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3 **Increased ocean heat transport into the Nordic Seas and Arctic Ocean**
4 **over the period 1993-2016**
5

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15
16 Warm water of subtropical-origin flows northward in the Atlantic Ocean and transports heat to
17 high latitudes. This poleward heat transport has been implicated as one possible cause of the
18 declining sea ice extent and increasing ocean temperatures across the Nordic Seas and Arctic Ocean,
19 but robust estimates are still lacking. Here we use a box inverse model and over 20 years of volume
20 transport measurements to show that the mean ocean heat transport was 305 ± 26 TW for 1993-2016.
21 A significant increase of 21 TW occurred after 2001, which is sufficient to account for the recent
22 accumulation of heat in the northern seas. Therefore, ocean heat transport may have been a major
23 contributor to climate change since the late 1990s. This increased heat transport contrasts with the
24 Atlantic Meridional Overturning Circulation (AMOC) slowdown at mid-latitudes and indicates a
25 discontinuity of the overturning circulation measured at different latitudes in the Atlantic Ocean.
26
27

28 The AMOC is the large-scale, bidirectional circulation in the Atlantic Ocean that transports warm
29 subtropical-origin Atlantic water (AW) northward near the surface and cold, dense water southward at
30 depth, connected by warm-to-cold water mass transformation at high latitudes¹. The bulk of the AW enters
31 the Norwegian Sea between Iceland and Scotland (Fig. 1a). The volume and temperature of the AW
32 transported into the Nordic Seas exert a strong influence on climate in northern Europe^{2,3}. As the warm
33 water progresses northward toward the Arctic Ocean, heat is continually released to the atmosphere and
34 the AW becomes colder and denser. Before reaching Fram Strait between Greenland and the Svalbard
35 archipelago the AW has attained sufficient density to supply the dense overflow water (OW) plumes that
36 pass through gaps in the Greenland-Scotland Ridge (GSR) and form the headwaters to the lower limb of
37 the AMOC⁴. Some additional transformation also takes place in the Barents Sea⁵ and Arctic Ocean⁶.
38 Recent observational programs emphasize the importance of this warm-to-cold water mass transformation
39 in the Nordic Seas and Arctic Ocean (jointly referred to as the Arctic Mediterranean) to the AMOC^{7,8}.

40
41 The Arctic Mediterranean has accumulated heat over the last decades as manifested by increased
42 ocean heat content and reduced sea ice volume⁹⁻¹². The Norwegian Sea experienced warming at a rate of

43 3.2 W m⁻² during 1995-2010⁹ and approximately 7 W m⁻² during 2011-2018¹⁰. Long-term hydrographic
44 monitoring has documented a widespread increase in upper ocean temperature along AW pathways from
45 the Rockall Trough to Fram Strait since 2000¹³⁻¹⁵. At higher latitudes the Eurasian Basin of the Arctic
46 Ocean and in particular the northern Barents Sea are going through a marine climate transition referred to
47 as “Atlantification”, which is associated with enhanced surface presence of AW, reduced sea ice cover, a
48 weakened halocline, and deeper wintertime convection¹⁶⁻¹⁸. Despite a dearth of hydrographic observations
49 in the Arctic Ocean, recent studies relying on data assimilation demonstrated that north of 70°N heat has
50 accumulated at a rate of approximately 1 W m⁻² (equivalent to ~12 TW over the entire region north of
51 70°N) since 2000^{11,12}.

52 Ocean currents through key gateways into the Arctic Mediterranean have been measured since the
53 late 1990s¹⁹. A recent compilation of volume transport measurements demonstrates that the exchange flow
54 across the GSR has remained constant or slightly increased (the increase is not statistically significant)²⁰.
55 This indicates that the strength of the northern component of the AMOC has been stable over the past two
56 decades. By contrast, measurements from a mooring array at 26.5°N in the North Atlantic indicate that the
57 AMOC has been in a reduced state since 2008²¹. This disconnect between a stable AMOC across the GSR
58 and a weakened AMOC at 26.5°N is presently not well understood^{7,22-24}.

59 Even though long-term volume transport measurements exist, robust estimates of ocean heat
60 transport into the Arctic Mediterranean have yet to be determined. The ocean heat transport is a product of
61 the temperature difference between compensating inflows and outflows and their volume transports. The
62 greatest impediment for such a calculation is the requirement of a mass-balanced ocean circulation across
63 the boundary of the Arctic Mediterranean²⁵. Although published long-term observation-based volume
64 transports conserve mass within uncertainties²⁰, observational time series can never fulfill the requirement
65 of complete mass balance. From this we can only calculate approximate heat transport across the GSR of
66 around 300 TW with no information about its temporal variability^{26,27}. Recently, however, such mass-
67 balanced heat transport estimates across the gateways to the Arctic Ocean were obtained using an inverse
68 box model to constrain the hydrographic and velocity measurements²⁸⁻³⁰. Here we extend this calculation
69 to include the Nordic Seas using the same method and the longer measurement time series across the GSR
70 to quantify the ocean heat transport into the Arctic Mediterranean over the period 1993-2016.

71

72 **Mass-balanced ocean circulation**

73 In order to quantify the ocean circulation across the gateways of the Arctic Mediterranean, we
74 consider published volume transport time series of 11 major ocean currents (Fig. 1). Following previous
75 work^{20,26,27}, the currents are divided into three distinct water mass groups: AW, OW, and polar water
76 (PW). AW is the warm and saline subtropical-origin water that flows northward across the GSR. The AW
77 inflow into the Arctic Mediterranean takes place primarily within the Iceland-Faroe (IF)¹⁵ and the Faroe-
78 Shetland Channel (FSC)³¹ branches between Iceland and Scotland, with some contributions also from the
79 North Icelandic Irminger Current (NIIC)³² west of Iceland and the European Shelf (ES)²⁰ branch which is a
80 combination of flows over the Scottish shelf and through the English Channel. OW is the cold and dense
81 water, commonly defined by a potential density greater than $\sigma_\theta = 27.8 \text{ kg m}^{-3}$, that returns southward
82 through gaps in the GSR as dense plumes. The major OW plumes pass through Denmark Strait (DS)³³ and
83 Faroe Bank Channel (FBC)²², with a minor contribution from Wyville Thomson Ridge (WTR)³⁴. The
84 dense water transport across the Iceland-Faroe Ridge is likely negligible³⁵ and has been disregarded. PW
85 is a cold and fresh water mass primarily originating in the Arctic Ocean that flows southward on both sides
86 of Greenland, through Davis Strait as the Baffin Island Current (here referred to as Davis Strait west)³⁶ and
87 through Denmark Strait as the light portion of the East Greenland Current (here referred to as EGC
88 shelf)³⁷. Within the PW category, water is also imported into the Arctic Mediterranean through Bering
89 Strait³⁸ and along the eastern side of Davis Strait (here referred to as Davis Strait east)³⁶. While these

90 inflows into the Arctic Mediterranean are not as cold and fresh as the PW emerging directly from the
91 Arctic Ocean, they have densities below $\sigma_\theta = 27.7 \text{ kg m}^{-3}$ and belong within the PW category³⁹.

92 The time series of volume transport were combined using a box inverse model, a widely used
93 method to quantify large-scale ocean circulation originally developed to resolve the classical
94 oceanographic problem of reference level velocity⁴⁰. We applied mass conservation as a constraint to
95 obtain monthly mass-balanced ocean transports across the boundary of the Arctic Mediterranean
96 (Supplementary Fig. 1). Each current was prescribed with an a priori uncertainty (Supplementary Table 1).
97 At each time step the inverse calculation then optimally adjusted the transport of each current within the
98 uncertainty bounds to obtain mass balance. Most of the modifications (58%) were applied to the PW,
99 which has a larger uncertainty, in particular the sparsely sampled EGC shelf (Fig. 1c). The modifications
100 of the volume transport are expressed primarily in the month-to-month variability and in the seasonal
101 cycle, while the long-term mean value is largely unchanged (Supplementary Fig. 2). The validity of our
102 23-year inversion was assessed by comparing the entire 1993-2016 period (Table 1) to the well-sampled
103 period of 2005-2012 (Supplementary Table 2). The statistics of the two periods are in excellent agreement,
104 which lends confidence to our long-term inversion. Details of the inverse calculation and time series
105 preparation are provided in the methods section.

106 The mass-balanced circulation obtained from the inversion is in good agreement with the in-situ
107 measurements in terms of means and variability on seasonal and longer time scales (Table 1; Fig. 2)²⁰.
108 Over the entire 1993-2016 period the AW and OW volume transports were stable apart from an increase of
109 approximately 0.5 Sv (1 Sv = 10^6 m s^{-1}) between 1998 and 2002 (Supplementary Fig. 3). The AW volume
110 transport is the highest in winter and the lowest in summer with a seasonal range of about 1.5 Sv, while the
111 OW volume transport does not have a clear seasonality (Fig. 2b).

113 **Ocean heat transport**

114 Combining the mass-balanced circulation and the transport-weighted temperatures for each
115 current, we have robustly quantified the ocean heat transport into the Arctic Mediterranean (Table 1; Fig.
116 3a). The monthly ocean heat transport time series has a large variability ranging from 200 to 450 TW, with
117 a long-term mean of 305 ± 26 TW. Details of the heat transport calculation and uncertainty estimates are
118 provided in the methods section. The heat transport has a distinct seasonal cycle, with maximum in
119 September-December and minimum in April-June, and an amplitude of about 120 TW (Fig. 3b). The
120 seasonality is the combined result of a temperature-driven component (with maximum in July-November)
121 and a velocity-driven component (which peaks in October-January; Supplementary Fig. 4). Considering
122 only the temperature transport across the GSR, we obtain a value of 281 ± 24 TW-eq. (Mass conservation is
123 not ensured when considering individual branches of the circulation, hence we use a different terminology
124 and unit, such as temperature transport and TW-eq^{29,41}). The temperature transport is in good agreement
125 with a recent estimate of 273 ± 27 TW based on repeat shipboard velocity and temperature measurements
126 along the GSR during 2009-2016⁸.

127 A key feature of the ocean heat transport time series is the substantial increase that took place
128 during 1998-2002, which is particularly evident in the filtered time series (Fig. 3a). A change point
129 detection method⁴² identified a discontinuity in the monthly ocean heat transport time series in July 2001 at
130 the 99.9% confidence level. Compared to the 1993-2000 average, an additional 21 TW of heat was
131 imported during 2002-2016, which is equivalent to a 1.5 W m^{-2} extra heat flux across the entire Arctic
132 Mediterranean. The difference in ocean heat transport between the two periods is statistically significant at
133 the 99.9% confidence level estimated using Welch's t-test as well as a bootstrap method⁴³. To understand
134 the cause of the increase, we decomposed the ocean heat transport time series into each water mass group
135 (Fig. 4a). The increase stems in full from the AW contribution. Further decomposition into the different
136 AW branches reveals that all three currents contributed to the increase at different times (Fig. 4b). We do

137 not consider the ES branch here due to its low transport and the lack of long-term measurements. The
138 temperature transport of the FSC increased first during 1998-2002, followed by the NIIC around 2003 and
139 the IF during 2004-2007. Relative to 1997, all of these three branches had 6-7 TW-eq higher temperature
140 transports in 2012-2014. Although such a decomposition into different water mass groups and currents no
141 longer ensures mass conservation, the same conclusion can be drawn using a constant reference
142 temperature near the freezing point of -1.8°C (Supplementary Fig. 5). While the increased temperature
143 transports in the NIIC³² and the IF¹⁵ have been documented, the present study quantifies their relative
144 contributions to the total ocean heat transport for the first time. Further decomposition sheds light on the
145 dominant causes of the heat transport variability (Fig. 4c). Both temperature- and velocity-driven
146 components contributed to the increased heat transport. The velocity-driven component increased first and
147 is a reflection of increased AW and OW flow across the GSR during 1998-2002 (Supplementary Fig. 3),
148 followed by an increase in the temperature-driven component during 2000-2004.

149 150 **OVERTURNING IN THE NORDIC SEAS**

151 Using a similar box inverse technique the ocean and sea ice heat transports across the gateways of
152 the Arctic Ocean for the period 2004-2010 were recently quantified as 180 ± 23 TW^{29,30}. If we consider the
153 same period of time and subtract this value from the heat transports into the entire Arctic Mediterranean,
154 we obtain 137 ± 34 TW for the Nordic Seas alone (Fig. 5). This is in good agreement with a comprehensive
155 heat budget in the Nordic Seas of 124 TW based on hydrographic measurements during 1990-1999⁴⁴.
156 Ocean heat transport into the Barents Sea was estimated to 73 TW for the period 1997-2007⁴⁵.

157 The water mass transformation from AW to OW that happens in the Arctic Mediterranean
158 requires 189 ± 14 TW of heat loss considering a temperature difference of 8.4°C and an OW volume
159 transport of 5.5 ± 0.3 Sv. If we assume that all of the heat loss from the Nordic Seas is used to form OW, the
160 137 ± 34 TW of ocean heat transport sustains an overturning circulation in the Nordic Seas of 4.0 ± 0.7 Sv.
161 This implies that the remainder of the overturning, 1.5 ± 0.8 Sv, takes place in the Arctic Ocean, primarily
162 in the Barents Sea⁵. This estimate agrees well with other inverse calculations which estimate a 1.3 ± 0.7 Sv
163 production of Arctic Intermediate water in the Barents Sea⁴⁶ and a net outflow of 1.5 Sv through Fram
164 Strait²⁹, most of which contributes to the Denmark Strait OW plume^{39,47}. These results further emphasize
165 the importance of water mass transformation in the Nordic Seas as a source of dense water to the lower
166 limb of the AMOC^{7,8}.

167 168 **WIDER IMPLICATIONS**

169 Since 2000, heat has accumulated at a rate of approximately 1 W m^{-2} or 12 TW in total in the
170 ocean north of 70°N which includes about half of the Nordic Seas and the entire Arctic Ocean^{11,12}. While
171 this warming is reflected in substantial sea ice loss and a more pronounced Atlantic Water influence at
172 high latitudes^{11,12,16-18}, the heat content of the Norwegian Sea has increased at an even higher rate^{9,10}.
173 Recent numerical and observational studies demonstrate that both surface heat flux and ocean heat
174 transport into the Norwegian Sea regulate the heat content variability on interannual to decadal time
175 scales^{10,48,49}. The increase in ocean heat transport across the GSR of 21 TW that took place after 2001 is
176 sufficient to account for most, if not all, of the heat accumulation in the Arctic Mediterranean. This
177 emphasizes the crucial role of ocean heat transport in the high latitudes climate system.

178 The AMOC transports warm water northward in the Atlantic Ocean and is a crucial component of
179 the climate system¹. In a warming climate the AMOC is expected to weaken⁵⁰. Measurements from a
180 mooring array at 26.5°N indicate that the AMOC has been in a reduced state since 2008²¹. By contrast, the
181 exchange flow of AW and OW across the GSR has been stable or slightly increased since 1992^{20,22,33}, and
182 our results demonstrate that also the ocean heat transport across the ridge has increased. This apparent

183 disconnect between a stable overturning at high latitudes and a weakened circulation at mid-latitudes in the
184 Atlantic Ocean is not well understood, but of key importance for accurate predictions of future climate
185 states^{7,22-24}.

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189

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311 **Data availability**

312 The mass-balanced ocean volume and heat transports are available at the Norwegian Marine Data Center
313 repository at <http://metadata.nmdc.no/metadata-api/landingpage/0a2ae0e42ef7af767a920811e83784b1>. The
314 volume transport time series for GSR branches (IF, FSC, NIIC, DS, FBC) are available at Oceansites
315 website (<http://www.oceansites.org/tma/gsr.html>). The WTR data is available through
316 <https://www.bodc.ac.uk>. The objectively mapped sections in Davis Strait are available via
317 <http://iop.apl.washington.edu/data.html>. The Bering Strait data is available at project website
318 (<http://psc.apl.washington.edu/HLD/Bstrait/bstrait.html>). The Arctic Ocean heat transport estimates during
319 2004-2010 is available at <https://doi.org/10.1594/PANGAEA.909966>. The ERA-Interim reanalysis data
320 were obtained from European Centre for Medium-Range Weather Forecasts
321 (<https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era-interim>). The PIOMAS were obtained
322 from the Polar Science Centre at University of Washington (<http://psc.apl.uw.edu/research/projects/arctic-sea-ice-volume-anomaly/>).
323

324

325 **Endnotes**

326 Correspondence and requests for materials should be addressed to T.T.

327

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342

343 **Author contributions**

344 T.T. and K.V. conceived and developed the study. T.T. integrated and analyzed the data with inputs from
345 all authors. T.T. and K.V. wrote the paper and all authors interpreted the results and clarified the
346 implications.

347

348 **Competing interests**

349 The authors declare no competing interests.

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354 Table 1: Means, uncertainties, and standard deviations (std.) of volume transports (Sv), heat transports
 355 (TW), and temperature transports (TW-eq) based on the monthly mass-balanced ocean circulation between
 356 January 1993 and April 2017. The transport estimates are categorized into water mass groups, net
 357 boundary transports, and AW components. Heat transports are printed in bold and are independent of a
 358 reference temperature. Temperature transports are calculated with reference to 0.0°C. The acronyms are:
 359 PW = Polar Water; AW= Atlantic Water; OW = Overflow Water; NIIC = North Icelandic Irminger
 360 Current; IF = Iceland-Faroe; FSC = Faroe-Shetland Channel; ES = European Shelf.
 361

Volume transports (Sv)		Heat / temperature transports (TW / TW-eq)	
Mean ± uncertainty	± std.	Mean ± uncertainty	± std.
Water mass groups			
PW	-2.7±1.3	±0.8	28±8
AW	8.0±0.7	±1.0	273±24
OW	-5.6±0.4	±0.5	4±0.4
Net boundary transports			
Ocean	-0.2±1.5	±0.2	305±26
Sea ice	-0.01±0.02	±0.02	5±0.1
Ocean plus sea ice	-0.2±1.5	±0.2	310±26
AW components			
NIIC	0.9±0.1	±0.3	24±3
IF	3.8±0.4	±0.5	124±13
FSC	2.7±0.4	±1.0	100±15
ES	0.6±0.3	±0.1	25±14

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365 **Figure legends**

366

367 **Figure 1: Major ocean currents of the region and data coverage.** (a) Schematic diagram of the major
368 currents crossing the boundary of the Arctic Mediterranean. The currents are color-coded according to
369 their water mass category: cyan for PW, red for AW, and black for OW. Only the currents included in the
370 inverse calculation are labeled. The size of the arrows reflect the magnitude of the volume transports. The
371 acronyms are: NIIC = North Icelandic Irminger Current; IF = Iceland-Faroe; FSC = Faroe-Shetland
372 Channel; ES = European Shelf; DS = Denmark Strait; FBC = Faroe Bank Channel; WTR = Wyville
373 Thomson Ridge. (b) The number of time series included in the inversion. (c) Data coverage of each current
374 included in the inversion.

375

376 **Figure 2: Mass-balanced volume transport across the boundary of the Arctic Mediterranean.** (a)
377 Monthly time series between January 1993 and April 2017. The time series are color-coded according to
378 their water mass category: cyan for PW, red for AW, blue for OW, and black for the sum. (b) Their
379 seasonalities and standard deviations.

380

381 **Figure 3: Mass-balanced ocean heat transport across the boundary of the Arctic Mediterranean.** (a)
382 Monthly time series between January 1993 and April 2017. The thick black line is a filtered time series
383 using a 61-month Hanning filter. The horizontal lines represent the means of the two periods (1993-2000
384 and 2002-2016) defined by the change point analysis. (b) Its seasonality and standard deviation.

385

386 **Figure 4: Filtered ocean heat transport change referenced to January 1997.** (a) Changes in ocean heat
387 transports (TW) in black and temperature transports of the distinct water masses (TW-eq) in colors. The
388 shading represents uncertainty estimates. The time series are filtered using a 61-month Hanning filter. (b)
389 Contributions to the heat transport change by the three AW branches (NIIC in blue, IF in cyan, and FSC in
390 orange). (c) Dominant causes of the heat transport change (temperature-driven in orange, velocity-driven
391 in blue, and the correlation term in cyan).

392

393 **Figure 5: Summary of ocean and sea ice heat transport estimates in the North Atlantic and Arctic**
394 **Mediterranean.** The ocean and sea ice heat transports into the Arctic Mediterranean during 1993-2016
395 and the Nordic Seas during 2004-2010 obtained in this study are printed in the rectangular red boxes. The
396 black ellipses show mass-balanced ocean and sea ice heat transports across a section (OSNAP, GSR) or
397 into enclosed regions (Nordic Seas, Barents Sea, Arctic Ocean) from the recent literature for comparison
398 and context.

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400

401

1

2 **Methods**3 **Data**

4 In order to derive a mass-balanced ocean circulation across the boundary of the Arctic
 5 Mediterranean, we consider the full mass budget of the Arctic Mediterranean. The mass budget can be
 6 expressed in the form:

7
$$\frac{\partial M}{\partial t} = F_{ma}^o + F_{ma}^i + F_{ma}^{surf} \quad (\text{Eq. 1})$$

8 where $\frac{\partial M}{\partial t}$ is the change in interior mass storage with time (storage flux), F_{ma}^o and F_{ma}^i are the horizontal
 9 exchanges of mass due to ocean circulation and sea ice export across the side boundary, and F_{ma}^{surf} is the
 10 addition or removal of mass due to surface freshwater fluxes. These freshwater fluxes are net precipitation
 11 minus evaporation (P-E), river runoff, Greenland ice sheet and land ice melt. We consider $\frac{\partial M}{\partial t}$ and F_{ma}^{surf}
 12 together in the inversion because this combined term is balanced by horizontal exchange of mass due to
 13 ocean circulation and sea ice export.

14 For F_{ma}^o , we consider published volume transport records of the exchanges across the boundary of
 15 the Arctic Mediterranean that cover different time periods between 1993 and 2016 (Fig. 1c). They are
 16 categorized into three distinct water mass groups: PW, AW and OW. For the PW we use observations
 17 from Bering Strait for the period 1992-2016³⁸ and from the east Greenland shelf north of Denmark Strait
 18 (EGC shelf) during 2011-2012³⁷. For the AW we use transport estimates of IF¹⁵ and FSC³¹ branches
 19 during 1993-2017 and NIIC during 1994-2015³². For the OW we use observations from DS³³ and FBC²²
 20 during 1996-2016, and WTR during 2006-2009 and 2011-2013³⁴. Some of the minor inflows and outflows
 21 lack sustained observations. For the ES branch we prescribe a constant AW inflow of 0.6 Sv with a
 22 seasonal cycle of amplitude 0.1 Sv²⁰. In Davis Strait we distinguish the southward Baffin Island Current on
 23 the western side and the northward West Greenland and West Greenland Slope Currents on the eastern
 24 side³⁶ using a separation longitude of 58°W²⁸ based on monthly velocity fields during 2004-2010³⁰.

25 The F_{ma}^i is quantified using monthly effective sea ice thickness and velocity data from the Pan-
 26 Arctic Ice Ocean Modeling and Assimilation System (PIOMAS) version 2.1 product⁵¹. The effective sea
 27 ice thickness is a product of sea ice thickness and sea ice concentration. The PIOMAS product covers the
 28 entire period of the inversion.

29 The F_{ma}^{surf} is prescribed based on the mean seasonal cycle of the surface freshwater flux estimate to
 30 the Arctic Ocean⁵² and the mean seasonal cycle of P-E to the Nordic Seas based on ERA-Interim⁵³.
 31 Greenland ice sheet and land ice melt contribution⁵⁴ are an order of magnitude smaller than other fresh
 32 water fluxes and have been neglected. The $\frac{\partial M}{\partial t}$ is prescribed with a satellite-derived storage flux in the
 33 central Arctic Ocean during 2003-2014⁵⁵ considering spatially uniform sea level change⁵⁶. Here we use the
 34 spatial average values of F_{ma}^{surf} and $\frac{\partial M}{\partial t}$ across the Arctic Mediterranean and consider only its mean seasonal
 35 cycle. We do not take interannual variability into account since the uncertainty of F_{ma}^{surf} is high⁵⁷. The
 36 combined surface freshwater flux and storage flux ($F_{ma}^{surf} - \frac{\partial M}{\partial t}$) has an annual mean value of 196 mSv with
 37 a pronounced seasonal cycle which peaks in June at approximately 650 mSv, and it is reduced to 80-250
 38 mSv during the rest of the year. A sensitivity test of the inverse model with constant ($F_{ma}^{surf} - \frac{\partial M}{\partial t}$) over the
 39 entire period confirms that temporal variability of this term has negligible impact (less than 0.1%) on the
 40 resulting mass-balanced ocean circulation and ocean heat transport.

41

42 **Short time series and data gaps**

43 Our inverse calculation extends from January 1993 to April 2017, coincident with volume
 44 transport records from the two major AW inflow branches (IF and FSC; Fig. 1c). These are the most
 45 critical measurements, as the bulk of the ocean heat transport variability originates in these AW branches
 46 (Table 1 and Fig. 4b). While measurements of the major AW and OW branches cover the period 1996-
 47 2016, some of the ocean transport records are shorter than the period of inversion, in particular some of the
 48 PW branches. The records occasionally also contain data gaps. These have various origins, such as data
 49 outliers or missing instruments. To obtain continuous monthly transport time series over the entire period,
 50 gaps shorter than 3 months duration are filled by linear interpolation, whilst gaps exceeding 3 months are
 51 filled using the mean seasonal cycle from the record. Short time series were extended using the average
 52 value of the record modulated by its mean seasonal cycle. These extended time series are assigned a
 53 greater uncertainty as described in the following section.

54

55 **Box inverse model settings**

56 We apply four mass constraints to the box inverse model, which arise from conserving mass in
 57 each of the three water mass groups (PW, AW, OW) as well as the total mass. The model considers four
 58 different mechanisms to satisfy the full mass constraint (Eq. 1). These are horizontal exchange of water
 59 due to ocean circulation (F_{ma}^o) and sea ice (F_{ma}^i), combined effect of net surface freshwater fluxes minus
 60 interior mass storage change ($F_{ma}^{surf} - \frac{\partial M}{\partial t}$). Diapycnal exchanges of water from one water mass to another
 61 within the interior of the Arctic Mediterranean are also considered to satisfy the mass constraints for the
 62 three water mass groups. In practice, the four mass constraints are expressed by an equation,

63 $\mathbf{Ab}^u = \mathbf{d} + \boldsymbol{\varepsilon},$ (Eq. 2)

64 where ($m \times n$) matrix \mathbf{A} contains information about the geometry of the system (m is the total number of
 65 constraints and n is the total number of unknowns). \mathbf{b}^u is an $n \times 1$ vector and contains sets of unknowns
 66 that the inversion solves. The ($m \times 1$) vector \mathbf{d} contains information about the magnitudes of the initial
 67 imbalances for each constraint. The ($m \times 1$) vector $\boldsymbol{\varepsilon}$ is an error vector. Full details of the box inverse model
 68 is available in the Appendix of a paper which performed a similar inversion for the Arctic Ocean²⁹.

69 The four mass constraints ($m=4$) are satisfied by modifying initial estimates of 25 unknown
 70 parameters ($n=25$), which consist of 11 F_{ma}^o velocity parameters and 11 F_{ma}^i velocity parameters for the 11
 71 defined ocean currents, 1 ($F_{ma}^{surf} - \frac{\partial M}{\partial t}$) velocity parameter, and 2 diapycnal velocity parameters between
 72 water mass interfaces below and above the AW. The parameters are initialized as follows. A first guess
 73 for the F_{ma}^o velocity parameters are obtained from the monthly volume transports scaled by the
 74 corresponding cross-sectional areas. The F_{ma}^i velocity parameters are initialized using the PIOMAS sea ice
 75 volume transport time series. The ($F_{ma}^{surf} - \frac{\partial M}{\partial t}$) velocity is initialized from the prescribed repeat seasonal
 76 cycle. The diapycnal velocities are set to zero. Note that salt is not constrained due to the substantial
 77 variability of PW salinity that we do not have sufficient observations to fully resolve. Unlike previous
 78 inverse model studies^{28,29}, we compute only ocean heat transports due to the lack of salt constraints. While
 79 PW temperature measurements are also scarce, the variability in temperature is comparatively small and
 80 sensitivity experiments indicate that this has only a minor impact on our ocean heat transport estimate (2%
 81 at most).

82 Row and column weights are used to prescribe uncertainties for both the constraints and the
 83 unknown parameters. The uncertainties of the constraints are: 2.0 Sv for the PW layer, 1.0 Sv for the AW
 84 layer, 0.5 Sv for the OW layer, and 0.01 Sv for the whole water column mass balance. For the F_{ma}^o
 85 velocities we apply the published uncertainty estimates listed in Supplementary Table 1. At times when
 86 volume transport estimates are not available, the uncertainty is doubled. For the F_{ma}^i and ($F_{ma}^{surf} - \frac{\partial M}{\partial t}$)

87 velocities the uncertainties are set to 30% and 100% of the initial estimates, respectively. For the diapycnal
88 velocity, the uncertainty is set to 10^{-4} m s^{-1} .

89 The uncertainties of the F_{ma}^0 (Supplementary Table 1) are key parameters to find the optimal
90 inverse adjustment to achieve the mass balance. We rely on published uncertainty estimates with the
91 exception of the EGC shelf branch. The reported error estimate is $\pm 0.46 \text{ Sv}$ which stems from the gridding
92 procedure alone³⁷. This is likely an underestimate because the mooring array did not extend across the
93 width of the Greenland shelf. We assign a higher uncertainty of $\pm 1.0 \text{ Sv}$, hence EGC shelf has the highest
94 uncertainty among the 11 branches. With this uncertainty, 58% of the inverse model modifications in
95 volume transport are applied to the PW layer. If we assign a lower uncertainty of $\pm 0.5 \text{ Sv}$ to the EGC shelf
96 transport, the inverse model modifications applied to the PW layer are reduced to 50%. This corresponds
97 to a change in 1 TW in the 1993–2016 mean ocean heat transport, which is well below the total uncertainty
98 of 26 TW and demonstrates that the inversion is not very sensitive to the high uncertainty of the EGC shelf
99 transport.

100

101 **Box inverse model adjustments**

102 Supplementary Fig. 1 shows the net volume transport time series before and after the inversion.
103 This net volume transport is balanced by other terms in the mass budget (Eq. 1), such as the F_{ma}^i and
104 $(F_{\text{ma}}^{\text{surf}} - \frac{\partial M}{\partial t})$. Without adjustments, the mean imbalance is $-0.3 \pm 1.4 \text{ Sv}$. The month-to-month changes range
105 from -4 to $+3 \text{ Sv}$. It has a clear seasonality, with a net inflow in winter and a net outflow in summer
106 because of the AW volume transport seasonality²⁰. During the most data-rich period of 2004–2010 the
107 imbalances were generally reduced. After the inversion the net volume transport becomes $-0.2 \pm 0.2 \text{ Sv}$ with
108 a distinct seasonal cycle that peaks in June. In June, a net outflow is largely balanced by the increased river
109 runoff to the Arctic Ocean⁵².

110 We next examine how inverse modification was introduced (Supplementary Fig. 2). The full
111 depth inverse model modifications are $0.1 \pm 1.3 \text{ Sv}$ (mean \pm standard deviations). For that, each water mass
112 inverse model modifications are $0.0 \pm 0.7 \text{ Sv}$ in the PW layer, $0.0 \pm 0.3 \text{ Sv}$ in the AW layer, and $0.0 \pm 0.2 \text{ Sv}$
113 in the OW layer. As mentioned in previous paragraph, 58% of the inverse model modifications in volume
114 transport are applied to the PW layer.

115

116 **Quantification of the ocean heat transport**

117 From the mass-balanced ocean circulation, the ocean heat transport F_{H}^0 can be quantified as the
118 product of the temperature difference between compensating inflows and outflows and their volume
119 transports. In practice, the ocean heat transport is estimated as a summation of temperature transports for
120 each branch j , such that

121
$$F_{\text{H}}^0 = \sum_{j=1}^N \rho_o c_p^o V_j (\Theta_j - \Theta_{\text{ref}}) \quad (\text{Eq. 3})$$

122 where ρ_o is the density of sea water, c_p^o is the specific heat capacity of sea water, V_j is the volume
123 transport estimate from the inverse model, Θ_j is the transport-weighted potential temperature, and Θ_{ref} is a
124 reference temperature here set to be 0°C . Note that we distinguish heat transport and temperature transport
125 by terminology and units^{29,41}. When the transport is sensitive to the choice of reference temperature, we
126 refer to it as temperature transport with unit W-eq (Watt-equivalent). This is the case when the sum of all
127 V_j is not zero, such as estimates from single gateways or water masses. The transport becomes
128 independent from the choice of reference temperature when mass is balanced (the sum of all V_j is zero),
129 then we refer to it as heat transport with unit W.

130 As summarized in Supplementary Table 3, for 5 of the 11 branches the transport-weighted
 131 potential temperatures Θ_j are calculated from published temperature transport measurements (Bering
 132 Strait³⁸, Davis Strait west³⁶, Davis Strait east³⁶, NIIC³² and IF¹⁵). For the remaining branches we use
 133 alternative information to represent the temperature variability. We use (1) moored temperature time series
 134 for DS⁵⁸ and FBC²²; (2) annual mean temperature variability observed at sustained long-term hydrographic
 135 transects¹⁴ with repeat seasonal cycle for FSC and ES branches⁵⁹; (3) fixed temperature for EGC shelf⁶⁰
 136 and WTR³⁴ branches. For the FSC and ES branches, published annual mean Shetland Shelf temperature
 137 (defined as temperature at salinity cores on the Shetland shelf)¹⁴ is used, but for the FSC the annual mean
 138 temperature is lowered by 1°C to represent the FSC temperature by considering the temperature and
 139 velocity distributions along the Fair Isle – Munken transect³¹. For the FSC and ES seasonal cycles with
 140 amplitudes of 0.8 and 1.3°C, respectively, that peak in September are prescribed on the basis of the
 141 observed upper ocean temperature variability along the section⁵⁹.

142

143 **Error estimates**

144 After the inversion, a posteriori uncertainties are calculated as the square root of the diagonal
 145 component of the error covariance matrix^{28,40}. Regarding the uncertainty of the long-term mean, we note
 146 that there are two different kinds of uncertainty: random and systematic⁶¹. Random uncertainty follows a
 147 probability distribution function and can be reduced by frequent sampling. Systematic uncertainty, on the
 148 other hand, is independent of the number of samples. We note that it is sometimes challenging to
 149 categorize the uncertainty into the two different types. Moreover, the sources of uncertainty are not always
 150 well documented in the literature. As such, we assume that the uncertainty is entirely systematic, which
 151 results in very conservative error estimates.

152

153 **Statistical significance of the heat transport increase**

154 The statistical significance of the difference in ocean heat transport between the 1993-2000
 155 average and 2002-2016 average of 21 TW was evaluated using Welch's t-test and a bootstrap method⁴³.
 156 For Welch's t-test, the statistic Z was calculated as follows:

$$158 Z = \frac{\bar{X}_1 - \bar{X}_2}{\sqrt{\frac{s_1^2}{N_1} + \frac{s_2^2}{N_2}}} \quad (\text{Eq. 4})$$

159 where \bar{X}_j , s_j , N_j are sample average, sample standard deviation, and sample size, respectively, over the j^{th}
 160 period. The statistical significance of the heat transport difference between the two periods was then
 161 evaluated using a two-tailed test and assuming a normal distribution (p-value 0.0006). The statistical
 162 significance was also evaluated using the bootstrap method, which is a procedure that involves random
 163 sampling with replacement from the dataset and does not require any assumptions about the underlying
 164 probability distribution. The original time series were randomly shuffled to generate 10,000 pseudo time
 165 series to test the statistical significance (p-value 0.0007). For both cases the increase in ocean heat
 166 transport of 21 TW between the two periods was found to be statistically significant above the 99.9%
 167 confidence level.

169

170 **Decomposition of the ocean heat transport**

171 To investigate causes of variability, the ocean heat transport $F_H^0\{V, \Theta\}$ is decomposed as follows:

172

173
$$F_H^o\{V, \theta\} = F_H^o\{\bar{V}, \bar{\theta}\} + F_H^o\{\bar{V}, \theta'\} + F_H^o\{V', \bar{\theta}\} + F_H^o\{V', \theta'\} \quad (\text{Eq. 5})$$

174 where overbar indicates average over time, and prime indicates deviation from the mean. Thus, the term
175 $F_H^o\{\bar{V}, \bar{\theta}\}$ represents the “mean transport” component with no temporal variability, the term $F_H^o\{\bar{V}, \theta'\}$ is the
176 “temperature-driven” component resulting from temperature variability in each current, $F_H^o\{V', \bar{\theta}\}$ is the
177 “velocity-driven” component resulting from volume transport variability in each current, and $F_H^o\{V', \theta'\}$ is
178 the “correlation term” resulting from the covariance of the two.

179

180

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182

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