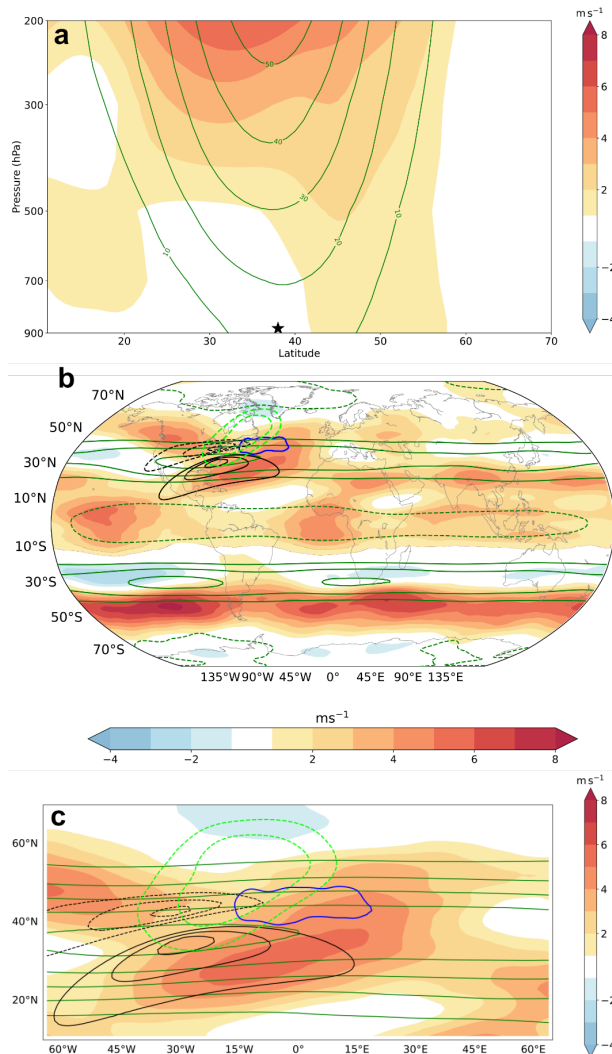
583 Overall, the jet response to warming resembles a southwest-to-northeast elongated pattern of
 584 enhanced wind downstream of the SST anomaly, which has a higher degree of similarity to that
 585 observed over the North Atlantic compared to a pure poleward shift seen in the SST front free SH.
 586 However, there are differences between the response in the idealized simulation and ERA5. The
 587 main contrast is that the slight poleward shift observed over the Gulf Stream is not reproduced
 588 over the SST anomaly. A possible reason why this change is not captured in these simulations
 589 is discussed in the next subsection and relates to local diabatic heating at the SST front and the
 590 findings from the two-dimensional frontal circulation model.

591 1) INCREASE IN DIABATIC HEATING

592 The increase in diabatic heating in the warmer atmosphere has its maximum on the warmer
593 side of the SST anomaly but downstream of its center, where baroclinic disturbances grow into
594 mature systems as they propagate poleward (blue contour in Fig. 9b,c). Latent heat release in the
595 accompanying warm conveyor belts in these systems is likely to contribute to the newly formed
596 maximum in diabatic heating in this sector. Therefore, there is a mismatch in the position of the
597 strongest increase in diabatic heating compared to ERA5 data in which the increase in diabatic
598 heating is located more over the Gulf Stream SST front. A potential cause could be a too weak
599 anchoring effect of diabatic heat release by the idealized SST anomaly in this fairly low-resolution
600 simulation (≈ 80 km grid spacing). For example, Sheldon et al. (2017) showed the inability to
601 simulate the impact of the Gulf Stream on the warm sector of cyclones for a resolution of 40 km,
602 which is twice as high than the resolution used here and in current typical climate simulations.
603 Further studies have demonstrated that higher atmospheric and oceanic resolution leads to a local
604 increase in diabatic heating due to enhanced vertical motion and heat and moisture fluxes over a
605 sharper SST front (Small et al. 2014; Parfitt et al. 2017; Schemm 2023). This may explain why
606 the increase in diabatic heating in this low-resolution simulation is greatest further downstream,
607 where synoptic systems have grown already into mature cyclones that are better resolved by the
608 model. Therefore, it is likely that the local response of the jet to diabatic heating, as suggested by
609 the two-dimensional frontal simulation, which requires the meso-gamma scale to be resolved, is
610 either not well represented or too weak.

623 2) MEAN BAROCLINICITY AND \vec{E} VECTOR CHANGE

624 Next, we compare the time mean baroclinicity response to warming to that found in reanalysis
625 data. At mid- and upper levels, the change in the baroclinicity exhibits a good resemblance to the
626 ERA5 trends (cf. Fig. 10a and Fig. 4a). The zonally averaged (30°W – 45°E) cross section of the
627 baroclinicity change depicts a tripolar pattern consisting of a positive trend at midlatitudes around
628 the jet axis. The positive response reaches down poleward of the climatological mean jet to the
629 500 hPa level. The increase in slope is flanked by a reduction of the isentropic slope equator-
630 and poleward (Fig. 10a). In the horizontal (Fig. 10b), the slope response averaged across the
631 troposphere (250 – 850 hPa) shows a clear triple pattern, with an area of enhanced baroclinicity



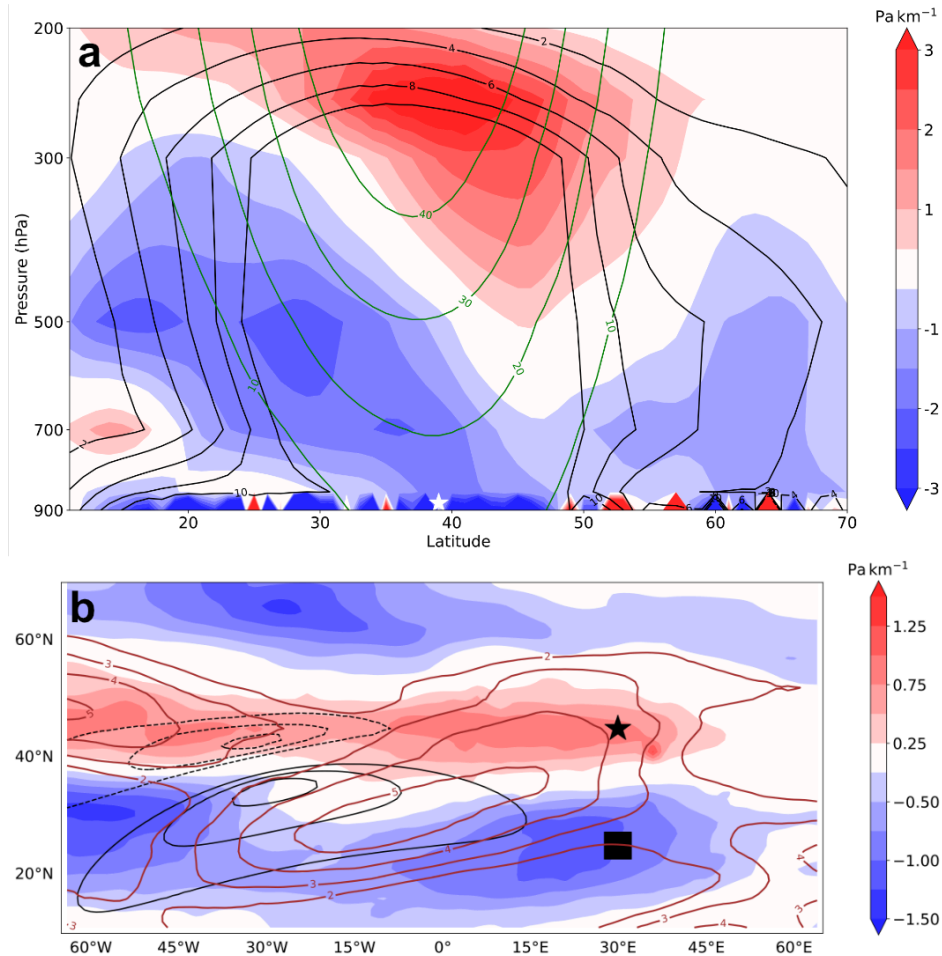
611 **FIG. 9. Response of zonal wind speed to uniform warming in an AGCM aquaplanet simulation:** a) Cross-
 612 section of the zonal mean zonal wind response (shading) between 30°W – 45°E in the aquaplanet simulation
 613 with a uniform warming of 4 K and the control simulation, both with an SST anomaly located at 30°W, 39°N
 614 and climatological average of the zonal mean zonal wind over the same sector in the control simulation (green
 615 contours). The black star above the x-axis indicates the latitudinal position of the SST anomaly. b) Difference
 616 between the zonal wind speed at 250 hPa (shading), climatological mean of the zonal wind speed for the control
 617 simulation (dark green contours, between 30 and 50 ms^{-1} with dashed black contour for zero zonal wind speed),
 618 diabatic heating change (blue contour, 0.4 K day^{-1}) and response in the zonal 500-hPa geopotential anomalies
 619 (light green contours at -400 and -600 $\text{m}^2 \text{s}^2$). The SST anomaly is represented by black contours (from -2.5 to 2.5
 620 K). To facilitate comparison with ERA5 results (Fig. 1b), all fields in panels a and b are shifted 50° westwards.
 621 Continents are shown for illustrative purpose only. c) Zoom of the zonal wind response, climatological zonal
 622 wind, diabatic heating and geopotential anomalies into the area near and downstream of the SST anomaly.

632 poleward of the jet extending downstream to 45°E (red shading in Fig. 10b). Parts of this region
633 downstream of the SST front (little star in Fig. 10b) coincide with the area of increased diabatic
634 heating (blue contour in Fig. 9c).

644 The increase in time mean baroclinicity is associated with changes in eddy momentum flux as
645 indicated by changes in the meridional component of E vectors (Fig. 11). Climatologically, the
646 control simulation is characterized by eastward pointing E vectors at the jet axis, by poleward
647 pointing E vectors poleward of the jet axis and by equatorward pointing E vectors equatorward
648 of the jet axis (Fig. 11a). In agreement with theory, E vector divergence occurs at the jet core
649 indicating eddy momentum flux convergence and acceleration (red shading in Fig. 11a). The
650 response to warming downstream of the SST front is characterized by increased eddy momentum
651 flux convergence in two locations (red shading in Fig. 11b). One increase is centered at 50°N
652 (little star in Fig. 11b), and it is associated with the increase in baroclinicity (little star in Fig.
653 10b) and also enhanced zonal wind speed. From a mechanistic viewpoint, enhanced diabatic
654 heat release increases the time mean baroclinicity, which strengthens the area of baroclinic wave
655 excitation and eddy momentum flux convergence in this region. This mechanism is also found in
656 ERA5. A second increase in eddy momentum flux convergence occurs equatorward of the mean
657 jet position (little square in Fig. 11b). However, this increase results from the poleward shift of
658 the wave excitation region (the main zone of baroclinicity), as it is also observed in the SH and has
659 no associated local increase in baroclinicity (little square in Fig. 10b). In the zonal mean cross
660 section (Fig. 11c), the change in the E vector is primarily characterised by an intensified poleward
661 orientation, with the more equatorward-located change being connected to the poleward shift and
662 the more poleward-located change being related to the increase in the slope.

663 In addition to the changes discussed above associated with the transient eddy mean flow feedback,
664 more locally above the SST front, we find a clear increase in the stationary circulation. (light green
665 contours in Fig. 9b,c). In this area, the response of the stationary circulation indicates the formation
666 of a stationary trough, which appears to cause the zonal wind change locally at the front.

667 As already argued in the previous section, since the diabatic heating is enhanced downstream
668 of the SST anomaly and not above its centre, the local strengthening and poleward push of the jet
669 seen in ERA5 and in the two-dimensional frontal-geostrophic model is not reproduced. A possible
670 cause of this discrepancy is the underresolved transient diabatic processes, which cause this local



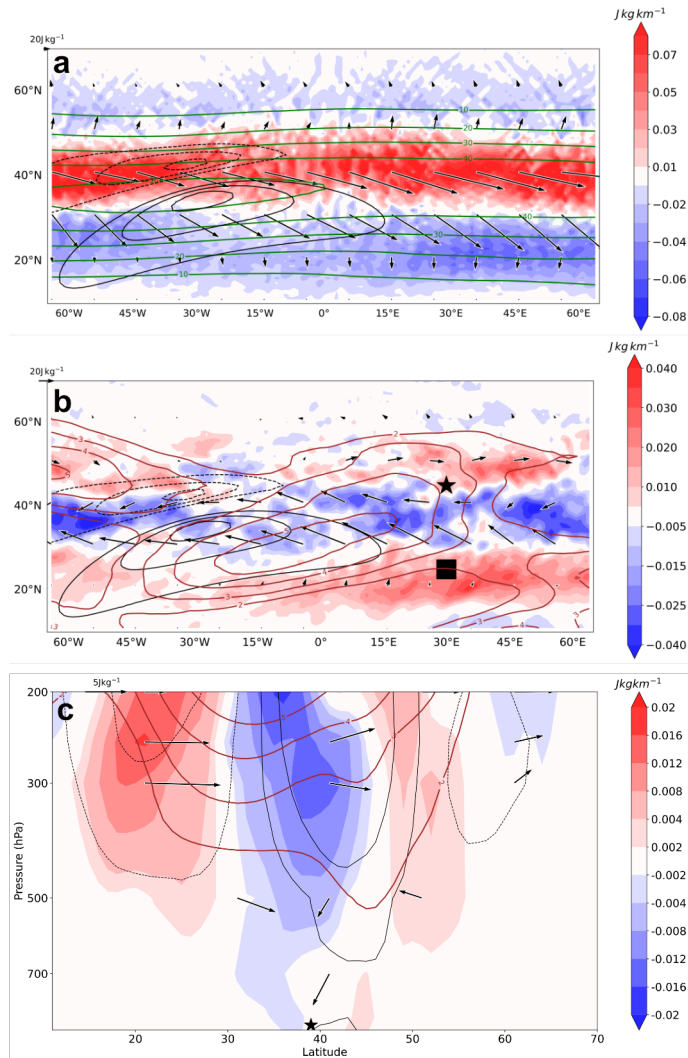
635 **FIG. 10. Response of isentropic slope to uniform warming in an AGCM aquaplanet simulation:** a)
 636 Cross-section of the zonal mean of the response in the slope of isentropic surfaces to a uniform warming of 4
 637 K in the idealized aquaplanet simulations with an SST anomaly centered at 30°W, 39°N. The slope is averaged
 638 over the area between 30°W and 45°E (downstream of the SST anomaly). The climatological mean of the slope
 639 in the control simulation is represented by black contours and the zonal mean of the zonal wind speed by green
 640 contours. The white star above the x-axis represents the latitudinal position of the SST perturbation. b) Slope
 641 response to the uniform warming averaged between 250 and 850 hPa (shading) and zonal wind response at 250
 642 hPa in the control simulation (brown contours). Black contours represent the SST anomaly. The star and the
 643 square mark regions of increased and decreased slope downstream of the SST anomaly (see text for discussion).

671 effect to be too weak and may be a major source of misrepresentation of the North Atlantic jet
 672 trend in low-resolution models. A thorough investigation of the effect of the model resolution on

673 diabatic processes is beyond the scope of this study, but diabatic heating increases at higher model
674 resolution have recently been demonstrated in a comparable aquaplanet setup by Schemm (2023).

686 To summarize, the key finding of this experiment is that a zonal surface asymmetry, here an
687 idealized representation of the land-sea contrast over the Gulf Stream sector, is sufficient to suppress
688 the general tendency of the jet to shift poleward under warming. The response to warming of the
689 idealized simulation includes major factors also found in reanalysis-based trends of the North
690 Atlantic jet: increased diabatic heating near the SST anomaly, increased tropospheric baroclinicity
691 at upper levels with a triple pattern of change throughout the troposphere at the mean storm track
692 location, and increased convergence of eddy fluxes feeding back on the zonal flow downstream of
693 the front causing an increase in the jet speed.

694 There are however some differences between this idealized run and reanalysis, namely that the
695 positive trend in baroclinicity does not extend to the surface (Fig. 10a). Some possible causes for
696 this mismatch could be a difference in ocean changes and air-sea exchange (Woollings et al. 2012)
697 between the reanalysis and idealized simulations or a weaker diabatic heating within baroclinic
698 weather systems. In addition, the idealized jet stream in the control simulation has a more zonal
699 orientation compared to the southwest-northeast tilt identifiable in ERA5. The lack of orography
700 in the idealized simulation could partially explain this difference in the jet orientation (Brayshaw
701 et al. 2009). In addition, although the eddy momentum convergence coincides with the area of
702 increased baroclinicity and stronger wind speed downstream of the SST anomaly and poleward
703 of the mean jet position (little stars in Figs. 10b and Fig. 11b), the pattern is different from
704 ERA5 near the SST perturbation, leading to a different jet response in this region, mainly driven
705 by changes in the stationary circulation. There is a positive response to warming of the E vector
706 divergence, which is not linked to a local increase in baroclinicity (little squares Figs. 10b and Fig.
707 11b). Furthermore, the increase in diabatic heating is located more downstream in the idealized
708 simulation in contrast to the reanalysis, where it is more anchored over the Gulf Stream. A cause
709 for this mismatch could be the coarse resolution of the idealized experiments. The local effect
710 of transient diabatic heating occurring during extratropical cyclone growth over the Gulf Stream
711 analyzed in the frontal-geostrophic simulation, which locally accelerates the jet and pushes it
712 poleward in both the frontal-geostrophic model and ERA5, appears thus too weak to affect the
713 mean response and thus, the changes near the SST anomaly are in contrast to the reanalysis.



675 **FIG. 11. Response of the E vector to uniform warming in an AGCM aquaplanet simulation:** a) Clima-
 676 tological mean of the E vector and its divergence (arrows and shading) in the control simulation at 250 hPa and
 677 climatological mean of the zonal wind speed in the control simulation (green contours). b) Response of E vector
 678 and its divergence to uniform warming at 250 hPa (arrows and shading, respectively). Brown contours depict
 679 the response to warming in the zonal wind at the same pressure level. The SST anomaly is represented by black
 680 contours in panels b and c. The star and the square mark regions of increased E vector divergence downstream of
 681 the SST anomaly (see text for discussion). c) Cross-section of the zonal mean difference of E vector divergence
 682 between warmed and control simulation (shading), climatological mean of the E vector divergence in the control
 683 simulation (black contours, between $-3 \cdot 10^{-5}$ and $3 \cdot 10^{-5} \text{ J kg}^{-1} \text{ m}^{-1}$), E vector response to uniform warming
 684 (arrows) and zonal wind response (brown contours) in the aquaplanet simulation with an SST perturbation located
 685 at 30°W, 39°N. The black star above the x-axis indicates the latitudinal position of the SST perturbation.

714 3) SENSITIVITY TO THE POSITION OF THE SST ANOMALY

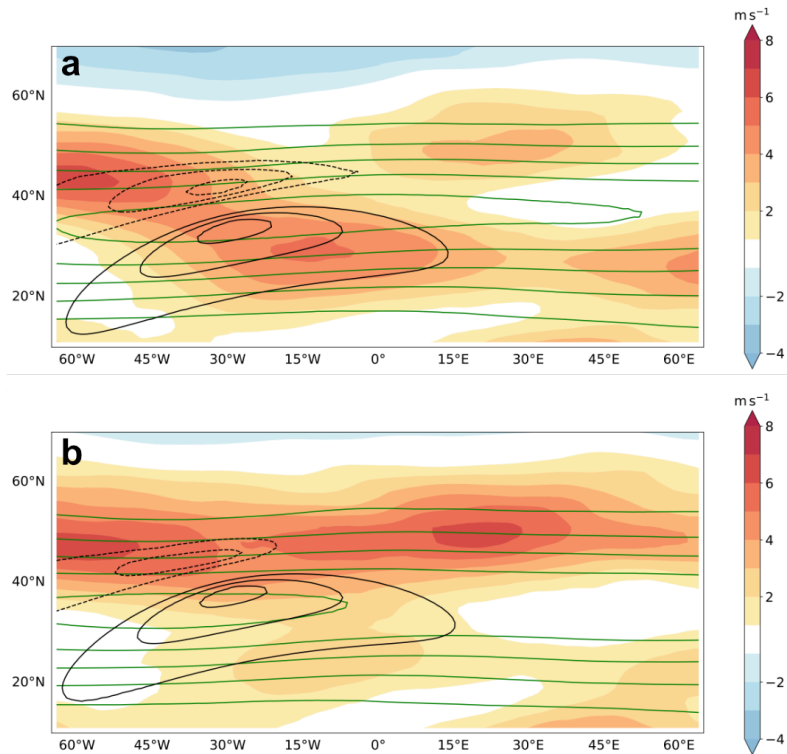
715 The purpose of the next simulations is to change the exact location of the convergence of the
716 upper-level eddy momentum flux and increase in diabatic heating relative to the mean jet position.
717 This is done by changing the location of the SST anomaly (to 38 and 42°N) relative to the mean
718 jet position, which is set by the large-scale SST gradients according to the Qobs SST distribution.

719 The response of the storm track to warming downstream of the SST anomaly (east of 0°) in both
720 simulations is characterized by an increase of the zonal wind speed poleward of the climatological
721 jet position (Fig. 12). Both simulations feature a triple pattern in the baroclinicity response
722 (Fig. 13a,b), which is more confined in the upper levels and is also slightly displaced towards
723 the equator in the simulation with the SST front at 38°N (Fig. 13c). In both simulations, the
724 increase in diabatic heating is located downstream of the idealized land-sea contrast (not shown),
725 in line with the simulation with an SST front at 39°N (Fig. 9b,c). In this downstream area (little
726 star in Fig. 13c,d), the response is dominated by an increase in baroclinicity and a corresponding
727 intensification of eddy momentum flux convergence. This behavior is independent of the position
728 of the SST asymmetry.

729 Conversely, the response near the SST anomaly differs between the two front positions. In the
730 case of a more equatorward SST anomaly, the jet shifts equatorwards (Fig. 12a) and the difference
731 between the warmed and control simulations features poleward pointing E vectors and a positive
732 difference in E vector convergence over the storm track entry region equatorwards of the mean jet
733 position (Fig. 13c), which pushes the jet equatorward. In the other sensitivity simulation with a
734 more poleward located SST anomaly, the jet clearly shifts poleward near the SST front and up-
735 and downstream of it (Fig. 12b), similarly to what occurs in the absence of the SST anomaly in
736 the SH (Fig. 9b). Accordingly, slope and eddy momentum convergence also shift polewards under
737 warming (Fig. 13b,d) and the increase in the isentropic slope is much less confined to upper levels.

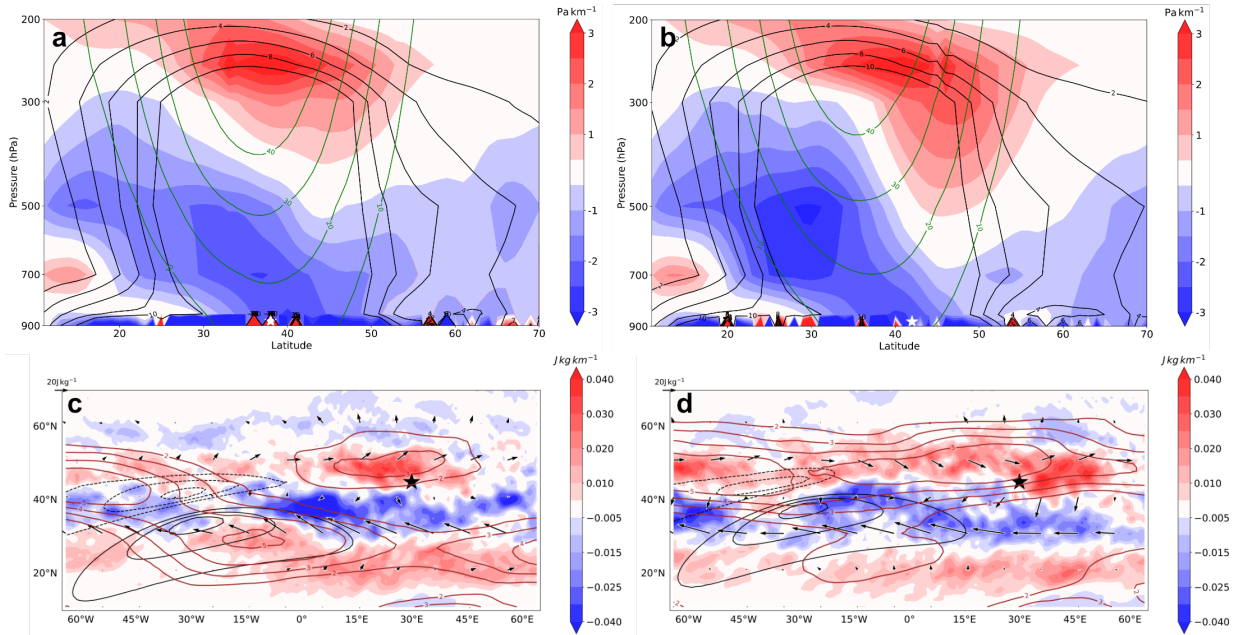
738 In the unperturbed SH there is also some variability between different simulations (not shown),
739 but the response is qualitatively similar featuring a poleward shift in all simulations, in contrast to
740 the NH where the position of the SST anomaly has a higher impact on the jet response than the
741 variability detected in the SH.

742 The three experiments with different positions of the SST anomaly show that if this perturbation
743 is located near the mean jet position the response to uniform is not a poleward shift as it occurs



751 FIG. 12. **Sensitivity of the zonal wind response to the position of the SST anomaly in aquaplanet**
 752 **simulations:** Difference between the zonal wind speed in the aquaplanet simulations with a uniform warming
 753 of 4 K and the control simulation at 250 hPa (shading), and climatological zonal wind speed in the control
 754 simulation (green contours between 30 and 50 ms^{-1}) with an SST anomaly located at 30° W, a) 38°N and b)
 755 42°N. Black contours represent the SST perturbation.

744 when there is no anomaly or when it is located far from the mean jet position. The role of changes in
 745 stationary circulation for the wind response near the anomaly is strongly dependent on its position
 746 as shown in the experiments with the SST perturbation at 38°N, where the stationary circulation has
 747 a limited influence, and the simulation with the SST anomaly at 39°N, where the response in this
 748 sector is dominated by changes in the stationary circulation. Downstream of the SST asymmetry,
 749 the response in all simulations is dominated by an increase in diabatic heating, which enhances
 750 baroclinicity and eddy momentum convergence poleward of the mean jet position.



756 **FIG. 13. Sensitivity of the baroclinicity and E vector response to the position of the SST anomaly in**
 757 **aquaplanet simulations:** Cross-section of the response of the slope of the isentropic surfaces under global
 758 warming (shading) and climatological mean of slope and zonal wind in control simulations (black and green
 759 contours, respectively) with an SST anomaly located at 30° W, a) 38°N and b) 42°N. White stars above the x-axis
 760 in panels a and b represent the latitudinal position of the SST anomaly. c), d) Response of the E vector and its
 761 divergence (arrows and shading, respectively) at 250 hPa for both sensitivity experiments and response in the
 762 zonal wind at the same pressure level (brown contours). The SST anomaly is represented by black contours in
 763 panels c and d. The stars in panels c and d mark a region of increased E vector divergence downstream of the
 764 SST anomaly (see text for discussion).

765 5. Fully-coupled CESM ensemble simulations with SSP3-7.0 scenario

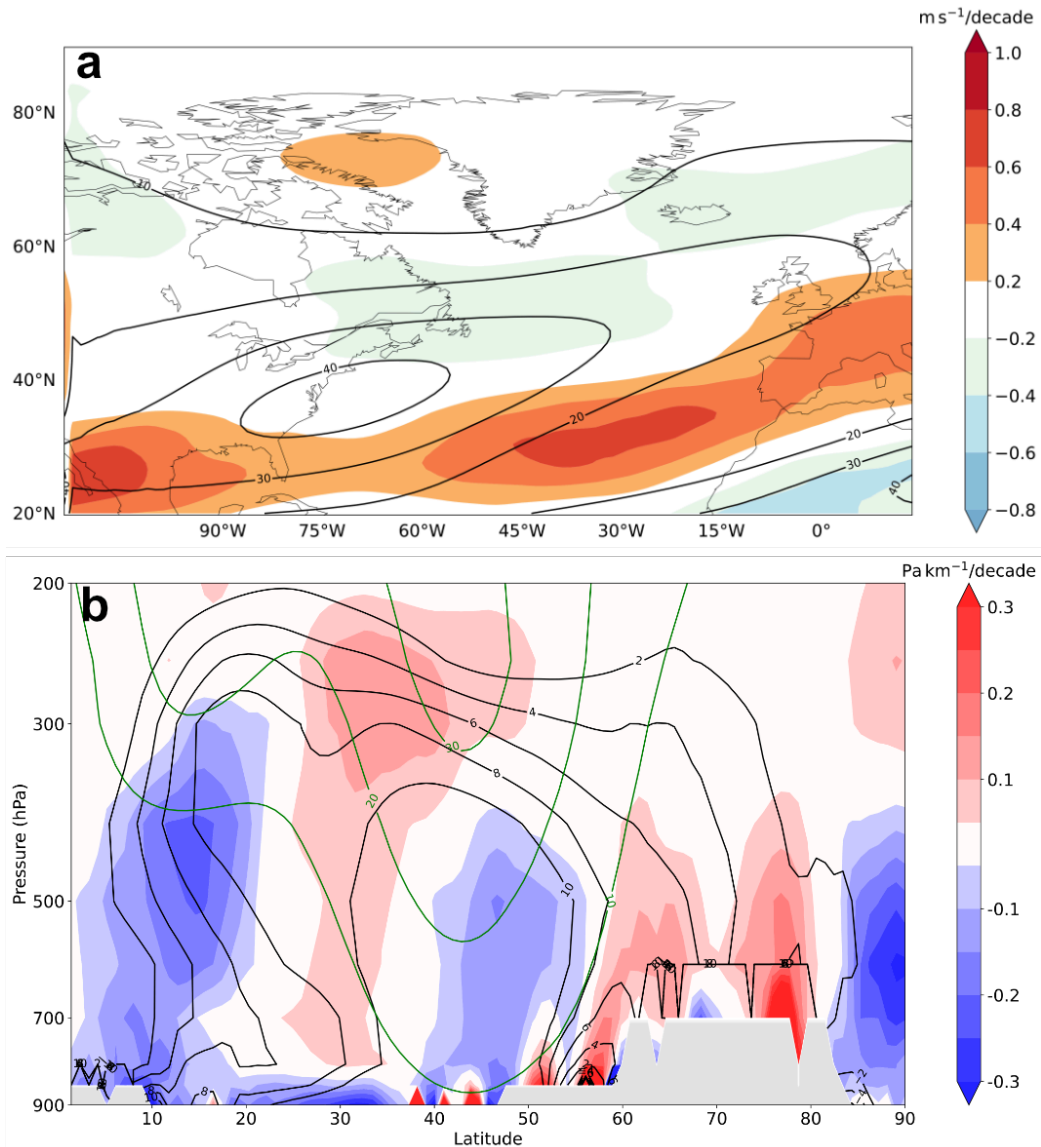
766 Finally, we examine the extent to which the trends identified in the reanalysis are reproduced
 767 by a historical run of fully-coupled climate model over the ERA5 period. It has been recognized
 768 that the current generation of fully-coupled Earth system models appears to be missing the trend
 769 in the North Atlantic jet stream (Blackport and Fyfe 2022). Here, we examine zonal wind and
 770 baroclinicity trends in an ensemble of fully-coupled climate simulations.

771 *a. CESM model setup*

772 The ensemble consists of 5 climate simulations produced with the Community Earth System
773 Model (CESM) version 2.1.2 (Danabasoglu et al. 2020), labeled 0900 to 1300. These are used to
774 compare the ERA5 trends with simulated historical trends and to analyze future projections for the
775 jet stream. The model is run in fully coupled mode including the Community Atmosphere Model
776 (CAM6) (Bogenschutz et al. 2018; Danabasoglu et al. 2020) with 32 vertical levels, the Community
777 Land Model (CLM5) (Lawrence et al. 2019), the Parallel Ocean Program version 2 (POP2, 60
778 vertical levels), the Los Alamos National Laboratory Sea Ice model (CICE5) (Hunke et al. 2015),
779 and the hydrological routing model Model for Scale Adaptive River Transport (MOSART) (Li
780 et al. 2013). The horizontal resolution is approximately 1° . The historical period with prescribed
781 forcing covers from 1850 to 2014, and from 2015 to 2100 the runs are forced with the SSP3-7.0
782 scenario, which constitutes a medium to high forcing pathway (O'Neill et al. 2016). To provide a
783 more meaningful comparison with the ERA5 data, the period 1980-2022, which is partially driven
784 by the SSP3-7.0 forcing, is used as the historical period.

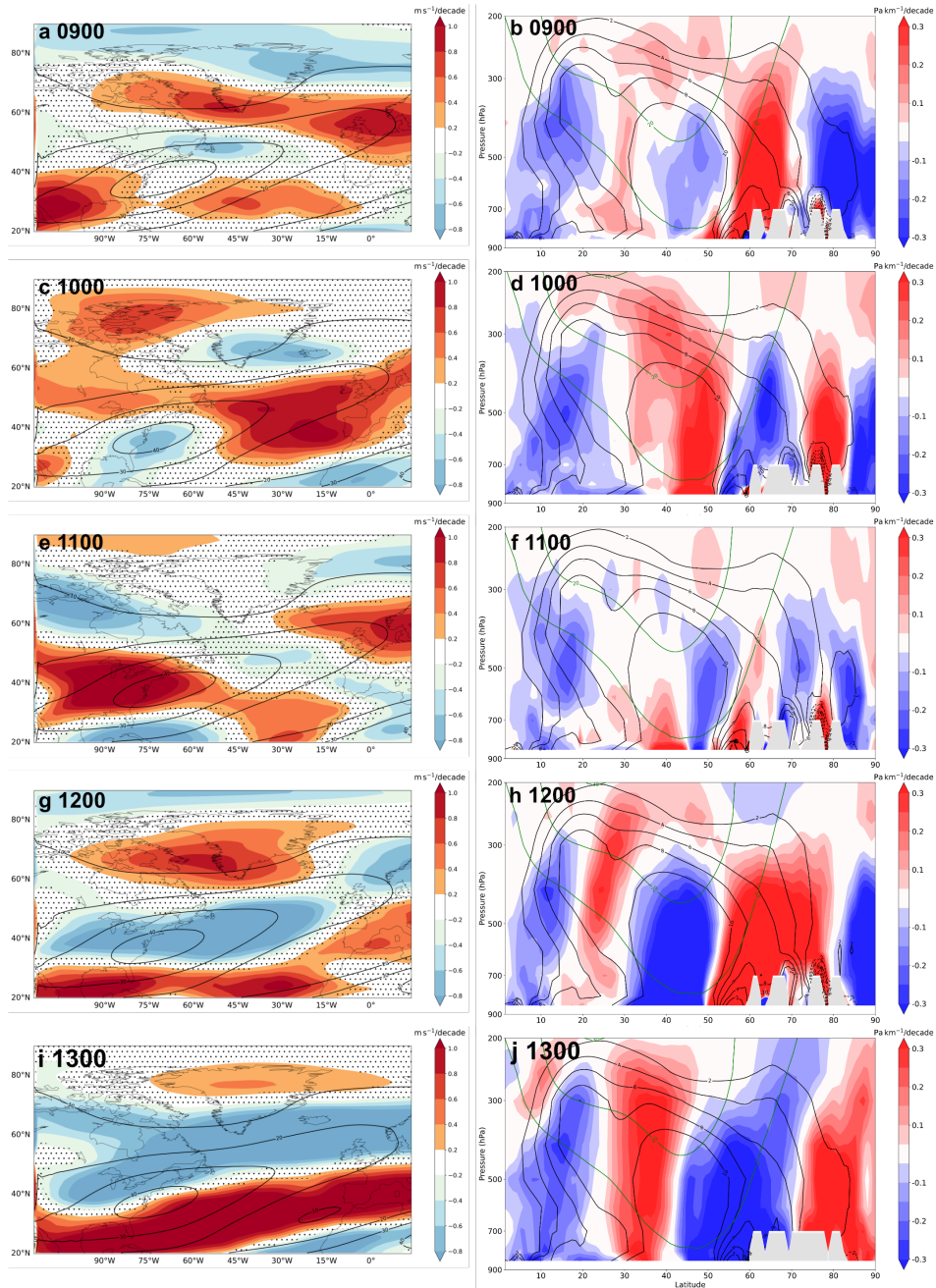
785 *b. Historical period (1980–2022):*

786 The ensemble mean trend for the historical period (1980–2022) displays a well-marked equa-
787 torward shift of the jet over the Gulf Stream region and downstream extension (Fig. 14a), which
788 is different from the local poleward shift identified in the reanalysis over the Gulf Stream sector.
789 A closer inspection of the ensemble mean of the slope trend shows that also the increase in the
790 baroclinicity is located equatorwards with respect to the jet position (between $25\text{--}40^\circ\text{N}$, Fig. 14b).
791 Considering the role of the local increase in transient diabatic heating in pushing the jet slightly
792 poleward, this effect seems to be either absent or not strong enough to result in a local increase and
793 poleward displacement of the jet stream. Nevertheless, some caution is required to interpret these
794 differences as model error given the small ensemble size, which may not be sufficient to isolate the
795 forced response. In addition, reanalysis trends as presented in section 2 are the result of a combi-
796 nation of external forcing and natural variability, which is particularly large in the historical period
797 as seen in the CESM ensemble spread. Appendix B further illustrates this aspect by analyzing
798 daily zonal wind data for a subset of 50 members of the CESM2 large ensemble (CESM-LENS)
799 (Rodgers et al. 2021).



800 FIG. 14. DJF trends in 250-hPa zonal wind speed and in a vertical cross-section of the slope of the
 801 isentropic surfaces for CESM simulations in the historical period (1980-2022): a) Trend in zonal wind
 802 speed (shading) and climatological mean (black contours) for the historical period. b) Trend in the slope of
 803 the isentropic surfaces (shading) and climatological mean (black contours) over the North Atlantic storm track
 804 (80–15° W). Zonal wind speed climatology is represented by green contours.

808 However, the ensemble spread of wind speed trends in the CESM ensemble simulation in the
 809 historical period is remarkably large, as is the direction of the meridional shift of the jet among
 810 the different ensemble members (Fig. 15). This diversity is aligned with different responses of



805 FIG. 15. DJF trends in 250-hPa zonal wind speed and in a vertical cross-section of the slope of the
 806 isentropic surfaces for each CESM ensemble member in the historical period (1980-2022): As in Fig. 14
 807 for each ensemble member.

811 the baroclinicity to warming across the ensemble, similarly to the effect of shifting the position of
 812 the SST anomaly in the aquaplanet simulations. In particular, there are differences in the exact

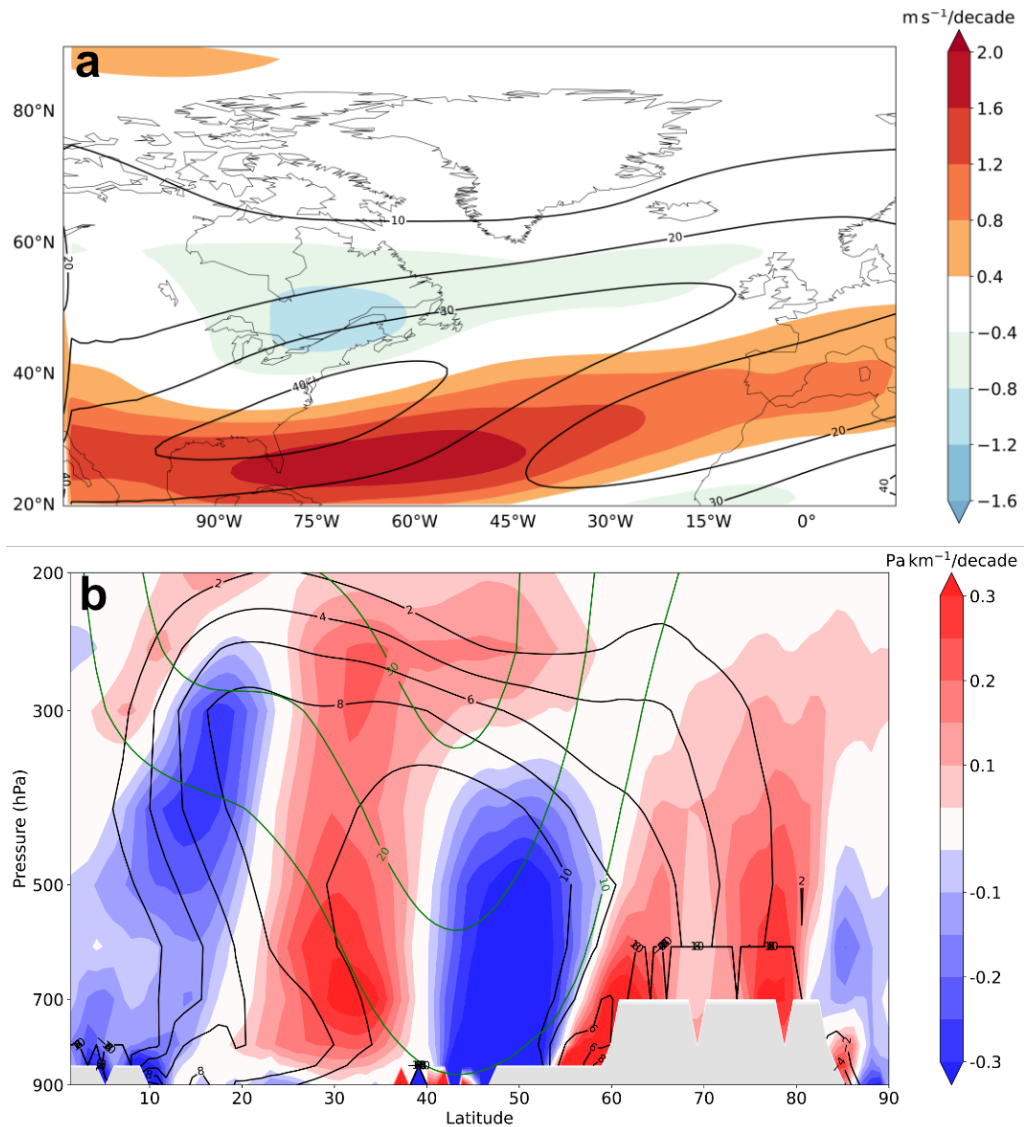
813 position of the positive baroclinicity trend relative to the climatological mean, which is essential in
814 determining the jet shift (Yuval and Kaspi 2016). For example, member 0900 displays a poleward
815 shift with a positive zonal wind trend around 60°N , which is accompanied by a positive slope in
816 trend in the same region (Fig. 15a,b). Conversely, member 1300 displays a tripolar pattern of
817 change (Fig. 15j) as expected from zonal asymmetries (Schemm et al. 2022). However, the positive
818 slope trend is also for this member located slightly equatorward of the mean jet position and
819 consequently the positive wind trend is displaced to the south (Fig. 15i), similar to the ensemble
820 mean (Fig. 14). The remaining members exhibit larger differences some members produce a
821 poleward shifted positive baroclinicity trend but no downstream extension, while some do not
822 reproduce the tripolar pattern in the slope trend at all. At upper levels, all members exhibit a
823 negative trend in stability due to either stratospheric cooling or reduced warming in midlatitudes,
824 which results in an overall positive slope trend, but there is widespread disagreement on the location
825 of the positive slope trend.

826 As a consequence of the large ensemble spread, the ensemble mean trend in momentum conver-
827 gence is close to zero as opposing trends in individual ensemble members cancel each other (not
828 shown). While the closest member to ERA5 shows a significant increase in the northward com-
829 ponent of the E vector, other members produce lower increases or even an equatorward pointing
830 trend.

831 *c. End of century (2057–2100)*

832 The ensemble mean of the trends for the end of the century remains fairly similar to that of
833 the historical period but is amplified (Fig. 16a). Again, baroclinicity features a notable increase
834 equatorward of the mean jet position between $25\text{--}35^{\circ}\text{N}$ throughout the troposphere and a negative
835 trend between $40\text{--}50^{\circ}\text{N}$ at mid- and low levels (Fig. 16b), which indicates that the exact location of
836 the modeled positive trend in baroclinicity is a potential source of uncertainty in future projections.

840 Differences in the ensemble spread are now reduced compared to the historical period. At the
841 end of the century, all ensemble members project an equatorward shift of the jet which is well
842 reflected in the ensemble mean (Fig. 16b). Contrary to what is found for the historical period, now
843 all members produce a positive slope trend around or equatorward of the mean jet position at low
844 levels (Fig. 17b) but 10° too equatorward compared to observed recent trends. This is associated



837 FIG. 16. Trends in 250-hPa zonal wind speed and in a vertical cross-section of the slope of the isentropic
 838 surfaces for CESM simulations in the end of the century period (2057-2100): As in Fig. 14 for the end of
 839 the century period. Note the different color scale for the wind trend compared to Fig. 14.

845 with a small warming trend at midlatitudes, which increases the meridional potential temperature
 846 gradient and thus, the slope. At upper levels, the pattern is similar to the one obtained for the
 847 historical period. In general, all members display a well-marked equatorward shift (Fig. 16a),
 848 similar to the aquaplanet simulation with zonal SST anomalies centered at 39°N. Further, the
 849 triple pattern in the slope and potential temperature trends emerges (Fig. 16b). Therefore, the
 850 baroclinicity and the jet trend seem both placed too far to the equator compared to recent trends in

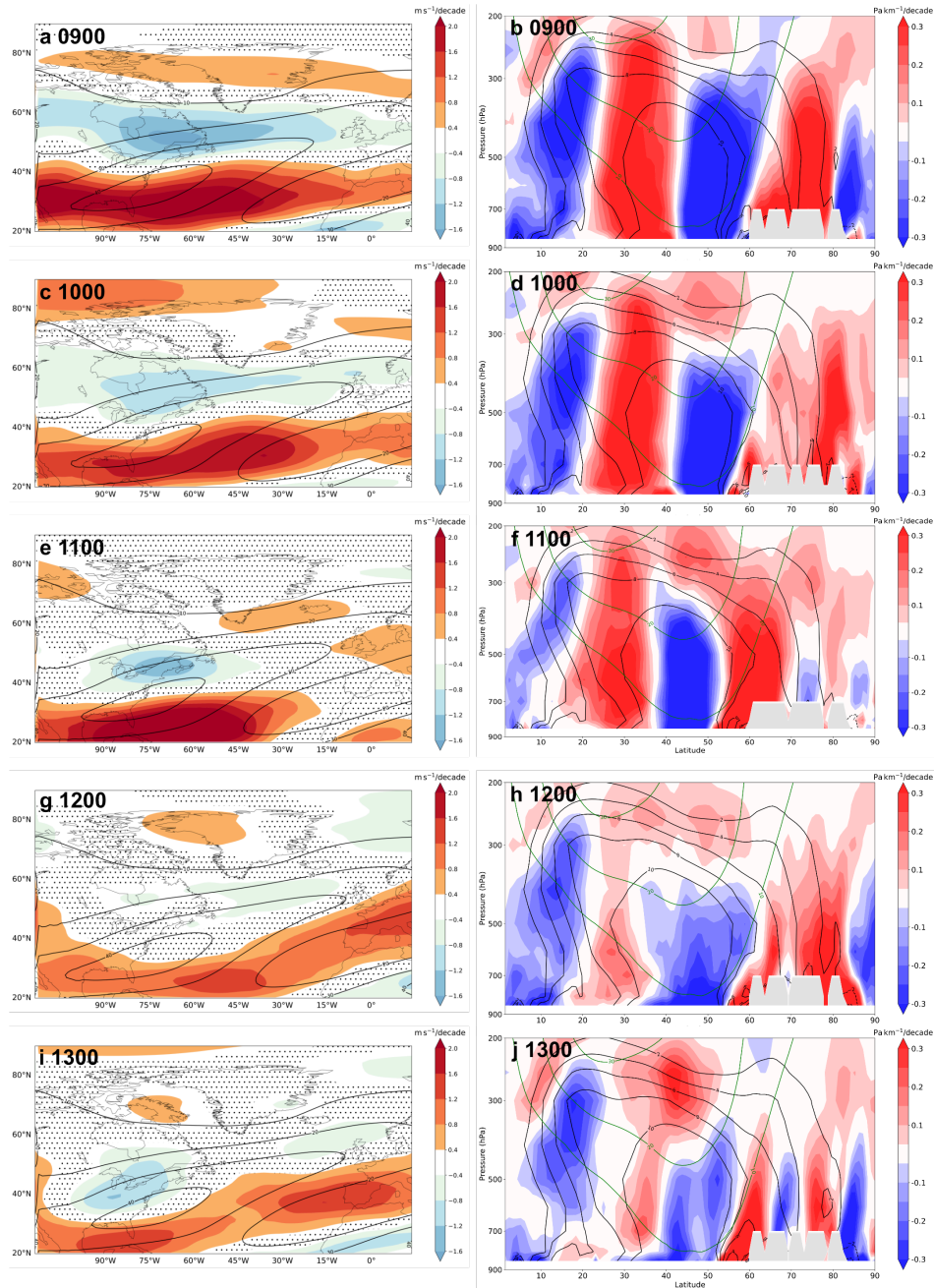
851 ERA5. The reduced ensemble spread at the end of the century and the large spread in the more
852 recent period indicate a small signal to noise ratio in the historical period, while it is stronger at the
853 end of the century, as a consequence of intensified warming in this period (Rodgers et al. 2021).

857 Overall, we find that the results over the Gulf Stream sector for the 5-member ensemble are
858 different from the reanalysis with an equatorward shift in the ensemble mean of the CESM
859 simulations, and a slightly poleward shift in ERA5. The downstream extension is found in
860 both reanalysis and CESM ensemble simulations, but the extension affects the Iberian peninsula
861 in CESM while it affects the UK in reanalysis data, due to the equatorward shift projected for
862 the entire North Atlantic in the climate simulations. Also the triple pattern of change is found in
863 vertical cross sections of changes in the isentropic slope, but again shifted towards the equator in
864 CESM. This mismatch is reduced when consideration is given to the CESM-LENS in the historical
865 period, where some members indeed feature a polewards shift as seen in ERA5, while the majority
866 of members still feature an equatorward shift (see Appendix B).

867 The signal for the end of the century consists of an equatorward shift over the Gulf Stream and
868 an extension of the jet towards southwestern Europe. As shown by the results of the idealized two-
869 dimensional frontal-geostrophic experiment, an increase in diabatic heating over the Gulf Stream
870 region leads to a slight poleward shift of the jet. Therefore, it remains an open question whether the
871 trend for the end of the century detected in the CESM climate simulations is the actual response
872 to the increase in greenhouse gases or it is the result of a misrepresentation of the local effect of
873 diabatic heating over the Gulf Stream. A more detailed analysis using a large ensemble as well
874 as higher resolution simulations, such as those produced by HighResMIP Haarsma et al. (2016),
875 is required to prove this aspect. Given the coarse resolution at which both the aquaplanet and the
876 fully-coupled climate simulations are run, this aspect could have an impact on the trends over the
877 Gulf Stream following the resolution argument of Sheldon et al. (2017). More research on the role
878 of model resolution in setting the jet response to warming is needed to quantify its effect on the jet
879 stream trends.

880 **6. Conclusions**

881 Atmospheric dynamics play an important role in the development of the North Atlantic jet
882 stream and storm tracks. In this study, two main mechanisms are outlined to better understand



854 FIG. 17. Trends in 250-hPa zonal wind speed and in a vertical cross-section of the slope of the isentropic
 855 surfaces for each CESM ensemble member in the end of the century period (2057-2100): As in Fig. 15 for
 856 the end of the century period. Note the different color scale for wind trends.

883 trends in the position and strength of the North Atlantic jet found in reanalysis data. At the heart
 884 of both is the effect of enhanced diabatic heating, which (i) produces a local transient negative PV

885 anomaly at upper levels (Wernli and Davies 1997; Pomroy and Thorpe 2000), and (ii) increases
886 the baroclinicity in the storm track entrance region at the core of the mean jet, which results in
887 enhanced eddy momentum fluxes downstream (Hoskins et al. 1983; Vallis 2017). We explore
888 each of these mechanisms with different diagnostics and modeling approach in the spirit of Peng
889 and Whitaker (1999) by analyzing the ERA5 reanalysis and a hierarchy of model simulations
890 with different complexity including a two-dimensional frontal-geostrophic simulation, aquaplanet
891 experiments with an atmospheric global circulation model and fully-coupled climate simulations.

892 Locally, over the Gulf Stream region, where synoptic waves grow but yet do not break, an
893 intensification and poleward shift of the jet is consistent with the response to more intense transient
894 diabatic heating pulses during the growth and rapid succession of growing baroclinic waves
895 (mechanism i). The response predicted by a frontal-geostrophic model forced with a heating
896 anomaly inspired in the long-term trend observed in ERA5 is sufficient to explain the magnitude of
897 the jet change seen in reanalysis data. However, this effect seems to be absent (or too weak) in the
898 aquaplanet, but also in the ensemble of CESM climate simulations. A potential cause is related to
899 the coarse model resolution and a too weak representation of the local diabatic effect and possibly
900 is one cause of the failure of climate model to capture the North Atlantic jet response to warming
901 as highlighted by Blackport and Fyfe (2022). Further analysis with larger ensembles and higher
902 resolution runs, for instance HighResMIP simulations (Haarsma et al. 2016) is required to quantify
903 the impact of model resolution on this mechanism.

904 Downstream of the Gulf Stream sector, the jet is further modified in ERA5 through an eddy
905 feedback, which is consistent with an increase in baroclinicity through diabatic heating (mechanism
906 ii) and intensified propagating Rossby waves away from the main wave source (Vallis 2017). This
907 mechanism affects the jet over the whole North Atlantic in contrast to the local effect over the
908 Gulf Stream produced by more intense diabatic heating pulses. The concomitant meridional
909 propagation of Rossby waves away from the main wave source is followed by a strengthening of
910 the eddy momentum convergence resulting in an intensification and zonal extension of the jet.
911 The mechanism is demonstrated through analysis of changes in E vectors in idealized aquaplanet
912 warming simulations in the presence of an SST front. The chain of mechanisms consists of
913 increased diabatic heating slightly downstream of the SST front, intensified mean baroclinicity and
914 enhanced eddy momentum convergence leading to an increase of the wind speed downstream of

915 the front. The mechanism seems to be rather independent of the exact position of the SST front.
916 However, the jet response at the location of the SST front is strongly influenced by the position
917 of the front and varies from a local poleward to an equatorward jet shift depending on the exact
918 position of the SST front. For the latter case, the formation of a stationary trough over the front is
919 a relevant factor in explaining the shift in the aquaplanet experiments.

920 The different methods considered here aim to foster a mechanistic understanding of the jet
921 response, primarily to increased diabatic heating. They could also help to explain the failure of
922 coarse climate models to simulate the recent trends in the North Atlantic (Blackport and Fyfe
923 2022). In our experiments, the 5-member ensemble of fully-coupled model in particular failed
924 to simulate the local poleward shift over the Gulf Stream region observed in ERA5. A larger
925 ensemble size is required to capture the observed trend within the ensemble spread, as can be
926 seen in the CESM-LENS (Rodgers et al. 2021). The idealized aquaplanet simulations are unable
927 to adequately represent the jet trend near the zonal asymmetry, which is intended to mimic the
928 effects of land-sea contrast and the Gulf Stream. This is, as indicated above, possibly due to
929 underrepresentation of the local effect of diabatic heat pulses, which has the potential to shift the
930 North Atlantic poleward and strengthen it. This goes hand-in-hand with a coarse resolution in the
931 SST fields (Sheldon et al. 2017), which strongly affects the location of enhanced baroclinicity and
932 diabatic heating. In the previously cited study, a well-resolved SST anomaly above the Gulf Stream
933 is shown to anchor the diabatic heating more consistently above its warm side and km-scale models
934 have been shown to increase the magnitude of the diabatic heating over the SST anomaly (Schemm
935 2023). The implication from our study is that a better anchoring and increased amplitude of local
936 transient heating is in general able to push the jet poleward. The coarse idealized aquaplanet
937 simulations reproduce some of the main aspects of the mechanisms previously outlined, namely
938 increased tropospheric baroclinicity at upper levels, and increased convergence of eddy momentum
939 flux associated with Rossby wave propagation downstream of the zonal SST asymmetry, but the
940 experiments also miss the local jet response directly above the SST front. Thus, the anchoring
941 effect and influence of local diabatic heating pulses are likely too weak and the area of increased
942 diabatic heating is moved downstream of the main cyclogenesis region pinpointing again to too
943 coarse model resolution.

944 The analysis performed here provides a dynamical interpretation of the trends detected in the
945 reanalysis, but an attribution to either anthropogenic forcing or natural variability has not been
946 conducted. There are multiple dynamical adjustment methods that aim to disentangle the forced
947 response of a magnitude of interest, usually temperature or precipitation, from natural variability.
948 These techniques have shown satisfactory results in previous studies (Smoliak et al. 2015; Lehner
949 et al. 2017; Guo et al. 2019; Sippel et al. 2019; Heinze-Deml et al. 2021). A foreseen study will
950 aim at adapting this technique to the North Atlantic jet stream trends to investigate which changes
951 are attributable to anthropogenic climate change.

952 *Acknowledgments.* This work was supported by the Swiss National Science Foundation through
953 project grant Nr. 204181.

954 *Data availability statement.* The ERA5 data are publicly available on the Copernicus Cli-
955 mate Change Service (C3S) Climate Data Store: <https://doi.org/10.24381/cds.adbb2d47>
956 (Hersbach et al. 2023). ICON simulations used for this paper will be archived at ETH Zurich's
957 Research Collection for scientific publications and research data under a CC-BY 4.0 license for
958 at least the upcoming 15 years (doi will be provided after acceptance). ETH Zurich's Research
959 Collection adheres to the FAIR principles. CESM simulation data will be stored at the Institute
960 of Atmospheric and Climate Science, ETH Zurich, for at least 10 years and will be available on
961 request.

962 APPENDIX A

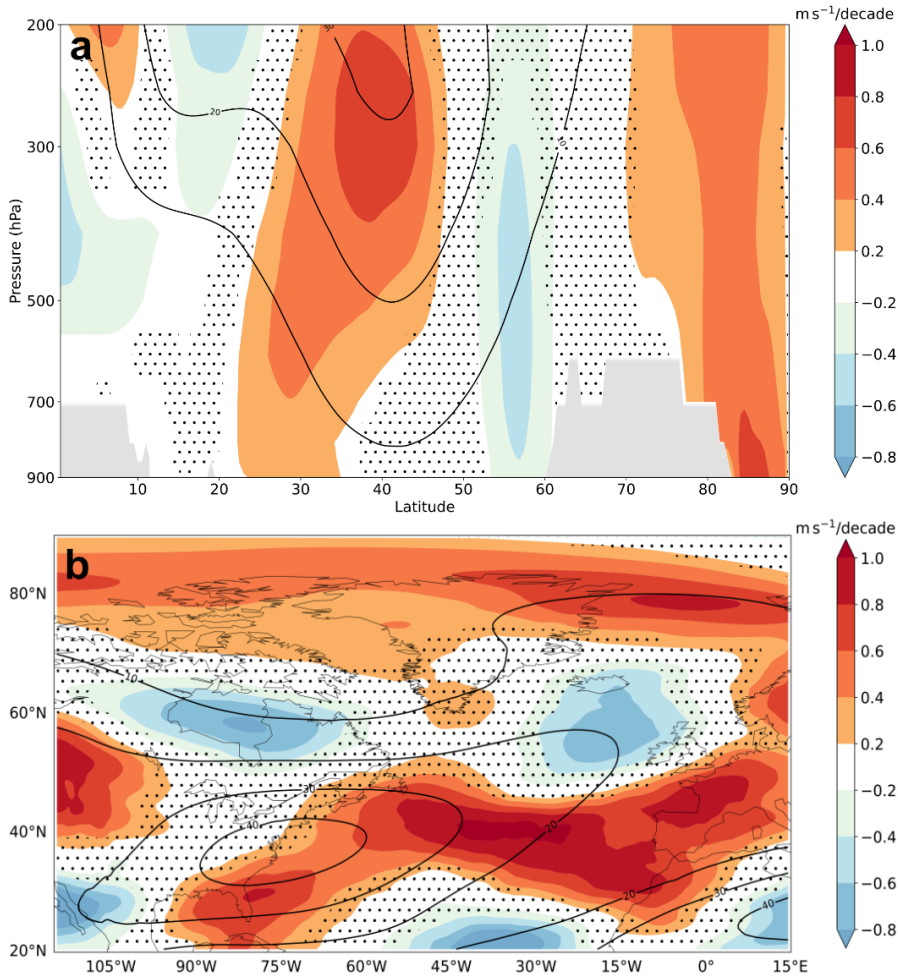
963 **Zonal wind trends in ERA in the period 1940-1978**

964 Wintertime trends in ERA5 before the satellite era (1940-1978) show an equatorward shift of
965 the jet in the Gulf Stream region with a defined southwest-northeast orientation (Fig. A1). This
966 contrasts with the trend for the more recent period, which is mainly characterized by a slight
967 poleward shift of the jet over this region (Fig. 1). Downstream over central and eastern North
968 Atlantic there is a strong intensification, an equatorward shift and an extension of the jet affecting
969 southwestern Europe. The trend shows a decrease of zonal wind speed over the Hudson Bay and
970 between Iceland and the British Isles and some areas at low latitudes, but there is no clear triple
971 pattern in the trends for this period. In addition, there is a clear positive trend at high latitudes,
972 in contrast to the later period shown in Fig. 1. The significant trends are mainly located over
973 the ocean, where the reanalysis might not be well constraint in this period given the scarcity of
974 observations in the considered period.

980 APPENDIX B

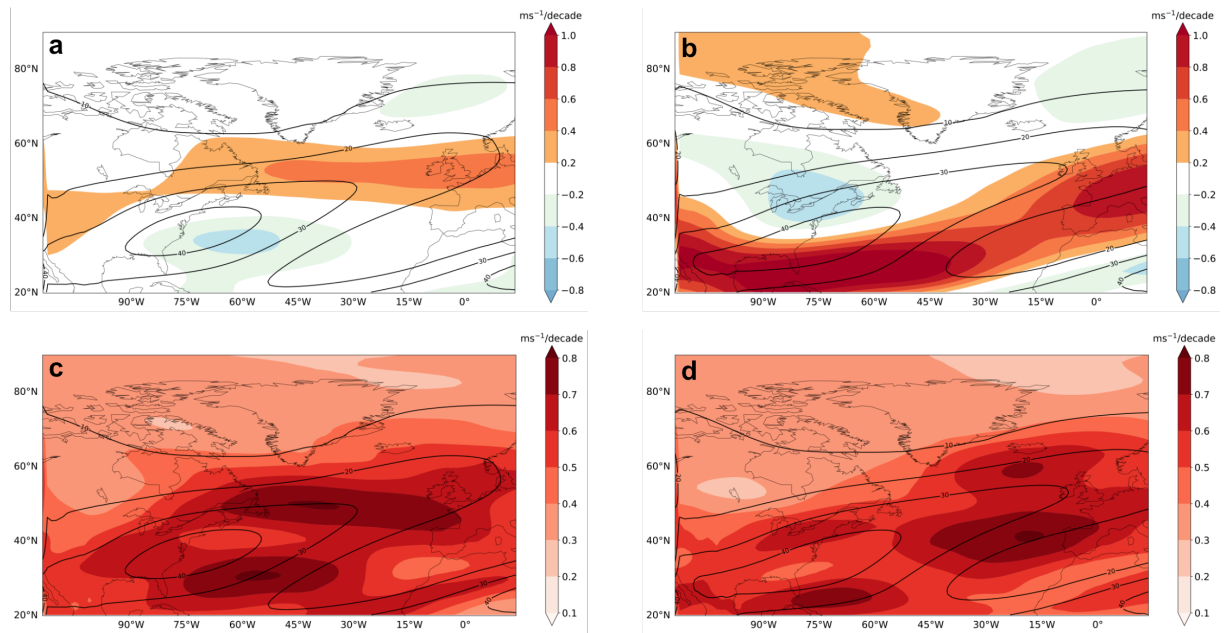
981 **Zonal wind trends in the CESM2 large ensemble (CESM-LENS)**

982 The ensemble mean of the selected members of the CESM2 large ensemble (CESM-LENS)
983 for the period 1980-2022 shows a poleward shift of the jet over the North Atlantic (Fig. B1a),



975 **FIG. A1. Wintertime zonal wind speed trend over the North Atlantic in ERA5 (1940-1978):** a) Mean of
 976 the trend (shading) and climatological average in zonal wind speed (black contours) over the North Atlantic storm
 977 track region (80°W – 15°W) for DJF in the period 1940-1978. b) Zonal wind trend (shading) and climatological
 978 mean (black contours) at 250 hPa. The stippling represents areas with p-values higher than p^* (see text for
 979 details).

984 in contrast to the small ensemble analyzed in section 5. However, the ensemble spread for the
 985 CESM-LENS is also remarkably large in this period with some members showing a poleward shift
 986 and others an equatorward shift as can be identified in the ensemble spread, which exhibits two
 987 areas of high spread equatorward and poleward of the mean jet (Fig. B1c). However, the trend
 988 displayed by ERA5 in mid and low latitudes is well captured by some ensemble members and
 989 contained within the distribution of the CESM-LENS (Fig. B2).

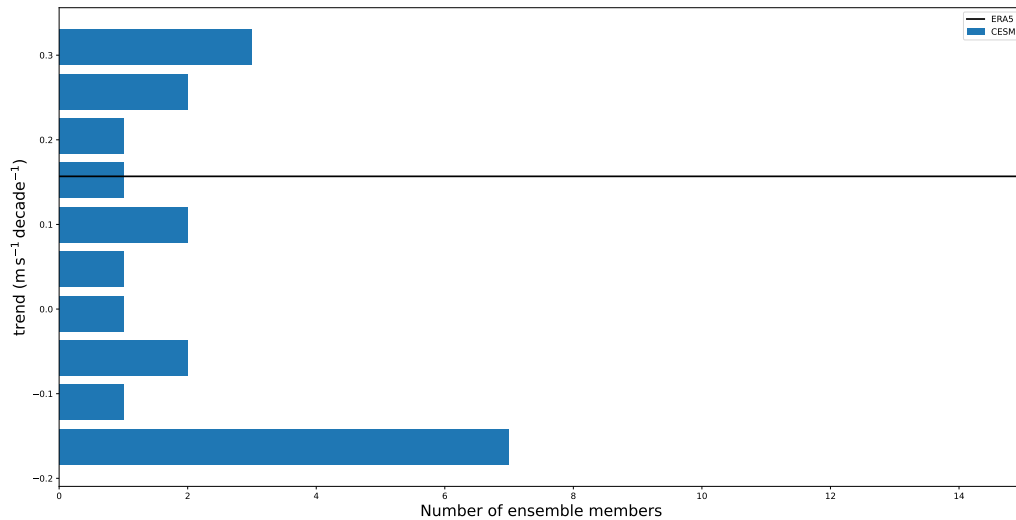


996 **FIG. B1. Mean and spread in zonal wind speed trends for the CESM-LENS:** Mean trend in zonal wind
 997 speed (shading) and climatological mean (black contours) for a) the historical period (1980-2022) and b) the
 998 end of the century period (2057-2100) at 250 hPa. c), d) Standard deviation of the trend in zonal wind speed
 999 (shading) and climatological mean of the zonal wind speed (black contours) for both periods at the same pressure
 1000 level.

990 An inspection of the ensemble mean and spread for the end of the century period (2057-2100)
 991 in the CESM-LENS shows a larger signal to noise ratio in this period (Fig. B1b,d) with most of
 992 the members showing positive trends equatorward of the mean jet over the Gulf Stream and an
 993 extension over Europe. However, there is still some variability in the position of the extension over
 994 Europe and the strength of the positive trend as it is depicted by both the 5-member ensemble and
 995 the CESM-LENS (Figs. 17 and B1d).

1004 **References**

1005 Ambaum, M. H. P., and L. Novak, 2014: A nonlinear oscillator describing storm track variability.
 1006 *Quarterly Journal of the Royal Meteorological Society*, **140**, 2680–2684, [https://doi.org/10.](https://doi.org/10.1002/qj.2352)
 1007 [1002/qj.2352](https://doi.org/10.1002/qj.2352).



1001 **FIG. B2. Distribution of zonal wind trends in the CESM-LENS:** Histogram of zonal wind speed trends
 1002 at 250 hPa for the period 1980-2022 averaged between 80°W – 15°W and 30°N – 50°N for the considered
 1003 CESM-LENS simulations (blue bars) and ERA5 (black line).

1008 Blackport, R., and J. C. Fyfe, 2022: Climate models fail to capture strengthening wintertime
 1009 North Atlantic jet and impacts on Europe. *Science Advances*, **8**, [https://doi.org/10.1126/sciadv.
 1010 abn3112](https://doi.org/10.1126/sciadv.abn3112).

1011 Bogenschutz, P. A., A. Gettelman, H. Morrison, V. E. Larson, C. Craig, and D. P. Schanen, 2018:
 1012 The path to CAM6: coupled simulations with CAM5.4 and CAM5.5. *Geoscientific Model
 1013 Development*, **11**, 235–255, <https://doi.org/10.5194/gmd-11-235-2018>.

1014 Bracegirdle, T. J., H. Lu, R. Eade, and T. Woollings, 2018: Do CMIP5 models reproduce observed
 1015 low-frequency North Atlantic jet variability? *Geophysical Research Letters*, **45**, 7204–7212,
 1016 <https://doi.org/10.1029/2018GL078965>.

1017 Brayshaw, D. J., B. Hoskins, and M. Blackburn, 2009: The Community Earth System Model
 1018 Version 2 (CESM2). *Journal of Advances in Modeling Earth Systems*, **66**, 2539–2558,
 1019 <https://doi.org/10.1175/2009JAS3078.1>.

1020 Brayshaw, D. J., B. Hoskins, and M. Blackburn, 2011: The basic ingredients of the North At-
 1021 lantic storm track. Part II: Sea surface temperatures. *Journal of the Atmospheric Sciences*, **68**,

- 1022 1784–1805, <https://doi.org/10.1175/2011JAS3674.1>.
- 1023 Browning, K. A., M. E. Hardman, T. W. Harrold, and C. W. Pardoe, 1973: The structure of
1024 rainbands within a mid-latitude depression. *Quarterly Journal of the Royal Meteorological*
1025 *Society*, **99**, 215–231, <https://doi.org/10.1002/qj.49709942002>.
- 1026 Charney, J. G., 1947: The dynamics of long waves in a baroclinic westerly current. *Journal*
1027 *of the Atmospheric Sciences*, **4**, 136–162, [https://doi.org/10.1175/1520-0469\(1947\)004<0136:
1028 TDOLWI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1947)004<0136:TDOLWI>2.0.CO;2).
- 1029 Danabasoglu, G., and Coauthors, 2020: The Community Earth System Model Version 2 (CESM2).
1030 *Journal of Advances in Modeling Earth Systems*, **12**, e2019MS001916, [https://doi.org/10.1002/
1031 qj.2378](https://doi.org/10.1002/qj.2378).
- 1032 Doms, G., and Coauthors, 2011: A description of the non-hydrostatic regional COSMO model.
1033 part II: Physical parameterization. Tech. rep., Deutscher Wetterdienst, Offenbach, Germany.
1034 https://doi.org/10.5676/DWD_pub/nwv/cosmo-doc_6.00_II,2011.
- 1035 Drouard, M., G. Rivière, and P. Arbogast, 2013: The north atlantic oscillation response to large-
1036 scale atmospheric anomalies in the northeastern pacific. *Journal of the Atmospheric Sciences*,
1037 **70** (9), 2854–2874, <https://doi.org/10.1175/jas-d-12-0351.1>.
- 1038 Drouard, M., G. Rivière, and P. Arbogast, 2015: The link between the North Pacific climate
1039 variability and the north atlantic oscillation via downstream propagation of synoptic waves.
1040 *Journal of Climate*, **28**, 3957–3976, <https://doi.org/10.1175/JCLI-D-14-00552.1>.
- 1041 Eady, E. T., 1949: The dynamics of long waves in a baroclinic westerly current. *Tellus*, **1**, 33–52,
1042 [https://doi.org/10.1175/1520-0469\(1947\)004<0136:TDOLWI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1947)004<0136:TDOLWI>2.0.CO;2).
- 1043 Eichler, T., and W. Higgins, 2006: Climatology and ENSO-related variability of North
1044 American extratropical cyclone activity. *Journal of Climate*, **19**, 2076–2093, [https://doi.org/
1045 10.1175/JCLI3725.1](https://doi.org/10.1175/JCLI3725.1).
- 1046 Grams, C. M., and Coauthors, 2011: The key role of diabatic processes in modifying the upper-
1047 tropospheric wave guide: a North Atlantic case-study. *Quarterly Journal of the Royal Meteorolo-*
1048 *gical Society*, **137**, 2174–2193, <https://doi.org/10.1002/qj.891>.

- 1049 Green, J. S. A., 1960: A problem in baroclinic stability. *Quarterly Journal of the Royal Meteorological Society*, **86**, 237–251, <https://doi.org/10.1002/qj.49708636813>.
- 1051 Guo, R., C. Deser, and L. Terray, 2019: Human influence on winter precipitation trends
1052 (1921–2015) over North America and Eurasia revealed by dynamical adjustment. *Geophysical Research Letters*, **46**, 3426–3434, <https://doi.org/10.1029/2018GL081316>.
- 1054 Haarsma, R. J., and Coauthors, 2016: High resolution model intercomparison project (HighResMIP
1055 v1.0) for CMIP6. *Geoscientific Model Development*, **9**, 4185–4208, [https://doi.org/10.5194/
1056 gmd-9-4185-2016](https://doi.org/10.5194/gmd-9-4185-2016).
- 1057 Harrold, T. W., 1973: Mechanisms influencing the distribution of precipitation within baro-
1058 clinic disturbances. *Quarterly Journal of the Royal Meteorological Society*, **99**, 232–251,
1059 <https://doi.org/10.1002/qj.49709942003>.
- 1060 Hartmann, D. L., 2007: The atmospheric general circulation and its variability. *Journal of the
1061 Meteorological Society of Japan. Ser. II*, **85B**, 123–143, <https://doi.org/10.2151/jmsj.85B.123>.
- 1062 Harvey, B., J. Methven, C. Sanchez, and A. Schaefer, 2020: Diabatic generation of negative
1063 potential vorticity and its impact on the North Atlantic jet stream. *Quarterly Journal of the
1064 Royal Meteorological Society*, **146**, 1477—1497, <https://doi.org/10.1002/qj.3747>.
- 1065 Heinze-Deml, C., S. Sippel, A. G. Pendergrass, F. Lehner, and N. Meinshausen, 2021: Latent Linear
1066 Adjustment Autoencoder v1.0: a novel method for estimating and emulating dynamic precip-
1067 itation at high resolution. *Geoscientific Model Development*, **14**, 4977–4999, [https://doi.org/
1068 10.5194/gmd-14-4977-2021](https://doi.org/10.5194/gmd-14-4977-2021).
- 1069 Held, I. M., 1993: Large-scale dynamics and global warming. *Bulletin of the American Me-
1070 teorological Society*, **74** (2), 228–242, [https://doi.org/10.1175/1520-0477\(1993\)074<0228:
1071 LSDAGW>2.0.CO;2](https://doi.org/10.1175/1520-0477(1993)074<0228:LSDAGW>2.0.CO;2).
- 1072 Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming.
1073 *Journal of Climate*, **19**, 5686–5699, <https://doi.org/10.1175/JCLI3990.1>.
- 1074 Hersbach, H., and Coauthors, 2020: The ERA5 global reanalysis. *Quarterly Journal of the Royal
1075 Meteorological Society*, **146**, 1999–2049, <https://doi.org/10.1002/qj.3803>.

- 1076 Hersbach, H., and Coauthors, 2023: ERA5 hourly data on single levels from 1940 to present.
1077 Copernicus Climate Change Service (C3S) Climate Data Store (CDS), dataset, [https://doi.org/](https://doi.org/10.24381/cds.adbb2d47)
1078 10.24381/cds.adbb2d47.
- 1079 Hogan, R. J., and A. Bozzo, 2018: A flexible and efficient radiation scheme for the ECMWF
1080 model. *Journal of Advances in Modeling Earth Systems*, **10**, 1990–2008, [https://doi.org/10.](https://doi.org/10.1029/2018MS001364)
1081 1029/2018MS001364.
- 1082 Hoskins, B. J., I. James, and G. White, 1983: The shape, propagation and mean-flow inter-
1083 action of large-scale weather systems. *Journal of the Atmospheric Sciences*, **40**, 1595–1612,
1084 [https://doi.org/10.1175/1520-0469\(1983\)040,1595:TSPAMF.2.0.CO;2](https://doi.org/10.1175/1520-0469(1983)040,1595:TSPAMF.2.0.CO;2).
- 1085 Hoskins, B. J., and D. J. Karoly, 1981: The steady linear response of a spherical atmosphere to ther-
1086 mal and orographic forcing. *Journal of the Atmospheric Sciences*, **38**, 1179–1196, [https://doi.org/](https://doi.org/10.1175/1520-0469(1981)038<1179:TSLROA>2.0.CO;2)
1087 10.1175/1520-0469(1981)038<1179:TSLROA>2.0.CO;2.
- 1088 Hoskins, B. J., and P. J. Valdes, 1990a: On the existence of storm-tracks. *Journal of the*
1089 *Atmospheric Sciences*, **47**, 1854–1864, [https://doi.org/10.1175/1520-0469\(1990\)047<1854:](https://doi.org/10.1175/1520-0469(1990)047<1854:OTEOST>2.0.CO;2)
1090 OTEOST>2.0.CO;2.
- 1091 Hoskins, B. J., and P. J. Valdes, 1990b: On the existence of storm-tracks. *Journal of Atmospheric*
1092 *Sciences*, **47** (15), 1854–1864, [https://doi.org/10.1175/1520-0469\(1990\)047<1854:OTEOST>](https://doi.org/10.1175/1520-0469(1990)047<1854:OTEOST>2.0.CO;2)
1093 2.0.CO;2.
- 1094 Hotta, D., and H. Nakamura, 2011: On the significance of the sensible heat supply from the
1095 ocean in the maintenance of the mean baroclinicity along storm tracks. *Journal of Climate*, **24**,
1096 3377–3401, <https://doi.org/10.1175/2010JCLI3910.1>.
- 1097 Hunke, E., W. Lipscomb, A. Turner, and S. Jeffery, N. ans Elliott, 2015: ICE: The Los Alamos Sea
1098 Ice Model documentation and software user’s manual version 5.1 LA-CC-06-012. Tech. rep.,
1099 Los Alamos National Laboratory, Santa Fe, NM, USA.
- 1100 Igel, A. L., and S. C. van der Heever, 2014: The role of latent heating in warm frontogenesis.
1101 *Quarterly Journal of the Royal Meteorological Society*, **140**, 139–150, [https://doi.org/10.1002/](https://doi.org/10.1002/qj.2118)
1102 qj.2118.

- 1103 Jablonowski, C., and D. L. Williamson, 2006: A baroclinic instability test case for atmospheric
1104 model dynamical cores. *Quarterly Journal of the Royal Meteorological Society*, **132**, 2943–2975,
1105 <https://doi.org/10.1256/qj.06.12>.
- 1106 Kidston, J., A. A. Scaife, S. C. Hardiman, D. M. Mitchell, N. Butchart, B. M. P., and L. J. Gray,
1107 2015: Stratospheric influence on tropospheric jet streams, storm tracks and surface weather.
1108 *Nature Geoscience*, **8**, 433–440, <https://doi.org/10.1038/ngeo2424>.
- 1109 Kuwano-Yoshida, A., S. Minobe, and S.-P. Xie, 2010: Precipitation response to the Gulf Stream in
1110 an atmospheric GCM. *Journal of Climate*, **23**, 3676–3698, <https://doi.org/10.1175/2010jcli3261>.
1111 1.
- 1112 Lawrence, D., and Coauthors, 2019: The Community Land Model version 5: Description of new
1113 features, benchmarking, and impact of forcing uncertainty. *Journal of Advances in Modeling*
1114 *Earth Systems*, **11**, 4245–4287, <https://doi.org/10.1029/2018MS001583>.
- 1115 Lee, J.-Y., and Coauthors, 2021: Future global climate: Scenario-based projections and near-term
1116 information. *Climate Change 2021: The Physical Science Basis. Contribution of Working Group*
1117 *I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*, V. Masson-
1118 Delmotte, P. Zhai, A. Pirani, S. Connors, C. Péan, S. Berger, N. Caud, Y. Chen, L. Goldfarb,
1119 M. Gomis, M. Huang, K. Leitzell, E. Lonnoy, J. Matthews, T. Maycock, T. Waterfield, O. Yelekçi,
1120 R. Yu, and B. Zhou, Eds., Cambridge University Press, Cambridge, United Kingdom and New
1121 York, NY, USA, chap. 4, 553–672.
- 1122 Lee, S., and H. kyung Kim, 2003: The dynamical relationship between subtropical and eddy-
1123 driven jets. *Journal of the Atmospheric Sciences*, **60** (12), 1490–1503, [https://doi.org/10.1175/
1124 1520-0469\(2003\)060<1490:tdrbsa>2.0.co;2](https://doi.org/10.1175/1520-0469(2003)060<1490:tdrbsa>2.0.co;2).
- 1125 Lehner, F., C. Deser, and L. Terray, 2017: Toward a new estimate of “time of emergence” of
1126 anthropogenic warming: Insights from dynamical adjustment and a large initial-condition model
1127 ensemble. *Journal of Climate*, **30**, 7739–7756, <https://doi.org/10.1175/JCLI-D-16-0792.1>.
- 1128 Li, C., and J. J. Wettstein, 2012: Thermally driven and eddy-driven jet variability in reanalysis.
1129 *Journal of Climate*, **25** (5), 1587–1596, <https://doi.org/10.1175/jcli-d-11-00145.1>.

- 1130 Li, H., M. S. Wigmosta, H. Wu, M. Huang, Y. Ke, A. M. Coleman, and L. R. Leung, 2013: A
1131 physically based runoff routing model for land surface and Earth system models. *Journal of*
1132 *Hydrometeorology*, **14**, 808–828, <https://doi.org/10.1175/JHM-D-12-015.1>.
- 1133 Lindzen, R. S., and B. Farrell, 1980: A simple approximate result for the maximum growth rate
1134 of baroclinic instabilities. *Journal of the Atmospheric Sciences*, **37**, 1648–1654, [https://doi.org/10.1175/1520-0469\(1980\)037<1648:ASARFT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<1648:ASARFT>2.0.CO;2).
- 1136 Lorenz, D. J., and E. T. DeWeaver, 2007: Tropopause height and zonal wind response to global
1137 warming in the IPCC scenario integrations. *Journal of Geophysical Research: Atmospheres*,
1138 **112 (D10)**, <https://doi.org/https://doi.org/10.1029/2006JD008087>.
- 1139 Lorenz, D. J., and D. L. Hartmann, 2003: Eddy–zonal flow feedback in the Northern Hemi-
1140 sphere winter. *Journal of climate*, **16 (8)**, 1212–1227, [https://doi.org/10.1175/1520-0442\(2003\)](https://doi.org/10.1175/1520-0442(2003)16<1212:EFFITN>2.0.CO;2)
1141 [16<1212:EFFITN>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)16<1212:EFFITN>2.0.CO;2).
- 1142 Maher, P., M. E. Kelleher, P. G. Sansom, and J. Methven, 2020: Is the subtropical jet shifting
1143 poleward? *Climate Dynamics*, **54**, 1741–1759, <https://doi.org/10.1007/s00382-019-05084-6>.
- 1144 Manney, G. L., and M. I. Hegglin, 2018: Seasonal and regional variations of long-term changes
1145 in upper-tropospheric jets from reanalyses. *Journal of Climate*, **31**, 423–448, <https://doi.org/10.1175/JCLI-D-17-0303.1>.
- 1147 Martin, J. E., 2021: Recent trends in the waviness of the Northern Hemisphere wintertime polar
1148 and subtropical jets. *Journal of Geophysical Research: Atmospheres*, **126**, e2020JD033 668,
1149 <https://doi.org/10.1029/2020JD033668>.
- 1150 Minobe, S., A. Kuwano-Yoshida, N. Komori, S.-P. Xie, and R. J. Small, 2008: Influence of the
1151 Gulf Stream on the troposphere. *Nature*, **452**, <https://doi.org/10.1038/nature06690>.
- 1152 Nakamura, H., 1992: Midwinter suppression of baroclinic wave activity in the Pacific. *Journal of*
1153 *the Atmospheric Sciences*, **49**, 1629–1642, [https://doi.org/10.1175/1520-0469\(1992\)049<1629:](https://doi.org/10.1175/1520-0469(1992)049<1629:MSOBWA>2.0.CO;2)
1154 [MSOBWA>2.0.CO;2](https://doi.org/10.1175/1520-0469(1992)049<1629:MSOBWA>2.0.CO;2).
- 1155 Neale, R. B., and B. J. Hoskins, 2001: A standard test for AGCMs including their physical
1156 parametrizations: I: The proposal. *Atmospheric Science Letters*, **101**, 101–107, <https://doi.org/10.1256/qj.06.12>.
- 1157

- 1158 Novak, L., M. H. P. Ambaum, and R. Tailleux, 2015: The life cycle of the North Atlantic storm track.
1159 *Journal of the Atmospheric Sciences*, **72**, 821–833, <https://doi.org/10.1175/JAS-D-14-0082.1>.
- 1160 O’Neill, B. C., and Coauthors, 2016: The scenario model intercomparison project (ScenarioMIP) for CMIP6. *Geoscientific Model Development*, **9**, 3461–3482, <https://doi.org/10.5194/gmd-9-3461-2016>.
- 1163 O’Reilly, H., S. Minobe, A. Kuwano-Yoshida, and T. Woollings, 2017: The Gulf Stream influence
1164 on wintertime North Atlantic jet variability. *Quarterly Journal of the Royal Meteorological
1165 Society*, **143**, 173–183, <https://doi.org/10.1002/qj.2907>.
- 1166 Orlanski, I., 1998: Poleward deflection of storm tracks. *Journal of the Atmospheric Sciences*,
1167 **55 (16)**, 2577–2602.
- 1168 Orr, A., P. Bechtold, J. Scinocca, M. Ern, and M. Janiskova, 2010: Improved middle atmosphere
1169 climate and forecasts in the ECMWF model through a nonorographic gravity wave drag parameterization.
1170 *Journal of Climate*, **23**, 5905–5926, <https://doi.org/10.1175/2010JCLI3490.1,2010>.
- 1171 Owen, L. E., J. L. Catto, D. B. Stephenson, and N. J. Dunstone, 2021: Compound precipitation
1172 and wind extremes over Europe and their relationship to extratropical cyclones. *Weather and
1173 Climate Extremes*, **33**, 100342, <https://doi.org/10.1016/j.wace.2021.100342>.
- 1174 Papritz, L., and T. Spengler, 2015: Analysis of the slope of isentropic surfaces and its tendencies
1175 over the north atlantic. *Quarterly Journal of the Royal Meteorological Society*, **141 (693)**,
1176 3226–3238.
- 1177 Parfitt, R., and A. Czaja, 2016: On the contribution of synoptic transients to the mean atmospheric
1178 state in the gulf stream region. *Quarterly Journal of the Royal Meteorological Society*, **142 (696)**,
1179 1554–1561, <https://doi.org/10.1002/qj.2689>.
- 1180 Parfitt, R., A. Czaja, and Y.-O. Kwon, 2017: The impact of SST resolution change in the ERA-
1181 interim reanalysis on wintertime Gulf Stream frontal air-sea interaction. *Geophysical Research
1182 Letters*, **44 (7)**, 3246–3254, <https://doi.org/10.1002/2017GL073028>.
- 1183 Peings, Y., J. Cattieux, S. J. Vavrus, and G. Magnusdottir, 2018: Projected squeezing of the
1184 wintertime North-Atlantic jet. *Environmental Research Letters*, **13**, 074016, <https://doi.org/10.1088/1748-9326/aacc79>.
- 1185

- 1186 Peng, S., and J. S. Whitaker, 1999: Mechanisms determining the atmospheric response to
1187 midlatitude SST anomalies. *Journal of Climate*, **12**, 1393–1408, [https://doi.org/10.1175/
1188 1520-0442\(1999\)012<1393:MDTART>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<1393:MDTART>2.0.CO;2).
- 1189 Pfahl, S., and H. Wernli, 2012: Quantifying the relevance of cyclones for precipitation extremes.
1190 *Journal of Climate*, **25** (19), 1288–1297, <https://doi.org/10.1002/qj.829>.
- 1191 Pomroy, H. R., and A. J. Thorpe, 2000: The evolution and dynamical role of reduced upper-
1192 tropospheric potential vorticity in intensive observing period one of FASTEX. *Monthly weather
1193 review*, **128**, 1817–1834, [https://doi.org/10.1175/1520-0493\(2000\)128<1817:TEADRO>2.0.
1194 CO;2](https://doi.org/10.1175/1520-0493(2000)128<1817:TEADRO>2.0.CO;2).
- 1195 Priestley, M. D. K., and J. L. Catto, 2022: Future changes in the extratropical storm tracks and
1196 cyclone intensity, wind speed, and structure. *Weather and Climate Dynamics*, **3**, 337–360,
1197 <https://doi.org/10.5194/wcd-3-337-2022>.
- 1198 Rivière, G., 2009: Effect of latitudinal variations in low-level baroclinicity on eddy life cycles
1199 and upper-tropospheric wave-breaking processes. *Journal of the Atmospheric Sciences*, **66**,
1200 1569–1592.
- 1201 Rivière, G., M. Wimmer, P. Arbogast, J.-M. Piriou, J. Delanoë, C. Labadie, Q. Cazenave, and
1202 J. Pelon, 2021: The impact of deep convection representation in a global atmospheric model on
1203 the warm conveyor belt and jet stream during NAWDEX IOP6. *Weather and Climate Dynamics*,
1204 **2**, 1011–1031, <https://doi.org/10.5194/wcd-2-1011-2021>.
- 1205 Rodgers, S.-S., K. B. Lee, and Coauthors, 2021: Ubiquity of human-induced changes in climate
1206 variability. *Earth System Dynamics*, **12**, 1393–1411, <https://doi.org/10.5194/esd-12-1393-2021>.
- 1207 Sampe, T., H. Nakamura, A. Goto, and W. Ohfuchi, 2010: Significance of a midlatitude SST frontal
1208 zone in the formation of a storm track and an eddy-driven westerly jet. *Journal of Climate*, **23**,
1209 1793–1814, <https://doi.org/10.1175/2009JCLI3163.1>.
- 1210 Schemm, S., 2023: Toward eliminating the decades-old “too zonal and too equatorward”
1211 storm-track bias in climate models. *Journal of Advances in Modeling Earth Systems*, **15** (2),
1212 e2022MS003482, <https://doi.org/https://doi.org/10.1029/2022MS003482>.

- 1213 Schemm, S., L. Papritz, and G. Rivière, 2022: Storm track response to uniform global warming
1214 downstream of an idealized sea surface temperature front. *Weather and Climate Dynamics*,
1215 <https://doi.org/10.5194/wcd-3-601-2022>.
- 1216 Schemm, S., and G. Rivière, 2019: On the efficiency of baroclinic eddy growth and how it
1217 reduces the North Pacific storm-track intensity in midwinter. *Journal of Climate*, **32**, 8373–8398,
1218 <https://doi.org/10.1175/JCLI-D-19-0115.1>.
- 1219 Schemm, S., G. Rivière, L. M. Ciasto, and C. Li, 2018: Extratropical cyclogenesis changes in
1220 connection with tropospheric ENSO teleconnections to the North Atlantic: Role of stationary
1221 and transient waves. *Journal of the Atmospheric Sciences*, **17**, 3943–3964, <https://doi.org/10.1175/JAS-D-17-0340.1>.
- 1222
- 1223 Schemm, S., H. Wernli, and H. Binder, 2021: The storm-track suppression over the western
1224 North Pacific from a cyclone life-cycle perspective. *Weather and Climate Dynamics*, **2**, 55–69,
1225 <https://doi.org/10.5194/wcd-2-55-2021>.
- 1226 Schemm, S., H. Wernli, and L. Papritz, 2013: Warm conveyor belts in idealized moist baroclinic
1227 wave simulations. *Journal of the Atmospheric Sciences*, **70**, 627–652, [https://doi.org/10.1175/](https://doi.org/10.1175/JAS-D-12-0147.1)
1228 [JAS-D-12-0147.1](https://doi.org/10.1175/JAS-D-12-0147.1).
- 1229 Screen, J. A., and I. Simmonds, 2010: The central role of diminishing sea ice in recent Arctic tem-
1230 perature amplification. *Nature*, **464 (7293)**, 1334–1337, <https://doi.org/10.1038/nature09051>.
- 1231 Shaw, T., and Coauthors, 2016: Storm track processes and the opposing influences of climate
1232 change. *Nature Geoscience*, **9 (9)**, 656–664.
- 1233 Sheldon, L., A. Czaja, B. Vanniére, C. Morcrette, B. Sohet, M. Casado, and D. Smith, 2017:
1234 A 'warm path' for Gulf Stream–troposphere interactions. *Tellus A: Dynamic Meteorology and*
1235 *Oceanography*, **69**, 1299–1307, <https://doi.org/10.1080/16000870.2017.1299397>.
- 1236 Simmons, A. J., 2022: Trends in the tropospheric general circulation from 1979 to 2022. *Weather*
1237 *and Climate Dynamics*, **3 (3)**, 777–809, <https://doi.org/10.5194/wcd-3-777-2022>.
- 1238 Simpson, I. R., T. A. Shaw, and R. Seager, 2014: A diagnosis of the seasonally and longitudinally
1239 varying midlatitude circulation response to global warming. *Journal of the Atmospheric Sciences*,
1240 **71**, 2489–2515, <https://doi.org/10.1175/JAS-D-13-0325.1>.

- 1241 Sippel, S., N. Meinshausen, A. Merrifield, F. Lehner, A. G. Pendergrass, E. Fischer, and R. Knutti,
1242 2019: Human influence on winter precipitation trends (1921–2015) over North America and
1243 Eurasia revealed by dynamical adjustment. *Journal of Climate*, **32**, 5677–5699, [https://doi.org/](https://doi.org/10.1175/JCLI-D-18-0882.1)
1244 [10.1175/JCLI-D-18-0882.1](https://doi.org/10.1175/JCLI-D-18-0882.1).
- 1245 Small, R. J., R. A. Tomas, and F. O. Bryan, 2014: Storm track response to ocean fronts in a
1246 global high-resolution climate model. *Climate dynamics*, **43**, 805–828, [https://doi.org/10.1007/](https://doi.org/10.1007/s00382-013-1980-9)
1247 [s00382-013-1980-9](https://doi.org/10.1007/s00382-013-1980-9).
- 1248 Smoliak, B. V., J. M. Wallace, P. Lin, and Q. Fu, 2015: Dynamical adjustment of the Northern
1249 Hemisphere surface air temperature field: Methodology and application to observations. *Journal*
1250 *of Climate*, **28**, 1613–1629, <https://doi.org/10.1175/JCLI-D-14-00111.1>.
- 1251 Stoelinga, M. T., 1996: A potential vorticity-based study of the role of diabatic heating and
1252 friction in a numerically simulated baroclinic cyclone. *Monthly weather review*, **124**, 849–874,
1253 [https://doi.org/10.1175/1520-0493\(1996\)124<0849:APVBSO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1996)124<0849:APVBSO>2.0.CO;2).
- 1254 Thompson, D. W. J., and T. Birner, 2012: On the linkages between the tropospheric isentropic
1255 slope and eddy fluxes of heat during northern hemisphere winter. *Journal of the Atmospheric*
1256 *Sciences*, **69**, 1811–1823, <https://doi.org/10.1175/JAS-D-11-0187.1>.
- 1257 Tiedke, M., 1989: A comprehensive mass flux scheme for cumulus parameterization in large-scale
1258 models. *Monthly Weather Review*, **117**, 1779–1800, [https://doi.org/10.1175/1520-0493\(1989\)](https://doi.org/10.1175/1520-0493(1989)117<1779:ACMFSF>2.0.CO;2)
1259 [117<1779:ACMFSF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1989)117<1779:ACMFSF>2.0.CO;2).
- 1260 Vallis, G. K., 2017: *Atmospheric and Oceanic Fluid Dynamics: Fundamentals and Large-Scale*
1261 *Circulation*. 2nd ed., Cambridge University Press, <https://doi.org/10.1017/9781107588417>.
- 1262 Van Delden, A., 1999: The slope of isentropes constituting a frontal zone. *Tellus A: Dynamic*
1263 *Meteorology and Oceanography*, **51**, 603–611, <https://doi.org/10.3402/tellusa.v51i5.14479>.
- 1264 Wallace, C., and M. Joshi, 2018: Comparison of land–ocean warming ratios in updated ob-
1265 served records and CMIP5 climate models. *Environmental Research Letters*, **13**, 114011,
1266 <https://doi.org/10.1088/1748-9326/aae46f>.

- 1267 Weijenborg, C., and T. Spengler, 2020: Diabatic heating as a pathway for cyclone clustering
1268 encompassing the extreme storm Dagmar. *Geophysical Research Letters*, **47**, e2019GL085777,
1269 <https://doi.org/10.1029/2019GL085777>.
- 1270 Wernli, H., and H. C. Davies, 1997: A lagrangian-based analysis of extratropical cyclones. I: The
1271 method and some applications. *Quarterly Journal of the Royal Meteorological Society*, **123**,
1272 467–489, <https://doi.org/10.1002/qj.49712353811>.
- 1273 Wilks, D. S., 2016: The stippling shows statistically significant grid points. *Bulletin of the American*
1274 *Meteorological Society*, **97**, 2263–2273, <https://doi.org/10.1175/BAMS-D-15-00267.1>.
- 1275 Wimmer, M., G. Rivière, P. Arbogast, J.-M. Piriou, J. Delanoë, C. Labadie, Q. Cazenave, and
1276 J. Pelon, 2022: Diabatic processes modulating the vertical structure of the jet stream above the
1277 cold front of an extratropical cyclone: sensitivity to deep convection schemes. *Weather and*
1278 *Climate Dynamics*, **3**, 863–882, <https://doi.org/10.5194/wcd-3-863-2022>.
- 1279 Woollings, T., 2010: Dynamical influences on european climate: an uncertain future. *Philosoph-*
1280 *ical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*,
1281 **368 (1924)**, 3733–3756, <https://doi.org/10.1098/rsta.2010.0040>.
- 1282 Woollings, T., M. Drouard, D. M. H. O'Reilly, C. H. Sexton, and C. McSweeney, 2023: Trends in
1283 the atmospheric jet streams are emerging in observations and could be linked to tropical warming.
1284 *Communications Earth and Environment*, **4**, 1–8, <https://doi.org/10.1038/s43247-023-00792-8>.
- 1285 Woollings, T., J. M. Gregory, J. G. Pinto, M. Reyers, and D. J. Brayshaw, 2012: Response of
1286 the North Atlantic storm track to climate change shaped by ocean–atmosphere coupling. *Nature*
1287 *Geoscience*, **5 (5)**, 313–317.
- 1288 Woollings, T., L. Papritz, C. Mbengue, and T. Spengler, 2016: Diabatic heating and jet stream
1289 shifts: A case study of the 2010 negative North Atlantic Oscillation winter. *Geophysical Research*
1290 *Letters*, **43**, 9994–10002, <https://doi.org/10.1002/2016GL070146>.
- 1291 Yadav, P., and D. M. Straus, 2017: Circulation response to fast and slow MJO episodes. *Monthly*
1292 *Weather Review*, **145**, 1577–1596, <https://doi.org/10.1175/MWR-D-16-0352.1>.
- 1293 Yu, B., and H. Lin, 2016: Tropical atmospheric forcing of the wintertime North Atlantic Oscillation.
1294 *Journal of Climate*, **29**, 1755–1772, <https://doi.org/10.1175/JCLI-D-15-0583.1>.

1295 Yuval, J., and Y. Kaspi, 2016: Eddy activity sensitivity to changes in the vertical structure of
1296 baroclinicity. *Journal of the Atmospheric Sciences*, **73**, 1709–1726, [https://doi.org/10.1175/
1297 JAS-D-15-0128.1](https://doi.org/10.1175/JAS-D-15-0128.1).

1298 Zängl, G., D. Reinert, P. Rípodas, and M. Baldauf, 2015: The ICON (ICOsahedral Non-hydrostatic)
1299 modelling framework of DWD and MPI-M: Description of the non-hydrostatic dynamical core.
1300 *Journal of the Atmospheric Sciences*, **141**, 563–579, <https://doi.org/10.1002/qj.2378>.