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Multi-decadal weakening of the Atlantic Overturning from a physics and observation-based reconstruction

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The Atlantic meridional overturning circulation (AMOC) is a key component of the global ocean circulation, transporting heat, salt, nutrients and carbon between hemispheres. Observational proxies and models suggest an AMOC slowdown over the 20th and 21st Centuries^{1,2,3,4,5}, yet its historical variability remains poorly constrained, and there are inconsistencies across studies and methodologies^{6,7,8}. Here we reconstruct AMOC strength and variability since 1940 using a physically based water-mass transformation framework⁹ applied to multiple observation-based ocean datasets and atmospheric reanalysis fields. This approach correctly estimates the overturning streamfunction in density space across the Atlantic by incorporating air-sea buoyancy fluxes, subsurface mixing, and volume tendencies. The reconstructed AMOC strength closely matches direct 21st Century measurements from the RAPID, OSNAP, and SAMBA arrays. Moreover, the reconstruction reveals a long-term AMOC weakening of 13.7, 8.3, and 7.3 Sverdrups (1 Sverdrup = $10^6 \text{ m}^3 \text{ s}^{-1}$) per century since 1960 at 34.5°S, 26°N, and in the subpolar cell, respectively. While air-sea fluxes account for ~80% of the time-mean AMOC strength at 26°N, the long-term decline is primarily driven by reductions in subsurface mixing and a decline in the upper-limb volume. These results provide a new practical framework for evaluating AMOC evolution and emphasize the importance of sustained ocean observations for tracking modern day AMOC changes.

The Atlantic meridional overturning circulation (AMOC) is a system of ocean currents that transports warm surface and thermocline waters to high latitudes in the North Atlantic and returns cold, dense deep waters southward across the Equator. A vast amount of heat carried by the upper branch of the AMOC is released to the atmosphere at midlatitudes, contributing to the temperate climate of northwest Europe and the UK¹⁰. The AMOC is projected to weaken over the 21st Century and beyond due to anthropogenic warming and the associated freshening of the subarctic Atlantic from increased rainfall, melting sea ice, and Greenland Ice Sheet and Arctic glacial melt^{4,11,12,13,14}. A slowdown of the AMOC would disrupt the northward transport of heat to the northern high latitudes¹⁵, with profound impacts on climate, sea-level, and marine ecosystems across the Atlantic and beyond.

Monitoring AMOC variability is crucial for assessing how the AMOC is responding to and influencing climate change. Multiple lines of observational evidence suggest a weakening of the AMOC since the 20th Century^{16,17,18,19}; however, the short duration of direct, continuous measurements has led to debate about the robustness of these changes. Ship-based hydrographic estimates of the AMOC at 25°N suggest a 30% slowdown in AMOC strength from 1957 to 2004^{16,20}, but these snapshot measurements could be aliased by large seasonal and subannual variability^{20,21,22}. This highlights the need for extended continuous observations to determine AMOC variability¹⁸, and also motivates the need for robust methods to reconstruct past AMOC change using incomplete data sets. The most comprehensive efforts to date for monitoring the AMOC include the Rapid Climate Change-Meridional Overturning Circulation and Heatflux Array (RAPID-MOCHA, hereafter, “RAPID”), which has been monitoring the AMOC at 26°N continuously since April 2004; the Overturning in the Subpolar North Atlantic Program (OSNAP), which has observed overturning in the Labrador Sea (OSNAP West) and between Greenland and Scotland (OSNAP East) since 2014; and the South Atlantic MOC Basin-wide Array (SAMBA), which has measured the AMOC at 34.5°S since 2009 (Fig. 1).

Although the RAPID observations have identified a weakening of the AMOC at 26°N during the past two decades^{18,23}, this observed change is over too short a period to distinguish between low-frequency natural variations and long-term anthropogenically-forced change^{11,18}. Approaches combining satellite altimetry with *in-situ* measurements from ships and Argo floats can extend the AMOC time series back to 1993^{17,24,25,26}, however these records still remain temporally limited for assessing long-term trends.

Due to the lack of sufficiently long direct measurements, the response of the AMOC to anthropogenic warming has been inferred indirectly from AMOC fingerprints^{1,2,3}. Multiple AMOC fingerprints have been proposed, including the ‘warming hole’ fingerprint^{1,2}, characterized by a region of surface cooling averaged over the subpolar North Atlantic relative to background global warming, and the South Atlantic sea surface salinity (SSS) pileup fingerprint^{3,27} (Fig. 2). While these fingerprints enable reconstruction of AMOC variability back to the late 19th Century, sea surface temperature (SST) and SSS are influenced by variability independent of the AMOC, and furthermore, climate models project the sensitivity of the warming hole fingerprint to decrease under future warming scenarios²⁷. Therefore, using indirect fingerprints introduces potential uncertainties in estimating the long-term AMOC trend¹¹.

Here we present a new approach that bridges direct measurements and indirect fingerprints, by directly quantifying the physical processes that characterize the AMOC using observationally-constrained ocean and atmosphere fields, enabling a reconstruction of the historical AMOC evolution (Extended Data Figs. 1–3). The meridional overturning in the North Atlantic is dominated by the buoyancy-forced thermohaline circulation²⁸. Hence the thermodynamic framework of Walin²⁹ (also known as the water-mass transformation, or WMT, framework; Extended Data Figs. 1) has been applied to estimate the buoyancy-forced circulation in the Atlantic^{9,28,30,31,32,33}. Previous studies have focused on the surface buoyancy-

driven circulation, while subsurface mixing has remained poorly parameterized in observation-based studies. Here we use observationally-constrained atmospheric reanalyses alongside subsurface ocean observations to estimate the AMOC streamfunction. This is based on a full volume budget in density space from the WMT framework⁹ that incorporates air-sea fluxes, shortwave radiation penetration³⁴, mesoscale horizontal and isopycnal mixing, small-scale vertical mixing, a state-of-the-art neutral slope calculation algorithm³⁵, observational-based mesoscale eddy³⁶ and vertical diffusivity estimates³⁷, and volume tendency (Fig. 1; Extended Data Figs. 1–3; Methods). This novel application of the methodology enables us to reconstruct the historical AMOC evolution from known physical processes, and in doing so, to re-examine whether the recent AMOC weakening is part of a long-term trend or low frequency variability.

Observation-based Atlantic overturning streamfunction

In ocean and climate models, the Atlantic overturning streamfunction is commonly computed by integrating the meridional or vertical volume flux as a function of depth or density and latitude^{31,38,39}. While this integration is relatively straightforward in models (since velocity fields are standard outputs), obtaining observation-based estimates of the AMOC is more challenging and has generally been limited to components driven by surface buoyancy fluxes. Here, for the first time, we derive the meridional volume flux along density surfaces through a volume budget that balances the rate of WMT due to surface buoyancy fluxes, shortwave radiation depth penetration, interior mesoscale neutral and horizontal mixing, and small-scale vertical mixing, with the local volume tendency (Extended Data Fig. 1). This approach, based on the WMT framework proposed by Walin²⁹ and further developed in more recent studies^{9,28,30}, enables us to reconstruct the AMOC streamfunction and its strength from multiple observation-based ocean temperature and salinity datasets combined with atmospheric reanalysis fields (see Methods, Extended Data Figs. 1–3 and Figs. 1, 3). Specifically, Equation 5 (Methods) is

applied within the Atlantic basin to compute the accumulated meridional volume transport as a function of latitude and density (Figs. 1, 3). The resulting time-mean AMOC streamfunction, evaluated over the common period 2004–2023 using the IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets, captures the upper and lower limbs of the meridional overturning circulation as well as the Atlantic subtropical and subpolar cells (Figs. 1, 3a; Extended Data Fig. 2).

The time-mean reconstructed AMOC streamfunction features the three main overturning cells within the upper limb of the AMOC in the Atlantic; namely, the anticlockwise shallow equatorial cell (blue cell in Fig. 3a, at approximately 0–100 m), the clockwise light subtropical cell in the North Atlantic (yellow contour in Fig. 1; upper red cell in Fig. 3a, at approximately 0–300 m and 10°–40°N), and the intermediate overturning cell⁴⁰ (or, subpolar cell north of 40°N) extending from the intermediate to deep water layers and spanning all Atlantic latitudes (red contour in Fig. 1; lower red cell in Fig. 3a, at approximately 300–2500 m). This structure indicates that light surface waters entering the North Atlantic are transformed into lighter waters within the equatorial cell and into denser Subtropical Mode Water within the subtropical cell, while both subtropical and intermediate waters – primarily Antarctic Intermediate Water originating from the Southern Ocean – contribute to the formation of Labrador Sea Water and North Atlantic Deep Water within the subpolar cell (Figs. 1, 3; Extended Data Fig. 2). The distribution of these three overturning cells is broadly consistent with estimates from a reanalysis product³² as well as simulations from ocean-only models and climate models in density space^{4,31,40}, although our reconstruction shows denser subtropical and subpolar cells compared with Coupled Model Intercomparison Project Phase 6 (CMIP6) simulations (Extended Data Fig. 4 of ref.⁴), likely due to CMIP6 model biases.

The North Atlantic subtropical cell is associated with WMT occurring around the Gulf Stream, its extension, and the North Atlantic Current^{39,41}, and reflects the formation of Subtropical Mode Water (Figs. 1, 3a; Extended Data Figs. 2, 4a). The peak cumulative

transport within the subtropical cell, based on the ensemble mean of all datasets, occurs between 10°N – 40°N and within the density range 25.5 – 26.5 kg m^{-3} , with a magnitude (and one standard deviation during 2004–2022) of approximately $24.6 \pm 3.0 \text{ Sv}$ ($\text{Sv} = \text{Sverdrup} = 10^6 \text{ m}^3 \text{ s}^{-1}$; Extended Data Fig. 4b). While our reconstruction exhibits a strong subtropical cell and the associated reversed southward transport forming its lower limb (Figs. 1, 3a; Extended Data Figs. 2, 4; see also refs.^{31,40}), this feature is absent in RAPID observations at 26°N ^{18,19} (Fig. 4b), likely because the depth-coordinate RAPID streamfunction does not directly resolve the subtropical cell⁴⁰. Nevertheless, our reconstruction captures the strength of the intermediate overturning cell (Fig. 4). Note that the subtropical cell represents only part of the upper branch of the overturning circulation; hence, the AMOC strength should not be defined solely by the subtropical cell.

The intermediate overturning cell is associated with the formation of Labrador Sea Water and North Atlantic Deep Water in the subpolar North Atlantic and their southward transport as the lower limb of the AMOC (Figs. 1, 3a). Following this physics-based protocol and existing definitions^{3,4,18,19,27}, we define the AMOC strength as the maximum streamfunction of this intermediate and deep-layer overturning cell (maxima within approximately 27.2 – 27.6 kg m^{-3} ; Methods; Figs. 1, 3a), representing the upper branch of AMOC transport that reaches the subpolar North Atlantic and supplies deep water formation, and examine its historical variability since the mid-20th Century (Fig. 3b). The peak cumulative AMOC volume transport has a time-mean strength and one standard deviation during 2004–2022 (chosen to match available estimates from the RAPID array) of approximately $17.2 \pm 1.9 \text{ Sv}$ within 10°N – 40°N (dark blue box in Fig. 3a), and $16.4 \pm 1.9 \text{ Sv}$ at 26°N (red line in Fig. 3). The reconstructed AMOC strength estimates closely match the time-mean AMOC strength of $17.4 \pm 3.2 \text{ Sv}$ estimated by RAPID arrays at 26°N through February 2023⁴² (hereafter, “RAPID-AMOC”; black line in Fig. 3b; Extended Data Fig. 5). The reconstructed peak cumulative northward

transport within the subpolar cell north of 40°N is approximately 16.0 ± 1.0 Sv during 2014–2020 (again chosen to match the available observation record; light blue box in Fig. 3a), only slightly weaker than the subpolar AMOC strength of 16.7 ± 3.7 Sv estimated by OSNAP arrays across the Labrador Sea and between Greenland and Scotland (grey line in Fig. 3b).

Observed changes in the AMOC strength

In Fig. 3b we show the time evolution of the reconstructed AMOC strength, defined as the maximum streamfunction within 10°N–40°N (dark blue line), at 26°N (red line), and within the subpolar cell north of 40° N (light blue line) for the period 1940–2023, alongside several existing estimates from the literature. The ensemble-mean AMOC streamfunction reconstructed using the WMT framework suggests that the AMOC strength within 10°N–40°N experienced a near monotonous decline from 1950 through to the early 2000s, followed by a rate of decline that has more than doubled since 2004. While the AMOC strength at 26°N and in the subpolar cell reached a local minima between the 1970s and 1990s, it temporarily rebounded in the late 1990s, but then has since exhibited a sustained decline. The disagreement among different datasets (ensemble spread, shown by light shading in Fig. 3b) occurs mainly before the 1980s, reflecting sparse measurements and differences in data sources and objective analysis methods employed in the reanalysis products. The reconstructed AMOC strength at 26°N and in the subpolar cell closely matches hydrographic estimates from nearby repeat ship-based transects in 1957, 1981, 1992, 1998, and 2004, corrected for seasonal variability by Atkinson et al.²⁰ (pink squares) and Kanzow et al.²² (purple triangles), as well as the AMOC strength derived from satellite altimetry for 1993–2014²⁴ (purple line, Figs. 3b, 4c). Although the reconstructed AMOC strength closely matches the time-mean RAPID-AMOC, it does not capture the high-frequency variability in that data set (Extended Data Fig. 6), as the reconstruction is based on annual averages derived from monthly mean subsurface temperature

and salinity fields.

The ensemble-mean AMOC strength, reconstructed using the WMT framework, reveals a long-term weakening of approximately 11.0, 8.3, and 7.3 Sv per century from 1960 to 2023 for the AMOC within 10°N–40°N, at 26°N, and in the subpolar cell, respectively. Extending the trend analysis back to 1940, these trends correspond to total declines of approximately 9.2, 7.0, and 6.1 Sv over the 84-year period since 1940, respectively. A comparison of the ensemble-mean AMOC strength during the Argo period (2004–2023) relative to 1960–1980 indicates a weakening of approximately 23%, 18%, and 16%, for the AMOC within 10°N–40°N, at 26°N, and in the subpolar cell, respectively (Figs. 3b, 5a). The long-term decline revealed by our reconstruction is overall consistent with an estimate of ~15% reduction since the mid-20th Century reported by Caesar et al.². In addition, the strength of the AMOC upper branch from the Southern Ocean to the subpolar cell in the North Atlantic (measured by the maximum cumulative transport from the surface to the intermediate-deep water layer at each latitude) has exhibited a sustained weakening since the mid-20th Century (Fig. 5a; Extended Data Figs. 7, 8).

A breakdown of the reconstructed AMOC strength at 26°N reveals that the air-sea flux component accounts for the majority (~80%) of the time-mean AMOC strength during 2004–2023 (Fig. 4c; Extended Data Figs. 3, 8), consistent with previous studies^{28,30,31,39}. However, the long-term weakening trend of the AMOC at 26°N is primarily driven by declines in the subsurface mixing and upper-limb volume tendency components, whereas the air-sea flux-induced overturning mainly exhibits decadal variability (Fig. 4c). This is consistent with the lack of an AMOC decline since the 1960s when inferred from air-sea heat flux reanalysis products^{8,43}, which do not incorporate subsurface mixing or volumetric changes in the AMOC. Subsurface mixing processes play a crucial role in modulating the AMOC, particularly in the subpolar North Atlantic⁴⁴ (Extended Data Figs. 3, 8, 9; Fig. 5). Our physics-based

reconstruction shows that the weakening of the vertical mixing component has contributed substantially to the AMOC decline across the Atlantic basin since the 1950s (Figs. 4c, 5; Extended Data Figs. 1, 8). This weakening is primarily driven by its salinity contribution (Extended Data Figs. 8, 9; Eq. 3.5), which is likely induced by enhanced upper-ocean salinity stratification within the density range 26.6–27.3 kg m⁻³ in the subpolar North Atlantic (Fig. 5e). The volume tendency component represents temporal changes in the volume of the AMOC upper limb and is associated with variations in advection, water-mass formation, and wind-driven processes. This term also exhibits a long-term decline alongside pronounced multidecadal variability, contributing significantly to the AMOC weakening (Fig. 4c; Extended Data Figs. 8, 9). Its variability tends to oppose that of the air–sea flux component, and its decline is associated with a thinning of the AMOC upper branch within and above the intermediate-water layer, and a thickening of the deep-water layer in the subpolar North Atlantic south of Greenland. Our reconstruction also highlights the importance of accounting for the depth penetration of solar radiation, which can affect the AMOC streamfunction estimate by 2–3 Sv (Extended Data Fig. 10), yet was often not considered in previous studies.

The South Atlantic MOC and fingerprints

Constrained by limited ship-based hydrographic measurements and mooring arrays, most observational assessments of historical AMOC evolution as well as numerical model projections of future change under anthropogenic climate forcing have focused on the North Atlantic. While often viewed as a passive conduit for deep waters formed in the North Atlantic and Southern Ocean, the South Atlantic also plays an active role in inter-basin water mass exchange in the upper ocean, particularly through the formation and export of low salinity Antarctic Intermediate Water originating from the Southern Ocean (Fig. 1). The northward transport of Antarctic Intermediate Water from the South Atlantic to the subpolar North

Atlantic constitutes a substantial contribution to the upper limb of the AMOC⁴⁵. To improve understanding of the South Atlantic's role in the AMOC system, the SAMBA observing array was deployed along 34.5°S in 2009 to monitor key components of the circulation (Figs. 1, 5a). The observed time-mean AMOC strength at 34.5°S from September 2013 to July 2017 is 17.3 ± 5.0 Sv⁴⁶, close to the reconstructed strength we estimate of 16.8 ± 1.5 Sv over the same period (Extended Data Figs. 7, 8a). Unlike mooring arrays or ship-based observations that are geographically limited, our reconstruction of the AMOC streamfunction using the WMT framework leverages globally available observational and reanalysis fields, enabling estimation of AMOC strength across all Atlantic latitudes (Figs. 1, 5a; Extended Data Fig. 2). This reconstruction shows that the AMOC reaches peak strength in the Southern Ocean, where the largest volume of Antarctic Intermediate Water is formed, and weakens as it moves northward toward the subpolar North Atlantic. Notably, this upper limb has undergone a sustained weakening across the Atlantic basin since the mid-20th Century, with the South Atlantic exhibiting both the largest decline and the strongest decadal variability (Fig. 5a; Extended Data Fig. 7). For example, our reconstruction indicates a long-term weakening trend of 13.7 Sv per century for the AMOC at 34.5°S from 1960 to 2023 (Fig. 3b). The pronounced weakening of the South Atlantic upper limb may further serve as a predictor of AMOC behaviour in the North Atlantic, with changes observed at 34.5°S propagating northward to 40°N over approximately 20 years, and reaching 60°N after a further 5 years (Extended Data Fig. 7b), suggesting a continued decline in the coming decades. Note that the temporal variability of the overturning streamfunction south of 34.5°S (beyond the tip of South Africa) is influenced by zonal advection across the open South Atlantic basin boundaries, therefore our analysis of AMOC strength changes is restricted to north of 34.5°S.

We also compare the reconstructed AMOC variability with that inferred indirectly from AMOC fingerprints, including the widely used “warming hole” fingerprint and a salinity-based

fingerprint, which reflects a salinity pileup over the subtropical South Atlantic associated with the anomalous salinity convergence caused by AMOC changes³ (Fig. 2). Both fingerprint-based indices indicate an overall AMOC slowdown from the mid-20th Century to the present. However, the SST-based index shows a rebound in the 1990s that is not as prominent in either the salinity-based index or our AMOC reconstruction. While AMOC fingerprints are useful for capturing long-term climate-driven AMOC variability, they provide only indirect estimates for possible AMOC changes. In contrast, our AMOC reconstruction using the WMT framework provides a direct, physically-consistent, and observation-constrained estimate of AMOC changes by quantifying key processes over the instrumental period combining historical measurements, atmospheric reanalyses, and Argo observations. Given the difficulties of directly and continuously measuring the basin-scale overturning circulation, this approach provides both a practical framework and a fresh perspective for quantifying historical AMOC variability and assessing future changes under continued climate forcing.

Summary and discussion

The AMOC is a fundamental component of the global climate system, redistributing heat across latitudes and influencing regional and global climate variability. To date, the most comprehensive observational efforts to continuously monitor AMOC variability include the RAPID array at 26°N since April 2004, the OSNAP array in the subpolar North Atlantic since 2014, and the SAMBA array at 34.5°S since 2009 (Fig. 1). While reconstruction approaches that combine satellite altimetry with *in-situ* measurements from ships and Argo floats have extended the AMOC record back to 1993^{17,24,25,26}, both the direct record and previous reconstructions remain temporally limited for distinguishing low-frequency climate variability from long-term anthropogenically-forced trends. Due to the scarcity of direct long-term measurements, multiple proxy-based AMOC fingerprints, particularly those based on SST and

SSS, have been widely used to investigate the response of AMOC to anthropogenic warming^{1,2,3}, generally suggesting a long-term AMOC weakening since the late 19th Century. However, fingerprint-based indices potentially introduce uncertainties that are not directly related to the AMOC, such that the degree to which the AMOC has changed cannot be confidently inferred from the indirect indexes.

In this study we provide a new comprehensive reconstruction of the AMOC using observations of ocean temperature and salinity combined with atmospheric reanalyses, exploiting a WMT framework^{9,28,29}. This approach provides a physically grounded estimate of the overturning streamfunction across the Atlantic basin, incorporating contributions from air-sea buoyancy fluxes, subsurface mesoscale and small-scale mixing, and volumetric tendencies (Figs. 1, 3; Extended Data Figs. 1–3). The reconstructed streamfunction quantifies the strength of the AMOC, defined by the maximum integrated northward volume transport from the surface to the intermediate and deep layers, for the low-latitude North Atlantic (within 10°N–40°N; 17.2 ± 1.9 Sv), at 26°N (16.4 ± 1.9 Sv), and within the subpolar cell (north of 40°N; 16.0 ± 1.0 Sv) (Fig. 3). This reconstruction allows assessment of the AMOC and its variability across all latitudes spanning the North and South Atlantic basins. The time-mean strength of the reconstructed AMOC closely matches direct 21st Century measurements from the RAPID, OSNAP, and SAMBA arrays (Fig. 5a). From 1960 to 2023, the reconstructed AMOC has exhibited a linear decline of approximately 13.7, 8.3, 11.0, and 7.3 Sv per century at 34.5°S and 26°N, within 10°N–40°N, and in the subpolar cell, respectively (Fig. 3b). We further find that AMOC variations in the South Atlantic upper limb may serve as a predictor of AMOC behaviour in the North Atlantic, with lead times of approximately 20–25 years (Extended Data Fig. 7b). Moreover, the continued weakening of the AMOC at 34.5°S may be a precursor to further declines in the North Atlantic over the coming decades.

The updated WMT framework enables a breakdown of the processes influencing AMOC strength and its change. At 26°N, while the time-mean AMOC strength is dominated (~80%) by air-sea fluxes, its long-term weakening is primarily driven by reductions in subsurface mixing and a decline in the volume of the AMOC upper limb (Fig. 4c). This agrees with the absence of an AMOC decline since the 1960s when considering only the surface buoyancy flux component, as found in past studies^{8,43}. The weakening vertical mixing component is largely attributable to its salinity contribution (Extended Data Figs. 8, 9; Eq. 3.5), likely reflecting enhanced upper-ocean salinity stratification in the subpolar North Atlantic (Fig. 5e). The volume tendency shows strong decadal variability alongside a long-term decline (Fig. 4c; Extended Data Fig. 8), likely linked to a thinning of the AMOC upper branch and a thickening of the deep-water layer in the subpolar North Atlantic.

In conclusion, using a WMT framework combined with observations and reanalyses provides a practical physics-based approach for assessing the historical evolution of the AMOC. This method could also be applied for better constraining future projections of AMOC behaviour compared to traditional velocity-based overturning calculations. This physics-based reconstruction is largely independent of existing AMOC strength estimates and fingerprints, exhibiting weaker decadal variability than previously proposed SST-based indices, and its extended record back to 1940 allows us to analyse a longer term record of AMOC behaviour. Our results reveal a long-term multi-decadal weakening of the AMOC, and highlight the essential roles of surface buoyancy forcing, subsurface mixing^{31,32,33}, and volume tendencies of intermediate and deep waters in shaping AMOC behaviour. At the same time, two main limitations should be acknowledged: first, the inferred AMOC weakening relies on estimates of surface buoyancy fluxes, subsurface mixing, and volume tendency changes, which can exhibit discrepancies across datasets especially when constrained by sparse observations prior to the Argo and satellite eras. And second, the diffusivities applied for mesoscale³⁶ and vertical

340 mixing³⁷ are taken to be constant in time. Nonetheless, consistency across reconstructions from
341 different datasets emerges after 1980 when observational data coverage improved markedly,
342 adding confidence in the methods applied in this study.

343 Continued AMOC weakening could trigger profound climate disruptions, including
344 colder winters in Europe, increased atmospheric disturbances in the northern mid-latitude jet
345 stream and storm track^{47,48}, reduced ocean carbon uptake⁴⁹, and shifts in tropical rainfall
346 patterns¹⁵. These potential impacts, together with the observation-based approach
347 demonstrated here, underscore the critical importance of sustained and expanded ocean
348 observing systems, such as RAPID, OSNAP, SAMBA, and deep-Argo arrays, to improve the
349 accuracy of AMOC monitoring and enhance the understanding of future climate projections.

Methods

Observational and reanalysis data.

The new reconstructions presented here are calculated using five different observational gridded ocean datasets: (1) The Institute of Atmospheric Physics (IAP) data (version 4.2 *in-situ* temperature and version 2 salinity observations for the upper 6000-m from refs.^{50,51}), monthly averaged from 1940 to 2023 with 1° horizontal resolution, referred to as “IAPv4.2 data”; (2) EN4.2.2 data from the UK Met Office Hadley Center (*in-situ* temperature and salinity observations for the upper 5500-m from ref.⁵²), monthly averaged from 1940 to 2023 with 1° horizontal resolution, referred to as “EN4 data”; (3) The Scripps Institution of Oceanography (SIO) Argo product (*in-situ* temperature and salinity observations constructed from Argo data for the upper 2000-m), monthly averaged from 2004 to 2023 with 1° horizontal resolution from ref.⁵³, hereafter referred to as “SIO RG-Argo”; (4) The Japan Meteorological Agency data (*in-situ* temperature and salinity observations version 7.3.1 for the upper 3000-m from ref.⁵⁴), monthly averaged from 1955 to 2023 with 1° horizontal resolution, referred to as “Ishii data”; and (5) World Ocean Atlas 2023 data (*in-situ* temperature⁵⁵ and salinity⁵⁶ observations for the upper 5500-m), monthly climatology averaged within the time periods 1975–1984, 1985–1994, 1995–2004, 2005–2014, and 2015–2022 with 1° horizontal resolution, referred to as “WOA 2023”. This study also uses IAP version 4 temperature combined with version 1 salinity (referred to as “IAPv4”) to construct a climatology for the period 2004–2023 (Fig. 4b; Extended Data Fig. 3a). The EN4 data used in this study includes all four ensemble members available from the latest version of the EN4 product, with four different sets of XBT and mechanical bathythermograph (MBT) bias corrections applied (see ref.⁵² for more details). When presenting results using the EN4 data, we choose to show the ensemble average from those four EN4.2.2 ensembles (referred to as “EN4-ESM”), to mitigate the uncertainty arising from different approaches applied in correcting biases in XBT and MBT data. The ensemble-

averaged AMOC reconstruction and its associated one standard deviation variation across time and data products are based on the IAPv4.2, EN4-ESM, SIO RG-Argo, and Ishii datasets. Observational data coverage improved substantially after the 1950s and 1960s with the introduction of mechanical bathythermographs (MBTs), conductivity-temperature-depth (CTD) instruments, and expendable bathythermographs (XBTs); therefore, the main trend analysis evaluated in this study begins in 1960.

All observed temperature and salinity measurements are converted to conservative temperature (Θ , ref.⁵⁷) and absolute salinity (S_A , ref.⁵⁸) using the Gibbs-Sea Water (GSW) Oceanographic Toolbox from the TEOS-10 software⁵⁹. Static stability of the water column on timescales of months is achieved by applying a vertical stabilization algorithm⁶⁰. The thermal expansion coefficient, α , haline contraction coefficient, β , cabbeling coefficient, C_b , thermobaricity coefficient, T_b , and surface-referenced potential density, ρ_0 , are then calculated using functions from the GSW Oceanographic Toolbox. The neutral density, γ^n , is calculated using functions from PreTEOS-10 software⁶¹.

The air-sea flux fields analyzed in this study are taken from the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis Version 5 (ERA5), using monthly averaged fields from 1940 to 2023 over a $1/4^\circ$ horizontal resolution grid. Variables analyzed from ERA5 include air-sea heat ($W m^{-2}$) and freshwater fluxes ($m s^{-1}$) at the ocean surface. The net air-sea heat flux is the sum of the net longwave and shortwave radiation fluxes and the latent and sensible heat fluxes. The surface freshwater flux is taken as the difference between precipitation F_P and evaporation F_E . Freshwater fluxes from river runoff and sea-ice melt are not included, as their effects are expected to be small in regions of North Atlantic Deep Water formation. Positive air-sea fluxes represent a buoyancy gain in the ocean surface layer. The observation-based SST ($^\circ C$) fields analyzed in this study are taken from the UK Met Office Hadley Centre Sea Surface Temperature (HadISST) Version 1.1 from 1870 to 2023, the NOAA

Extended Reconstruction SSTs Version 5 (ERSSTv5) from 1854 to 2023, and the ERA5 SST from 1940 to 2023.

AMOC reconstruction using the WMT framework

Definition of AMOC strength. We calculate the annual mean overturning streamfunction in density space using various observation-based oceanic and atmospheric datasets (e.g., IAPv4.2, EN4.2.2, SIO RG-Argo, Ishii, WOA23, and ERA5), incorporating contributions from surface buoyancy fluxes, shortwave radiation penetration, parameterized mesoscale and small-scale eddy mixing, and volume tendency components (Fig. 3a; Extended Data Figs. 1–3, 8–10). The overturning streamfunction is computed using the water-mass transformation (WMT) framework by closing a volume budget in density space (detailed below). By calculating the AMOC streamfunction in density coordinates, we capture the natural pathways of water flowing along isopycnals, which provides a more continuous representation of the overturning between the subtropical and subpolar North Atlantic, and highlights the WMT that occurs as the water circulates cyclonically around the subpolar North Atlantic^{31,32,38,40}. Following refs.^{3,4,19,27,40}, where the AMOC is defined as the Atlantic Ocean mid-depth overturning cell, we calculate AMOC strength in density space from the maximum value of the Atlantic overturning streamfunction within the intermediate and deep-water layer below $\gamma^n = 26.6 \text{ kg m}^{-3}$ (Figs. 1, 3, 4; Extended Data Fig. 2). Specifically, we define the maximum streamfunction at 26°N as the ‘AMOC 26°N strength’ (red line in Figs. 3b, 4), over 10°N – 40°N as the ‘AMOC upper-limb strength’ (dark-blue line in Fig. 3), and over 40°N – 60°N as the ‘AMOC subpolar strength’ (light-blue line in Fig. 3).

Our reconstructed AMOC time series is compared with several existing estimates reported in the literature. The RAPID 26°N AMOC streamfunction and time series evaluated in density

space (through February 2023)⁴² were averaged to monthly means from the 10-day averages provided (red line in Extended Data Fig. 5b). We additionally applied a density threshold of $\sigma_0 > 26.6 \text{ kg m}^{-3}$ to derive the intermediate and deep-layer overturning strength (green line in Extended Data Fig. 5b; black line in Fig. 3b). The monthly OSNAP AMOC time series for 2014–2020 across the full array, and across the OSNAP West (southern Labrador to the southwestern tip of Greenland in the Labrador Sea) and OSNAP East (southeastern Greenland to Scotland) subsections⁶², were also analysed (grey line in Fig. 3b), as were the daily SAMOC time series from the SAMBA array at 34.5°S from September 2013 to July 2017⁴⁶ (yellow triangle in Fig. 5a; yellow line in Extended Data Fig. 7a). The northward volume transport from the surface to 1100 m, observed by repeat hydrographic sections along the latitude 25°N in 1957, 1981, 1992, 1998, and 2004, and adjusted for seasonal variability by Atkinson et al.²⁰, were analysed (shown as pink squares in Fig. 3b). The accompanying pink bars indicate a standard deviation of 2.9 Sv from RAPID observations over 2004–2009, as reported in Table 5 of ref.²⁰. Hydrographic estimates of the maximum streamfunction at the same latitude and years, corrected for seasonal variability by Kanzow et al.²², are shown as purple triangles in Fig. 3b. The monthly AMOC time series for 1993–2014, derived from satellite altimetry sea surface height²⁴, are included in Fig. 3b (purple line) for comparison with the AMOC strength that we reconstructed using the WMT framework.

The water-mass transformation (WMT) framework. We derive the meridional overturning circulation streamfunction based on the WMT framework following refs.^{9,29,63,64,65}. Consider a water mass volume sandwiched between two neutral density surfaces $\gamma^n - \Delta\gamma^n/2$ and $\gamma^n + \Delta\gamma^n/2$, and bounded by its intersection with the sea surface and by a latitude transect to the south in the North Atlantic as the interior boundary (Extended Data Fig. 1). The volume (ΔV) of this water mass is determined by $\Delta V = \iiint dV$, and the volume transport across the interior boundary is written as $\Delta\psi$. The dia-neutral volume flux of fluid crossing the $\gamma^n - \Delta\gamma^n/2$ and

450 $\gamma^n + \Delta\gamma^n/2$ iso-neutral surfaces can be written as $G(\gamma^n - \Delta\gamma^n/2)$ and $G(\gamma^n + \Delta\gamma^n/2)$,
 451 respectively. We choose a sign convention such that $\Delta\psi > 0$ indicates that water leaves the
 452 domain, and G is positive if directed toward increasing density.

453 Assuming incompressibility, the volume budget for the water mass bound by the iso-neutral
 454 surfaces $\gamma^n - \Delta\gamma^n/2$ and $\gamma^n + \Delta\gamma^n/2$, the ocean surface, and an interior boundary (e.g., a
 455 latitude in the North Atlantic), can be written as:

$$\underbrace{\frac{\partial(\Delta V)}{\partial t}}_{\text{volume tendency}} + \underbrace{\Delta\psi}_{\text{outflow}} = \underbrace{G(\gamma^n - \Delta\gamma^n/2) - G(\gamma^n + \Delta\gamma^n/2)}_{\text{water mass formation into } (\gamma^n - \Delta\gamma^n/2, \gamma^n + \Delta\gamma^n/2)}, \quad (1)$$

456 where $\partial(\Delta V)/\partial t$ represents the temporal change in layer volume, $\Delta\psi$ denotes the outflow of
 457 water volume across the interior boundary via ocean circulation, G represents the WMT due to
 458 air-sea buoyancy fluxes and mesoscale and small-scale ocean mixing (Extended Data Fig. 1),
 459 $\Delta G = G(\gamma^n - \Delta\gamma^n/2) - G(\gamma^n + \Delta\gamma^n/2)$ represents the convergence of G , which quantifies the
 460 water mass formation (or destruction) within the layer bound by $(\gamma^n - \Delta\gamma^n/2, \gamma^n + \Delta\gamma^n/2)$,
 461 so that $\Delta G > 0$ means that volume is added to the budget domain.

462 Following Groeskamp et al.⁹, the dia-neutral volume transport (quantified by the WMT
 463 rate) is written as

$$G(\gamma^n) = \frac{\partial}{\partial \gamma^n} \iiint_{V(\gamma_*^n \leq \gamma^n)} \frac{b \gamma_*^n}{\rho} [\rho_i] dV. \quad (2)$$

464 Here $b = |\nabla\gamma|/|\nabla\rho_i|$ is the ratio required to construct an approximate global neutral surface
 465 from surfaces that are only locally defined⁶⁴, ρ_i is the locally referenced potential density, ρ is
 466 the *in-situ* seawater density, $[\rho_i]$ is represented as a sum of physical process influencing the
 467 material derivative of density, and dV is the volume of water within the density range $\gamma_*^n \leq \gamma^n$.
 468 In the discrete approximations of Eq. (2), the WMT is calculated by partitioning the ocean into

469 $\Delta\gamma^n = 0.05 \text{ kg m}^{-3}$ density bins, i.e., $\gamma_*^n \in [\gamma^n - \Delta\gamma^n/2, \gamma^n + \Delta\gamma^n/2]$, spanning the density
 470 range $21 < \gamma^n < 29 \text{ kg m}^{-3}$.

471 The material time evolution of locally referenced potential density, $[\rho_l]$, due to surface
 472 air-sea buoyancy fluxes, $(\dot{\rho}_l)_F$, shortwave penetration fluxes, $(\dot{\rho}_l)_{SW}$, mesoscale neutral
 473 mixing (i.e., cabbeling and thermobaricity), $(\dot{\rho}_l)_N$, horizontal mixing, $(\dot{\rho}_l)_H$, and small-scale
 474 isotropic mixing (i.e., vertical mixing), $(\dot{\rho}_l)_V$, is determined by

$$(\dot{\rho}_l)_F = \left[-\frac{\alpha}{C_p} (Q_{lw} + Q_{lh} + Q_{sh}) - S_A \beta \rho_0 (F_E + F_P) \right] \delta(z - \eta), \quad (3.1)$$

$$(\dot{\rho}_l)_{SW} = -\frac{\alpha}{C_p} \nabla \cdot (Q_{sw} F(z)), \quad (3.2)$$

$$(\dot{\rho}_l)_N = \rho K_N C_b |\nabla_N \Theta|^2 + \rho K_N T_b \nabla_N P \cdot \nabla_N \Theta, \quad (3.3)$$

$$(\dot{\rho}_l)_H = \alpha \nabla \cdot \left[-\rho K_H \left(\frac{\partial \Theta}{\partial x} + \frac{\partial \Theta}{\partial y} \right) \right]_H - \beta \nabla \cdot \left[-\rho K_H \left(\frac{\partial S_A}{\partial x} + \frac{\partial S_A}{\partial y} \right) \right]_H, \quad (3.4)$$

$$(\dot{\rho}_l)_V = \alpha \nabla \cdot \left(-\rho D \frac{\partial \Theta}{\partial z} \right) - \beta \nabla \cdot \left(-\rho D \frac{\partial S_A}{\partial z} \right). \quad (3.5)$$

475 Here $C_p = 3992 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity of seawater, C_b is the cabbeling
 476 coefficient, T_b is the thermobaricity coefficient, and P is pressure. $\nabla_N \Theta$ and $\nabla_N P$ indicate the
 477 gradients of Θ and P , respectively, along a neutral surface. We calculate monthly neutral slopes
 478 and tracer gradients using the VENM algorithm^{35,61}, a vertically non-local method for
 479 calculating neutral slopes and tracer gradients based on a search algorithm. The terms Q_{sw} ,
 480 Q_{lw} , Q_{lh} , and Q_{sh} are the shortwave, longwave, latent, and sensible heat flux terms. $F_E + F_P$
 481 is the surface freshwater flux taken to be the difference between precipitation, F_P , and
 482 evaporation, F_E . The Dirac delta function, $\delta(z - \eta)$, is nonzero only at the ocean surface, $z =$
 483 $\eta(x, y, t)$, with units of inverse length. The penetration of shortwave radiation with depth,
 484 previously shown to be important for estimating WMT³⁴ (Extended Data Fig. 10), is taken as

485 $Q_{sw} F(z)$, where $F(z) = R e^{-\frac{z}{h_1}} + (1 - R) e^{-\frac{z}{h_2}}$ is an exponential decay function proposed by
 486 Paulson and Simpson⁶⁶, in which z is depth (positive downward), $R = 0.58$, $h_1 = 0.35$ m, and
 487 $h_2 = 23$ m.

488 The mesoscale neutral diffusivity in the ocean interior, K_N , and horizontal diffusivity in the
 489 mixed layer, K_H , are taken from Groeskamp et al.³⁶, who combined mixing length theory⁶⁷,
 490 surface mode theory⁶⁸, and mean-flow suppression theory⁶⁹ to estimate the full-depth
 491 mesoscale eddy diffusivity from World Ocean Atlas 2018 and altimeter measurements. The
 492 small-scale isotropic diffusion, D , is taken from de Lavergne et al.³⁷, obtained by applying a
 493 tidal mixing parameterization to an observational climatology of stratification, accounting for
 494 both bottom-intensified local mixing and stratification-dependent remote mixing, to estimate
 495 the effects of small-scale turbulent mixing. All three diffusivities are spatially varying but
 496 remain constant in time. Ocean mixed layer depth (MLD) is calculated as the depth at which
 497 the monthly-averaged potential density in the water column first exceeds the density at 10-m
 498 depth by 0.03 kg m^{-3} , following past work^{70,71}.

499 **Constructing the observation-based AMOC streamfunction.** For the water volume bounded
 500 by the iso-neutrals $\gamma^n - \Delta\gamma^n/2$ and $\gamma^n + \Delta\gamma^n/2$ (Extended Data Fig. 1), the outflow volume
 501 flux crossing the interior domain boundary ($\Delta\psi$) equals the convergence of WMT (ΔG) minus
 502 the temporal change in layer volume ($\partial(\Delta V)/\partial t$) from Eq. (1):

$$\underbrace{\Delta\psi}_{\text{outflow}} = \underbrace{\Delta G}_{\text{water mass formation}} - \underbrace{\frac{\partial(\Delta V)}{\partial t}}_{\text{volume tendency}}. \quad (4)$$

503 When applying Equation (4) across latitudes from 70°N to 35°S in the Atlantic basin (basin
 504 mask shown in Fig. 1), the outflow volume flux across each latitude (i.e., interior boundary)
 505 and within each density bin is estimated. Here $\Delta\psi > 0$ represents a southward flow leaving the
 506 domain; therefore, a minus sign is applied to Equation (4) to derive the corresponding

northward volume fluxes, $-\Delta\psi$. The Atlantic meridional overturning circulation (AMOC) streamfunction is then calculated as the cumulative northward volume transport, integrated from lighter to denser layers within the density range $21 < \gamma^n < 29 \text{ kg m}^{-3}$ and from 70° N southward, and apportioned by latitude^{9,28,30} (Fig. 3a):

$$-\psi = -G + \frac{\partial V}{\partial t} = -(G_F + G_{SW} + G_H + G_N + G_V) + \frac{\partial V}{\partial t}, \quad (5)$$

where $G_F + G_{SW}$ are the WMT due to air-sea buoyancy fluxes and shortwave radiation penetration, $G_H + G_N$ is the sum of the WMT due to mesoscale horizontal and neutral mixing, and G_V is small-scale vertical mixing (Extended Data Figs. 1, 3). Here ψ , G , and V are all integrated over the volume below $\gamma^n = 21 \text{ kg m}^{-3}$. Due to the 2000 m-depth limitation of the gridded SIO RG-Argo, the AMOC streamfunction in density layers below $\gamma^n = 27.8 \text{ kg m}^{-3}$ is excluded from discussion for this data, as these layers are not fully sampled within the available profiles. Since the overturning streamfunction is derived by closing the volume budget for a given density layer, the WMT-based streamfunction includes not only the buoyancy-driven thermohaline circulation but also other processes that contribute to volume tendencies within that layer. Note that temporal changes in layer volume south of 35° S are affected by zonal advection across the open zonal Atlantic basin boundaries. Accordingly, while we present the climatological mean overturning streamfunction south of 35° S to illustrate the full circuit of the AMOC upper and lower limbs, we recognize that its temporal variability also reflects net zonal volume advection across the basin boundaries.

Following previous studies^{3,4,19,27,40}, the reconstructed AMOC intensity is defined as the maximum overturning streamfunction within the intermediate and deep layers of the North Atlantic, based on the AMOC streamfunction derived from the WMT framework. To perform the WMT-based volume budget calculation and reconstruct the AMOC streamfunction, the ocean is discretized into density bins. In this study, we use a bin size of $\Delta\gamma^n = 0.05 \text{ kg m}^{-3}$. The

choice of bin size influences how well the vertical structure of WMT is captured and affects the estimated maximum streamfunction. Although finer bins suit for resolving sharp vertical gradients, such as those near the thermocline, they do not always yield better results – particularly when the bin size exceeds the vertical resolution of gridded ocean data, which can introduce numerical noise⁹. To mitigate this, we interpolate vertical profiles to 10-m intervals in the upper 2500 m and 50-m intervals below that depth. After testing a range of density bin sizes from 0.02 to 0.2 kg m⁻³, we chose $\Delta\gamma^n = 0.05$ kg m⁻³ with a 0.1 kg m⁻³ smoothing window for IAP, Argo, and Ishii data and a 0.2 kg m⁻³ smoothing window for EN4 data. This choice provides the optimum bin-size configuration to ensure sufficient resolution to capture the AMOC structure, whilst also factoring in the resolution and data coverage of each product, thereby yielding consistent reconstructions across datasets.

Definition of AMOC fingerprints

SST fingerprint. The SST-based ‘warming hole’ AMOC fingerprint was originally proposed by Rahmstorf et al.¹ and is calculated here following the latest version as employed by Caesar et al.², using the November-May average of SST from the HadISST, ERSSTv5, and ERA5 datasets. First we calculate the SST trend at each grid point over the full data period and identify a persistent surface cooling in the subpolar North Atlantic south of Greenland¹ (Fig. 2). We then compute the area-weighted mean SST over the cooling region in the subpolar North Atlantic for each of the SST datasets. The ‘warming hole’ index is obtained as the SST anomaly over the subpolar North Atlantic relative to the global mean SST² (Fig. 2c). Motivated by this SST-based index, we extend the approach to subsurface depths by repeating the calculations using vertically averaged conservative temperature over the upper 5 m and 1000 m of the ocean, respectively, based on the IAPv4.2 data. The upper 5m depth-averaged temperature is used as a proxy for SST in the IAPv4.2 climatology, as that data set does not include SST separately.

SSS fingerprint. The ‘salinity pileup’ index was originally proposed by Zhu and Liu³ and is calculated as the difference between the annual mean SSS averaged over the subtropical South Atlantic between 10°S and 34°S, and the annual mean SSS averaged over the subtropical South Indo-Pacific between 10°S and 34°S (Fig. 2f). The annual mean SSS is taken from the IAPv4.2, EN4-ESM, and Ishii datasets, using the absolute salinity at 5 m depth.

Data Availability

The processed observational-based data used to recreate the figures presented in this study have been deposited online in the Zenodo database under (*will be published online upon acceptance for publication*). The gridded SIO RG-Argo product is available at http://sio-argo.ucsd.edu/RG_Climatology.html. These data were collected and made freely available by the International Argo Program and the national programs that contribute to it (<http://www.argo.ucsd.edu>, <https://www.ocean-ops.org>). The Argo Program is part of the Global Ocean Observing System. The gridded IAP temperature and salinity data is available at <http://www.ocean.iap.ac.cn/>. The latest version of EN4 data (EN4.2.2) is available from <https://www.metoffice.gov.uk/hadobs/en4/download-en4-2-2.html>. The World Ocean Atlas 2023 data is available from <https://www.ncei.noaa.gov/products/world-ocean-atlas>. The latest version of Ishii data (version 7.3.1) can be downloaded at <https://climate.mri-jma.go.jp/pub/ocean/ts/v7.3.1/>. The GSW Oceanographic Toolbox is available at <http://www.teos-10.org/software.htm>. PreTEOS-10 software is available online at http://www.teos-10.org/preteos10_software/neutral_density.html. The ERA5 reanalysis data can be obtained from <https://www.ecmwf.int/>. Data from the RAPID AMOC observing project is funded by the Natural Environment Research Council, U.S. National Science Foundation (NSF) with support from NOAA. These data are available from <https://rapid.ac.uk/>. OSNAP data were collected and made available by the OSNAP (Overturning in the Subpolar North

Atlantic Program) project with support from the national programs that contribute to it (www.o-snap.org). Data from the SAMBA array can be obtained from https://www.aoml.noaa.gov/phod/SAMOC_international/samoc_data.php. The gridded bathymetry data (GEBCO_2025 Grid) can be obtained from <https://www.gebco.net/data-products-gridded-bathymetry-data/gebco2025-grid>.

Code Availability

All analyses and figures shown in this manuscript are reproducible via MATLAB. Codes needed to process and reproduce all diagrams presented in this study will be published online in the Zenodo database upon acceptance for publication.

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Author Contributions

M.H.E., Z.L., and S.G. conceived and designed the study. Z.L. and S.G. developed the WMT approach to reconstruct AMOC streamfunction. Z.L. undertook the calculations and data analysis and produced all the graphics used in the study. All authors contributed to the research direction of the study by providing scientific advice and help with interpretation of results

603 through all stages of the project. Z.L. wrote the first draft of the paper. All authors contributed
604 to refinement of the paper and interpretation of the results.

605

606 **Competing Interests**

607 The authors declare no competing interests.

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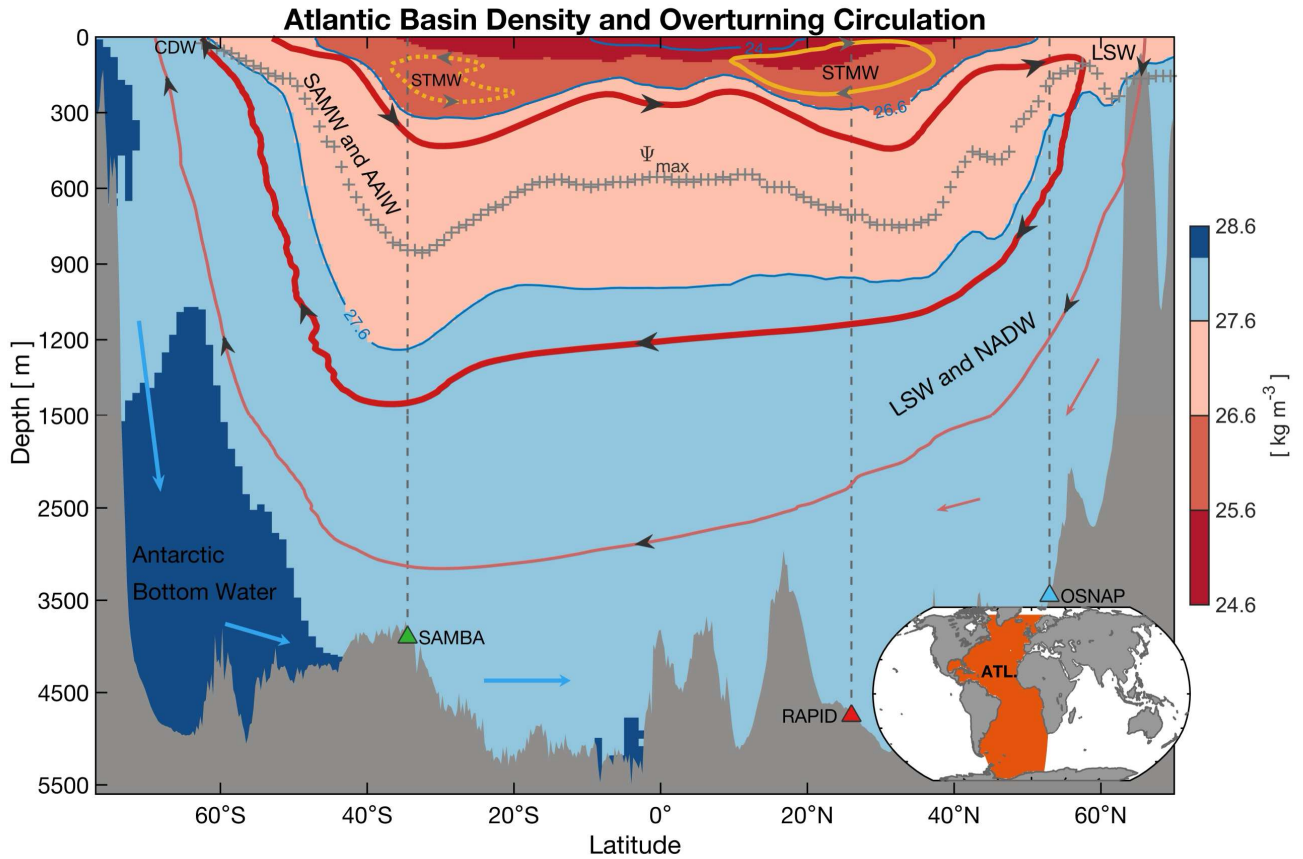


Figure 1 | Meridional transect of Atlantic basin density and overturning circulation. Zonally averaged annual mean neutral density (kg m^{-3}) across the Atlantic basin during 2004–2023, based on the IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets. Blue contours represent neutral density surfaces $\gamma^n = 24.0, 26.6$, and 27.6 kg m^{-3} ; abbreviations indicate Subtropical Mode Water (STMW), Subantarctic Mode Water (SAMW), Antarctic Intermediate Water (AAIW), Circumpolar Deep Water (CDW), Labrador Sea Water (LSW), and North Atlantic Deep Water (NADW). Superimposed reconstructed streamfunction contours highlight the overturning: solid and dashed yellow cells represent the clockwise subtropical cell in the North Atlantic and anticlockwise subtropical cell in the South Atlantic, respectively; the red bold contour (9 Sv of the meridional overturning streamfunction) illustrates the northward flow of intermediate waters and the southward return flow of LSW and NADW, which together form the upper and lower branches of the AMOC; the thin red contour (4 Sv) and light red arrows depict the southward flow of LSW and NADW that constitutes the lower limb. Grey crosses mark the depth of maximum overturning streamfunction within the intermediate and deep-water layer at each latitude; light blue arrows indicate the circulation pathway of Antarctic Bottom Water. Vertical dashed grey lines depict the SAMBA array at 34.5°S , the RAPID array at 26°N , and the approximate latitude of the OSNAP array at 53°N , though OSNAP moorings span $53^\circ\text{--}60^\circ\text{N}$. Bathymetry is averaged between 20°W and 25°W near the A16 transects. The inset in the bottom right shows the Atlantic basin mask used in this study (orange-red shading).

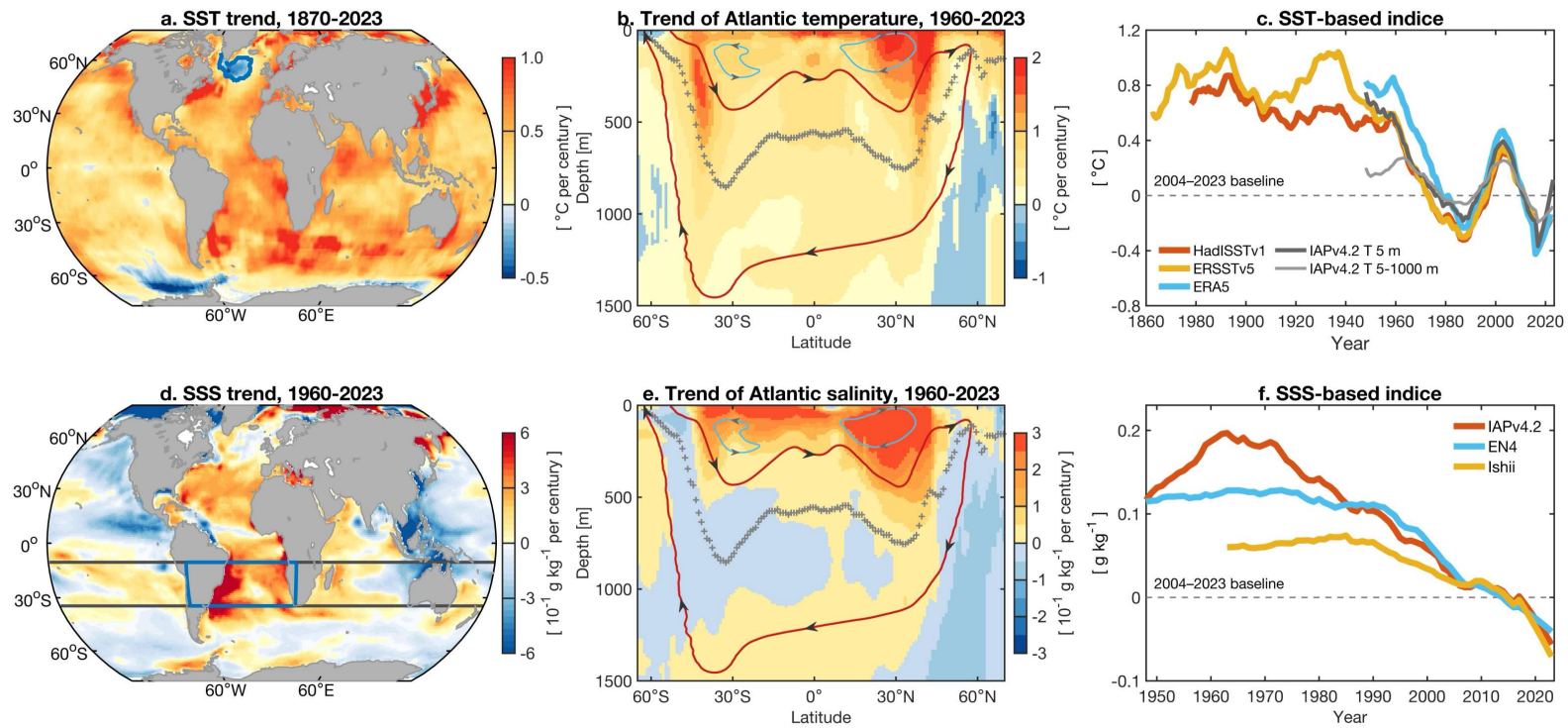


Figure 2 | Sea surface temperature (SST)- and salinity (SSS)-based AMOC fingerprints.
a, Linear trends of annual SST ($^{\circ}\text{C}$ per century) over 1870–2023 from HadISST. Blue contours indicate the ‘cold blob’ in the subpolar North Atlantic, where long-term SST trends are negative.
b, Linear trends of zonally averaged annual mean temperature ($^{\circ}\text{C}$ per century) across the Atlantic basin over 1960–2023 from IAPv4.2. **c**, Time series of the SST-based AMOC index^{1,2} ($^{\circ}\text{C}$) from HadISST (red), ERSST (yellow), ERA5 (blue), and subsurface temperature fields from IAPv4.2 (grey), relative to the 2004–2023 mean. Time series are shown as 15-year smoothed November–May averages, with the last 7 years smoothed using a locally weighted linear regression (‘lowess’) approach. **d**, **e**, Linear trends of annual SSS ($10^{-1} \text{ g kg}^{-1}$ per century) and zonally averaged annual mean salinity ($10^{-1} \text{ g kg}^{-1}$ per century) across the Atlantic basin, over 1960–2023 from IAPv4.2. Superimposed reconstructed streamfunction contours are taken from Fig. 1 to highlight the overturning: light blue cells represent the subtropical cells; the red contour illustrates the northward flow of intermediate waters and the southward return flow of LSW and NADW. **f**, Time series of the SSS-based AMOC index³ (g kg^{-1}) from IAPv4.2 (red), EN4-ESM (blue), and Ishii (yellow), relative to the 2004–2023 mean. Details of the definitions of these AMOC indices are given in the Methods.

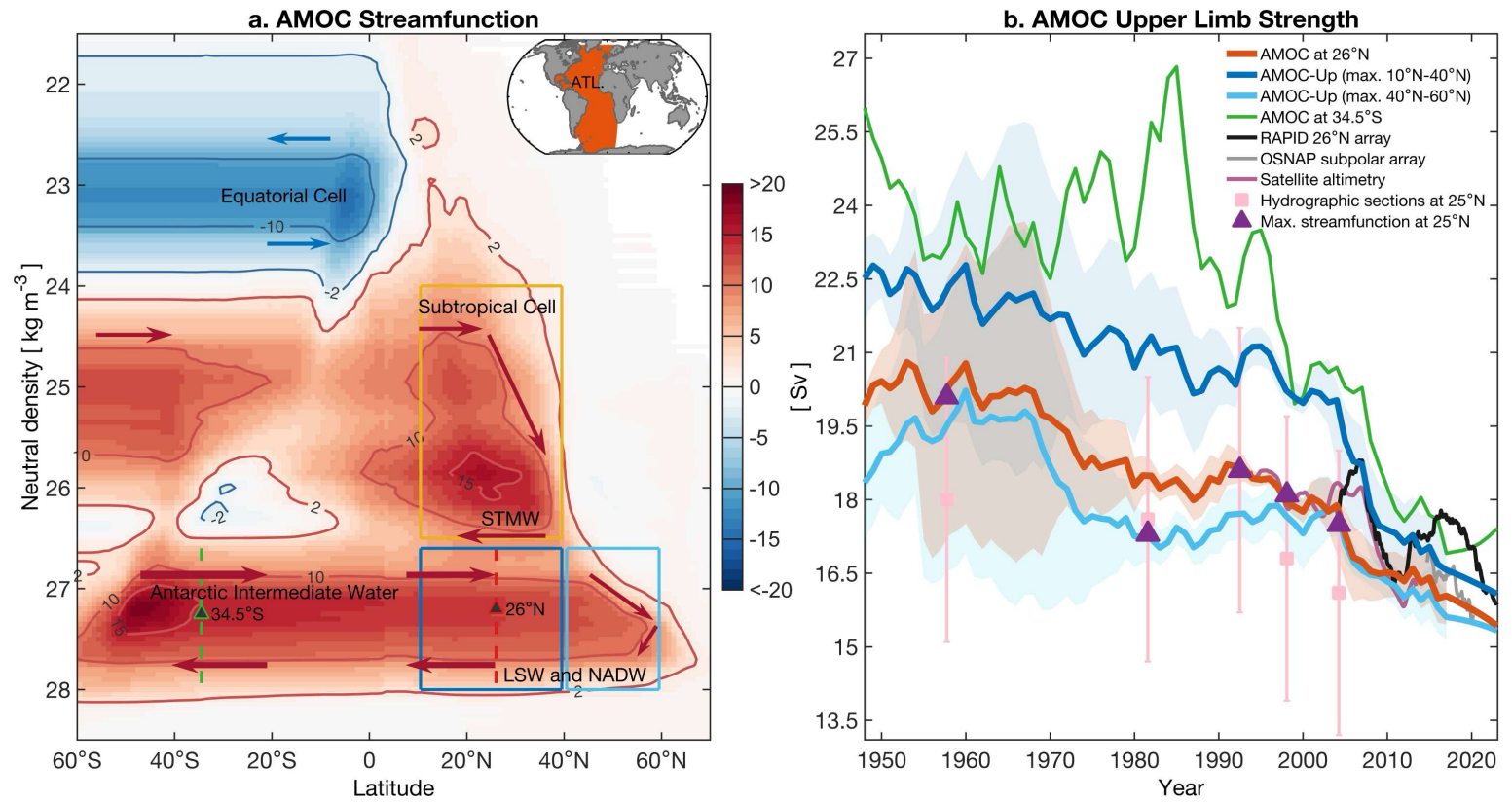


Figure 3 | Structure and time evolution of the Atlantic meridional overturning circulation (AMOC). **a**, Ensemble-mean AMOC streamfunction ($\text{Sv} = \text{Sverdrup} = 10^6 \text{ m}^3 \text{ s}^{-1}$) for the period 2004–2023, reconstructed using the WMT framework from the IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets (Methods). Superimposed contours show streamfunction values of -10 , -2 , 2 , 10 , and 15 Sv . The inset in the top right shows the Atlantic basin mask used in this study (orange-red shading). **b**, Ensemble-averaged AMOC strength (Sv) shown as annual means (thick lines with shadings indicating one standard deviation across datasets), derived from IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo. Red line shows the AMOC strength at 26°N ; green line shows the AMOC strength at 34.5°S ; dark blue line shows the AMOC upper-limb strength within 10°N – 40°N ; light blue line shows the subpolar AMOC strength within 40°N – 60°N . All reconstructed lines represent 15-year running averages, with the last 7 years smoothed using a locally weighted linear regression (‘lowess’) approach. Also shown are the AMOC strength observed by the RAPID array at 26°N (black line; 5-year running mean) and the OSNAP array in the subpolar North Atlantic (grey line; 5-year running mean). Pink squares represent ship-based hydrographic estimates of the integrated transport over the upper 1100 m at 25°N , with pink bars indicating the 2.9 Sv standard deviation from RAPID observations over 2004–2009, as reported in Table 5 of ref.²⁰. Purple triangles indicate hydrographic estimates of the maximum streamfunction at 25°N ²². The purple line shows the AMOC estimate derived from satellite altimetry for 1993–2014²⁴.

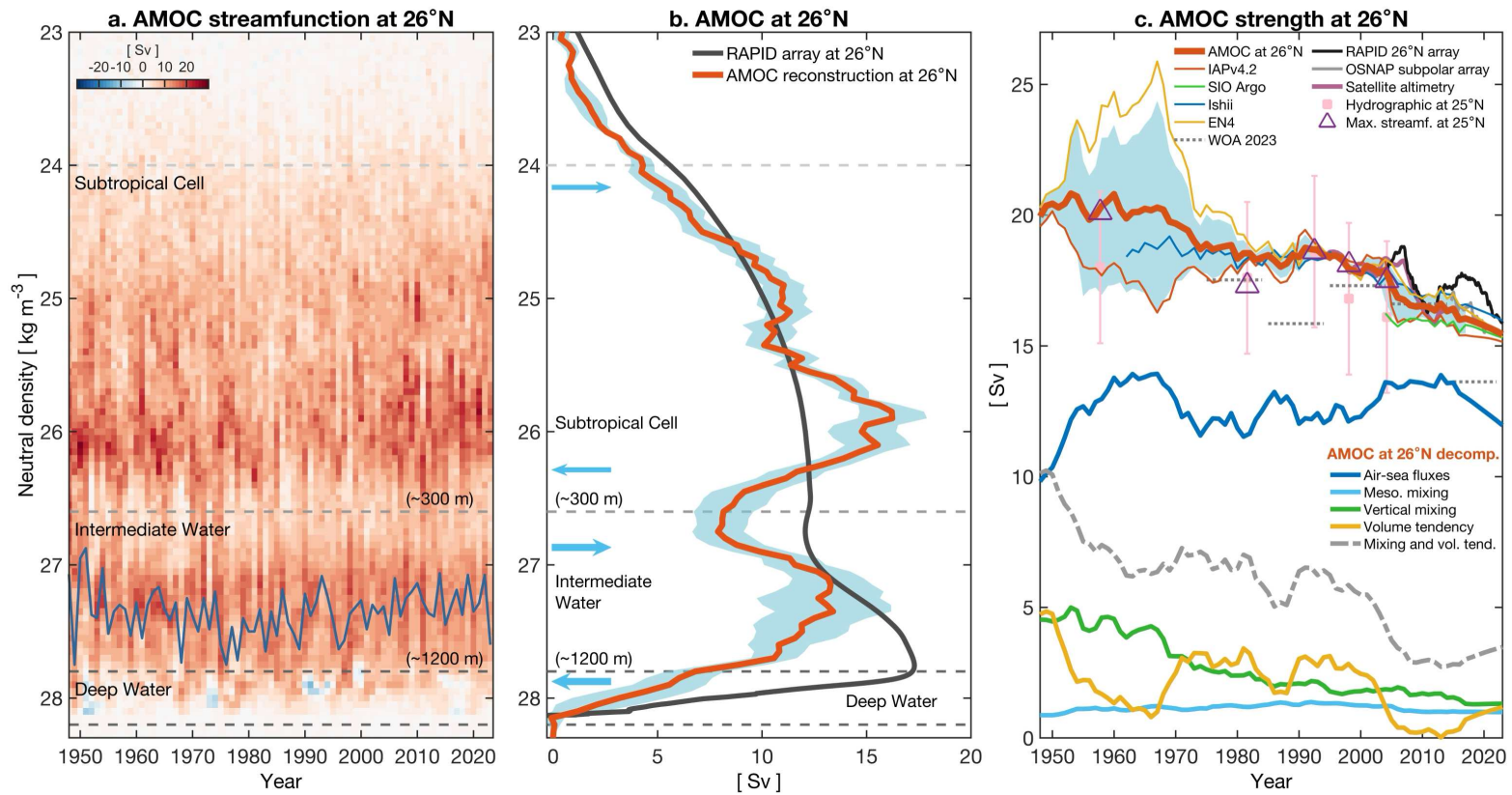


Figure 4 | Reconstructed AMOC at 26°N and its air-sea flux and subsurface components.

a, Hovmöller diagram of reconstructed annual mean AMOC streamfunction (Sv) at 26°N in density space. Blue line indicates the neutral density layer with the maximum streamfunction.

b, Vertical structure of the reconstructed AMOC streamfunction at 26°N using a density bin size of 0.05 kg m^{-3} from IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo. The black line shows the climatological mean overturning streamfunction observed by the RAPID array at 26°N. **c**, Reconstructed annual mean AMOC strength at 26°N (Sv) for IAPv4.2 (thin red), EN4-ESM (yellow), SIO RG-Argo (green), and Ishii (blue), with the ensemble mean shown as a thick red line and associated one standard deviation across datasets as blue shading. The reconstructed AMOC strength at 26°N is further decomposed into contributions from air-sea fluxes (dark blue), mesoscale mixing (light blue), vertical mixing (green), and upper-limb volume tendency (yellow). The dashed grey line indicates the combined contribution from all subsurface processes (mesoscale mixing, vertical mixing, and upper-limb volume tendency). All reconstructed lines represent 15-year running averages, with the last 7 years smoothed using a locally weighted linear regression ('lowess') approach. Decadal mean AMOC strength estimates at 26°N from WOA 2023 are shown in dashed grey for 1975–1984, 1985–1994, 2005–2014, and 2015–2022. Observational estimates include hydrographic section transports (pink squares; ref.²⁰), hydrographic streamfunction maxima (purple triangles; ref.²²), satellite altimetry (purple line; ref.²⁴), and continuous observations from RAPID (black line) and OSNAP (grey line), as in Fig. 3b.

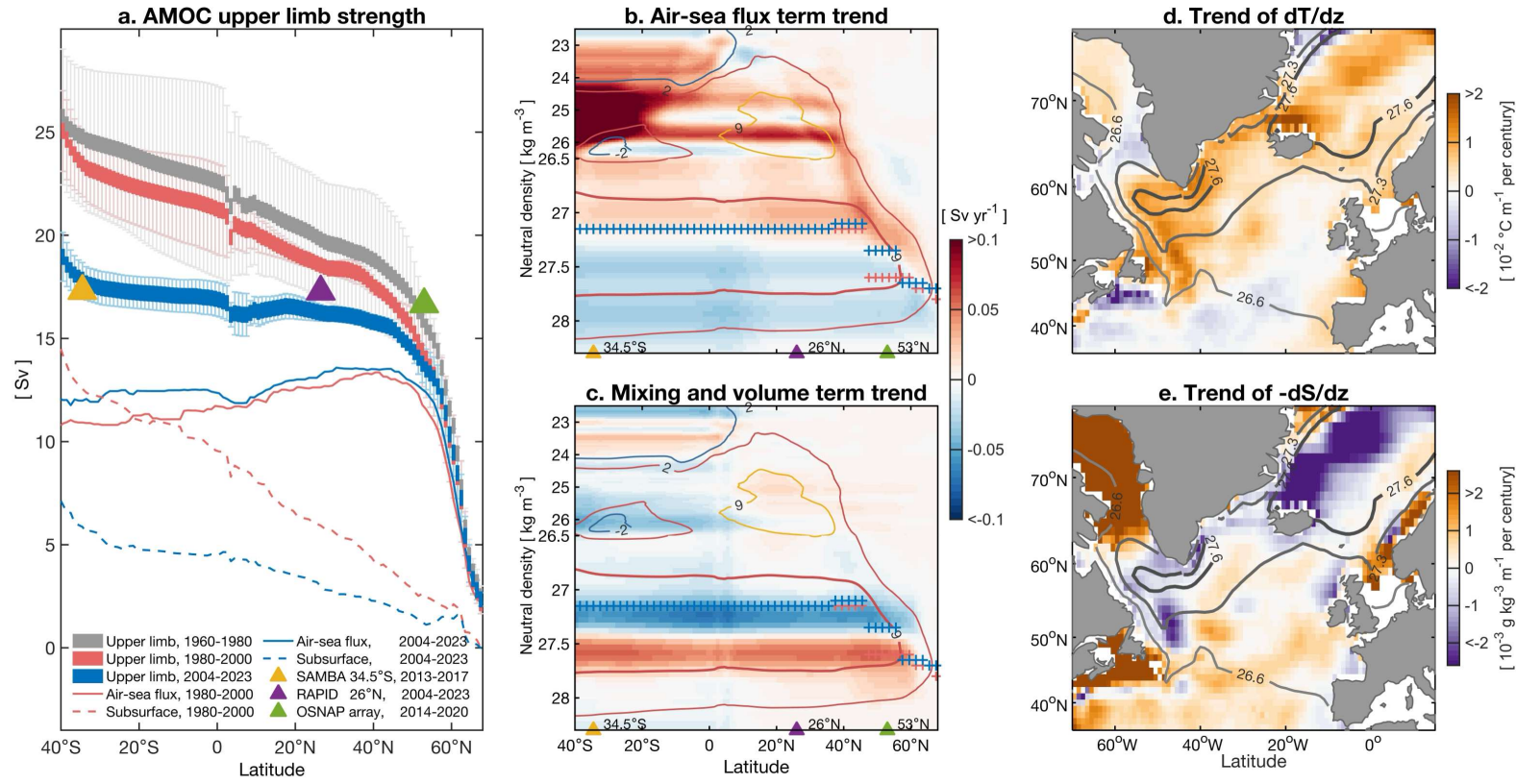
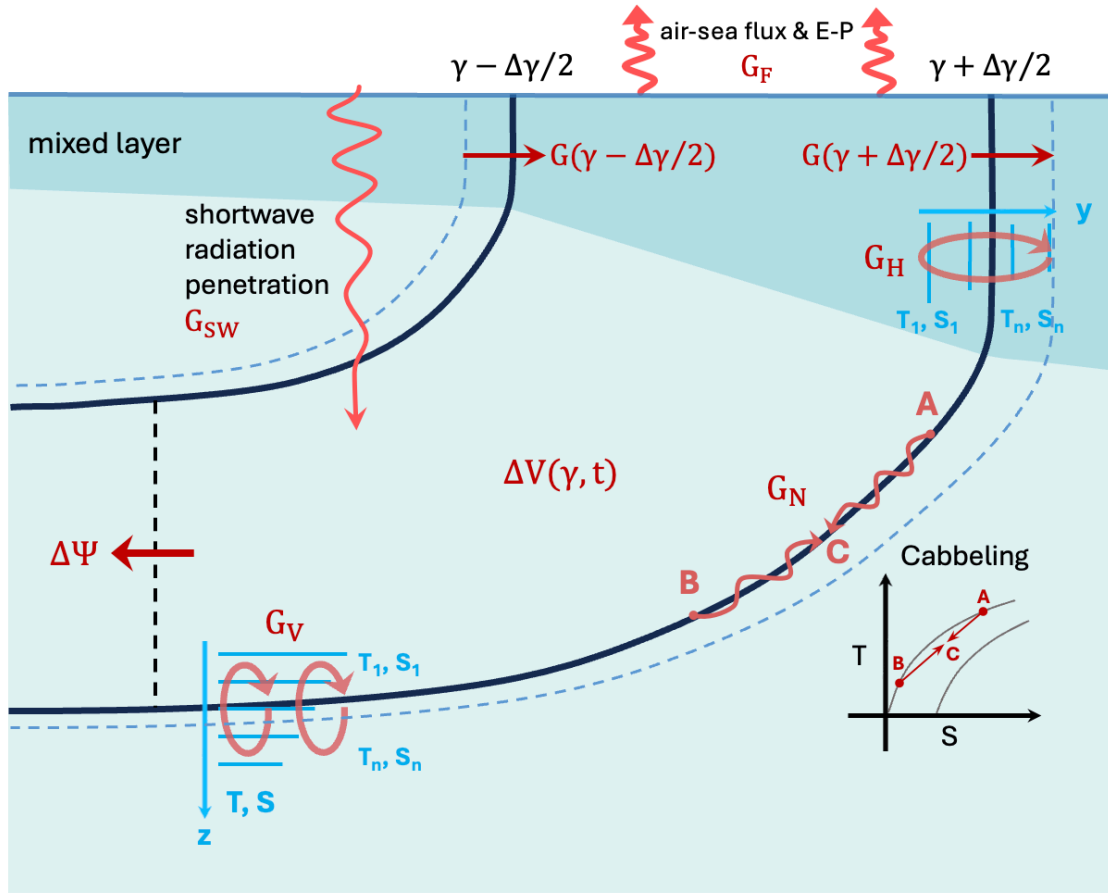
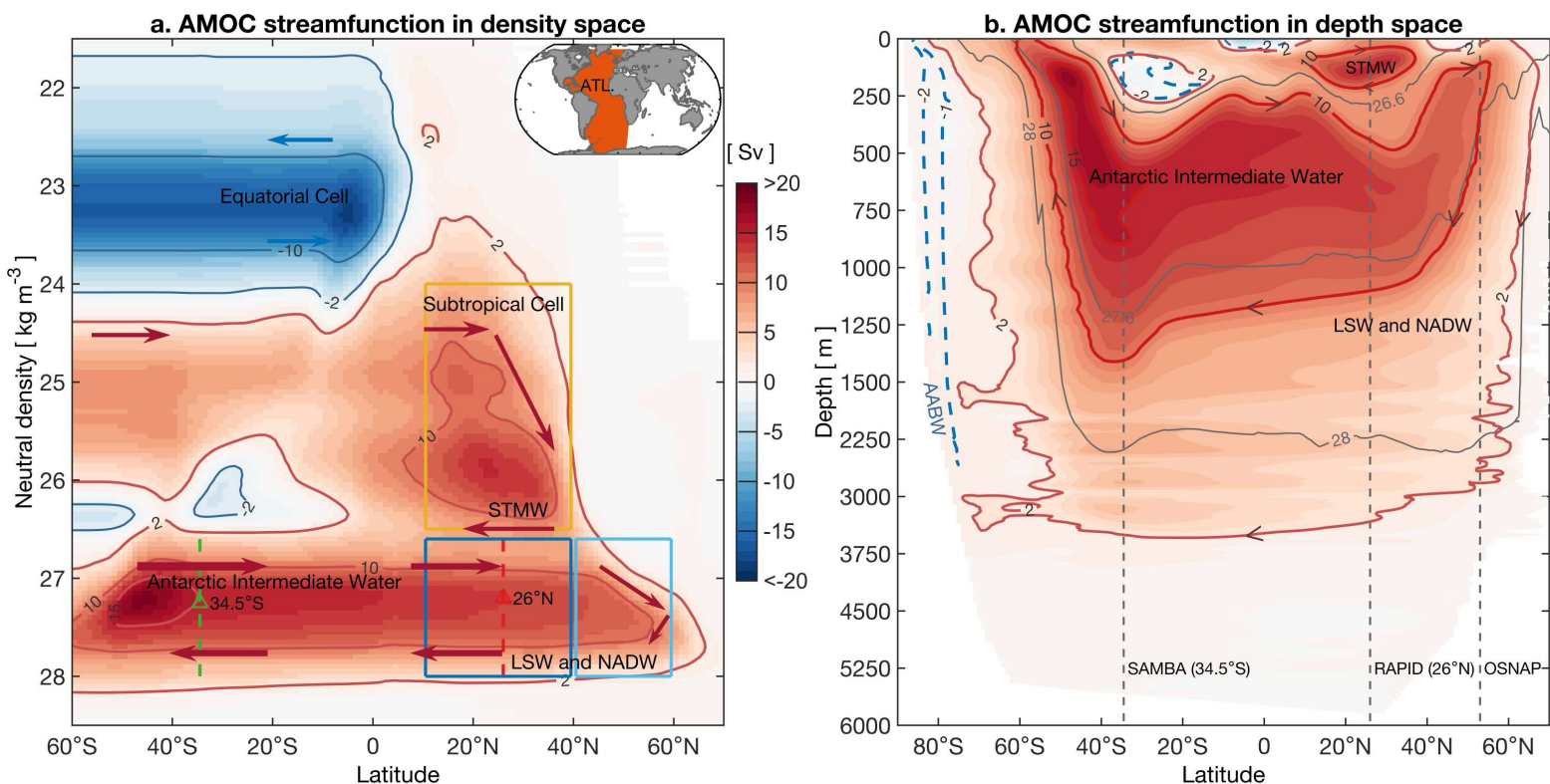


Figure 5 | AMOC strength changes and component breakdown. **a**, Ensemble-averaged annual mean AMOC strength (Sv), defined as the maximum intermediate and deep-layer streamfunction at each latitude, for the periods 1960–1980 (grey), 1980–2000 (red), and 2004–2023 (blue). The ensemble means and associated one standard deviation (bars) are based on IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets. The purple triangle represents the AMOC strength observed by the RAPID array at 26°N (April 2004–February 2023); the green triangle shows the OSNAP subpolar AMOC strength (2014–2020); and the yellow triangle shows the SAMOC strength observed by the SAMBA array at 34.5°S (September 2013–July 2017). Solid lines show the air-sea flux term for 1980–2000 (red) and 2004–2023 (blue); dashed lines show the sum of subsurface terms (mixing and upper-limb volume tendency) for the same periods. **b–c**, Long-term trends in reconstructed streamfunction from 1960 to 2023 decomposed into contributions from **(b)** air-sea fluxes and **(c)** ocean mixing and volume tendency terms. Crosses mark the layer of maximum streamfunction for 1980–2000 (red) and 2004–2023 (blue). **d–e**, Long-term trends in upper 100-m **(d)** ocean temperature stratification and **(e)** salinity stratification during 1960–2023. Grey contours represent neutral density surfaces $\gamma^n = 26.6, 27.3$, and 27.6 kg m^{-3} averaged over the upper 100-m.



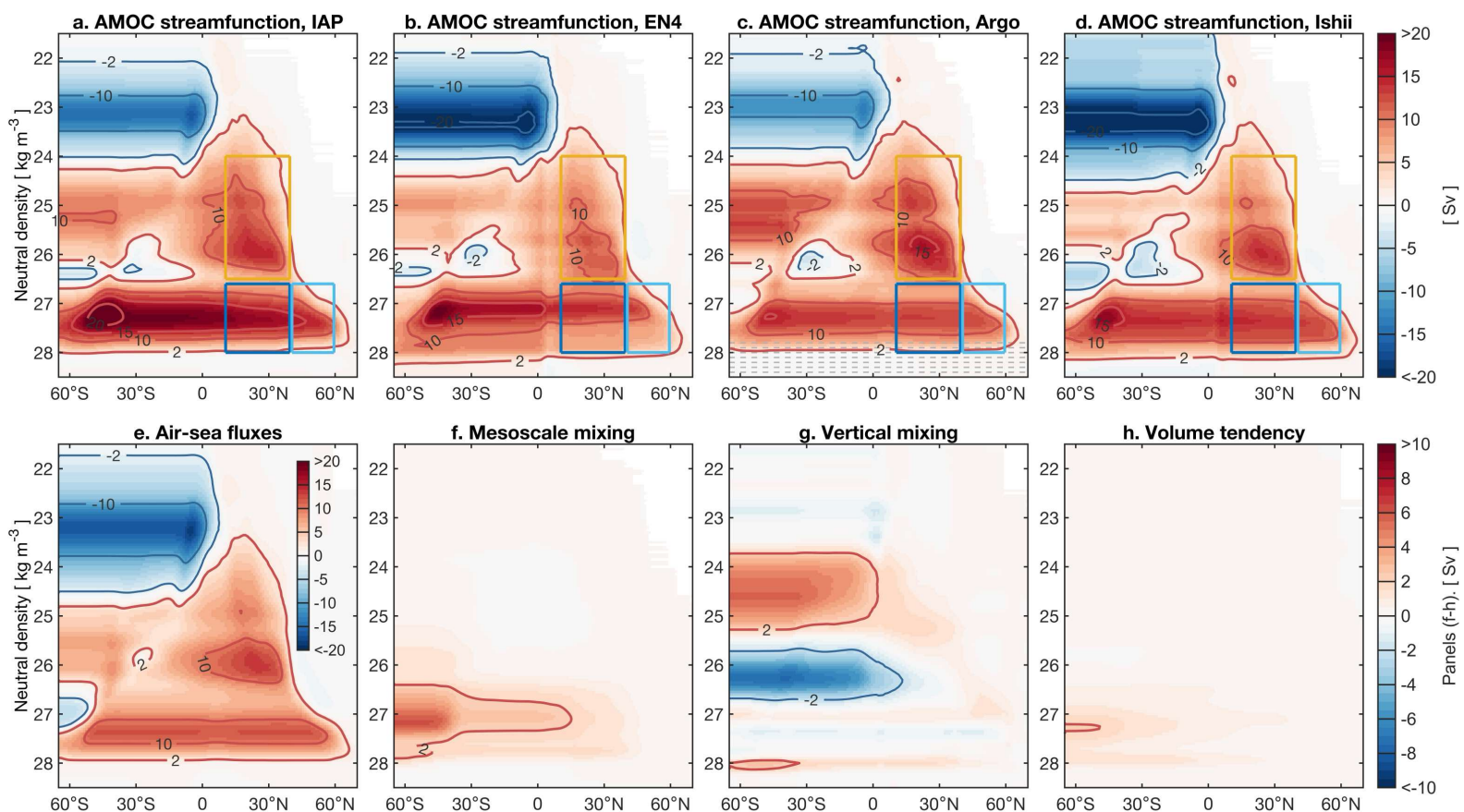
$$\Delta\psi = (\Delta G_F + \Delta G_{SW} + \Delta G_H + \Delta G_N + \Delta G_V) - \frac{\partial(\Delta V)}{\partial t}$$

Extended Data Figure 1 | Illustration of the water-mass transformation (WMT) framework, adapted from Fig. 3 in Groeskamp et al.⁹ ΔV represents the volume of water within $\gamma^n \pm \Delta\gamma^n/2$, bounded by the ocean surface and an interior boundary. Red arrows indicate volume transport across the interior boundary ($\Delta\psi$) and across neutral density surfaces ($G(\gamma^n - \Delta\gamma^n/2)$, $G(\gamma^n + \Delta\gamma^n/2)$) driven by multiple physical processes: air-sea heat and freshwater fluxes (G_F ; upward arrows), shortwave radiation penetration (G_{SW} ; downward arrows), horizontal mixing (G_H ; horizontal circles), mesoscale mixing along neutral density surfaces (G_N ; arrows along neutral surfaces), and small-scale vertical mixing (G_V ; vertical circles). Dashed blue lines indicate the ΔV varies with time and density. $\Delta G = G(\gamma^n - \Delta\gamma^n/2) - G(\gamma^n + \Delta\gamma^n/2)$ is the convergence of G , which quantifies the water mass formation (or destruction) within the layer bounded by $\gamma^n \pm \Delta\gamma^n/2$, so that $\Delta G > 0$ means that volume is added to the budget domain. The volume budget equation shown below the schematic corresponds to Equation (5).

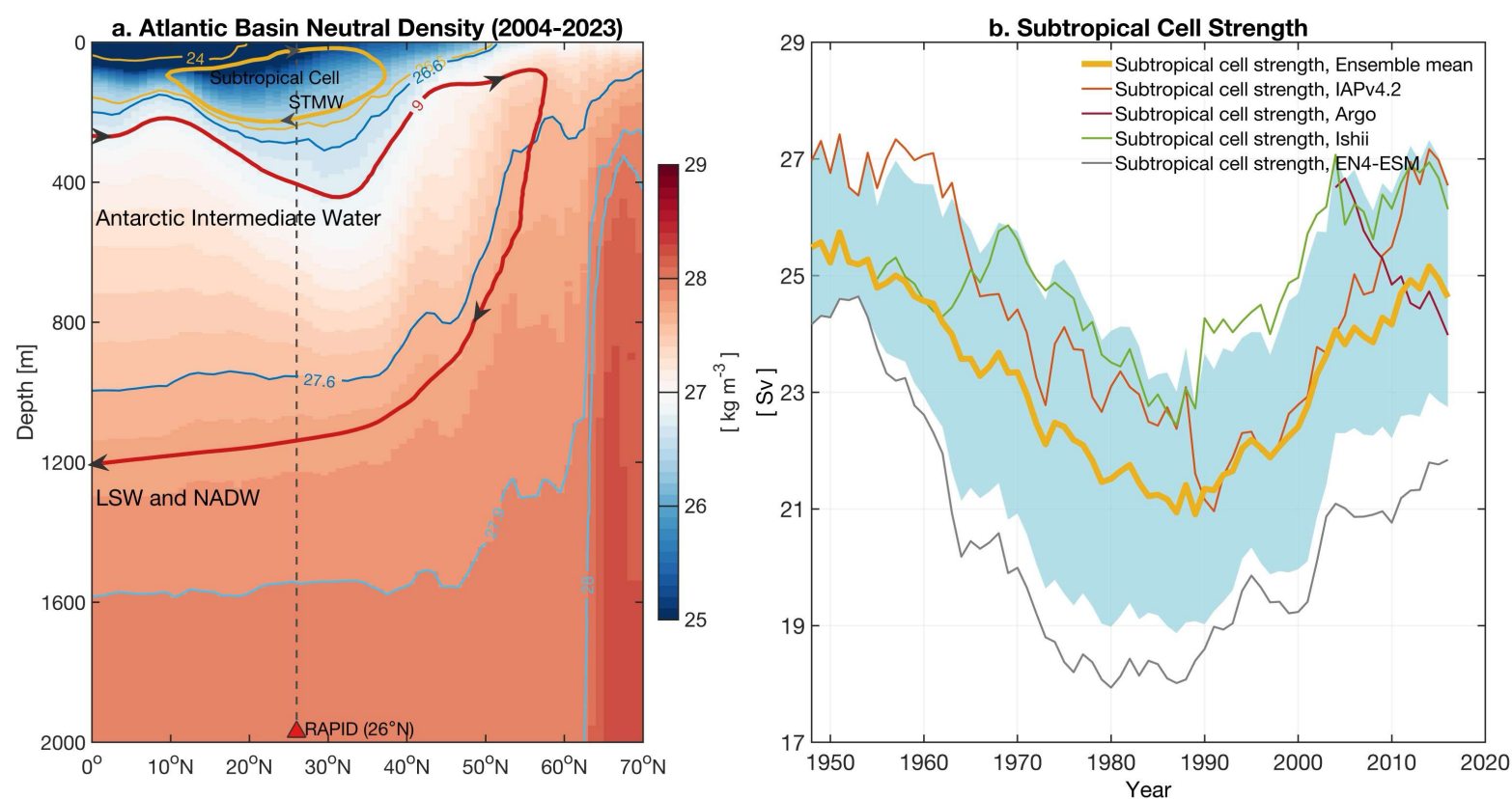


Extended Data Figure 2 | AMOC streamfunction in density and depth coordinates.

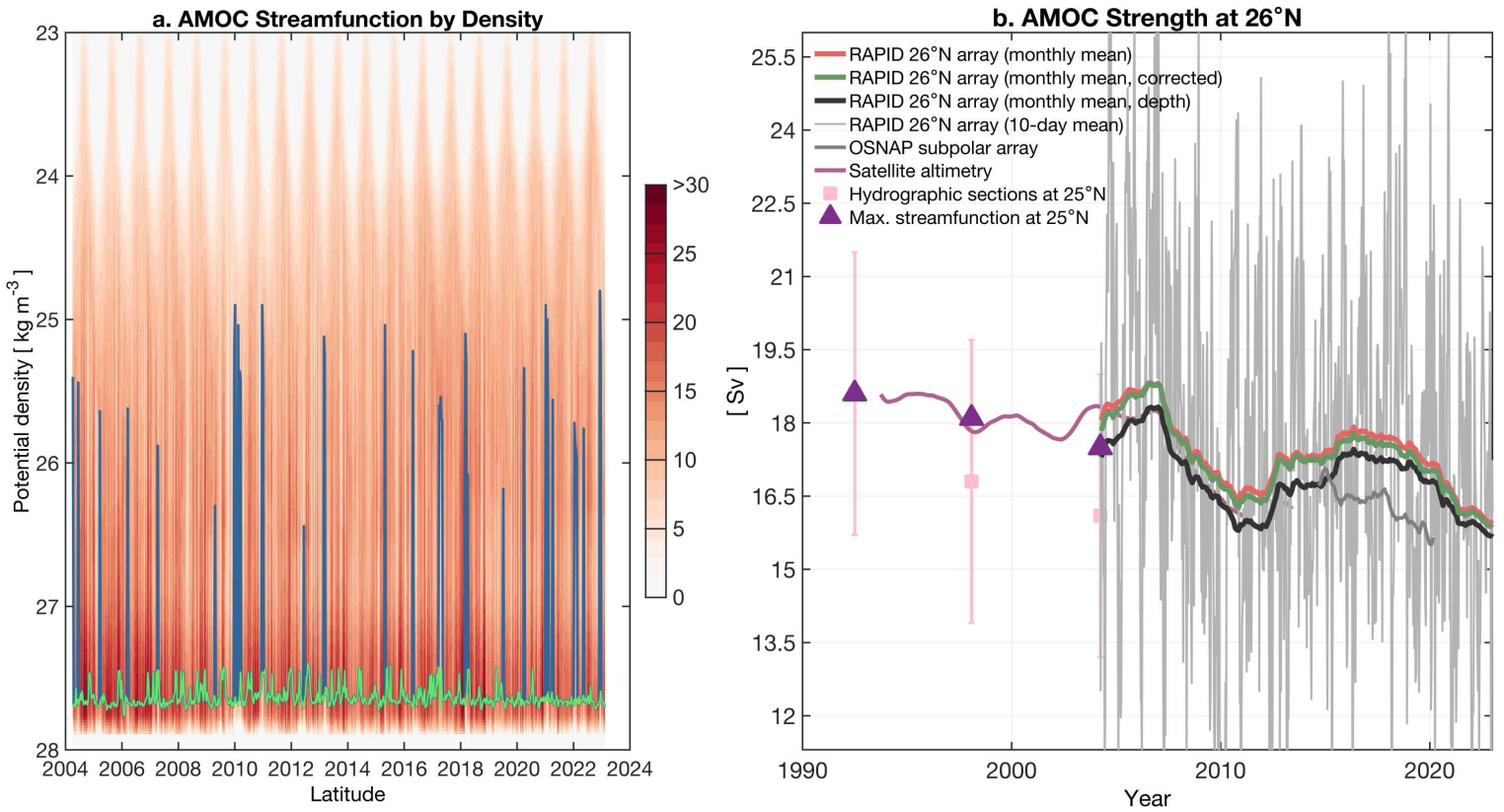
Ensemble-averaged reconstructed AMOC streamfunction (Sv) during 2004–2023 from the IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets (Methods), shown in (a) density coordinates (same as Fig. 3a) and (b) depth space remapped from density coordinates using the zonal mean depth of each density layer. Superimposed contours show streamfunction values of $-10, -2, 2, 10, \text{ and } 15 \text{ Sv}$. The inset in (a) shows the Atlantic basin mask used in this study (orange-red shading). Blue contours in (b) represent neutral density surfaces $\gamma^n = 24.0, 26.6, 27.6, \text{ and } 28.0 \text{ kg m}^{-3}$.



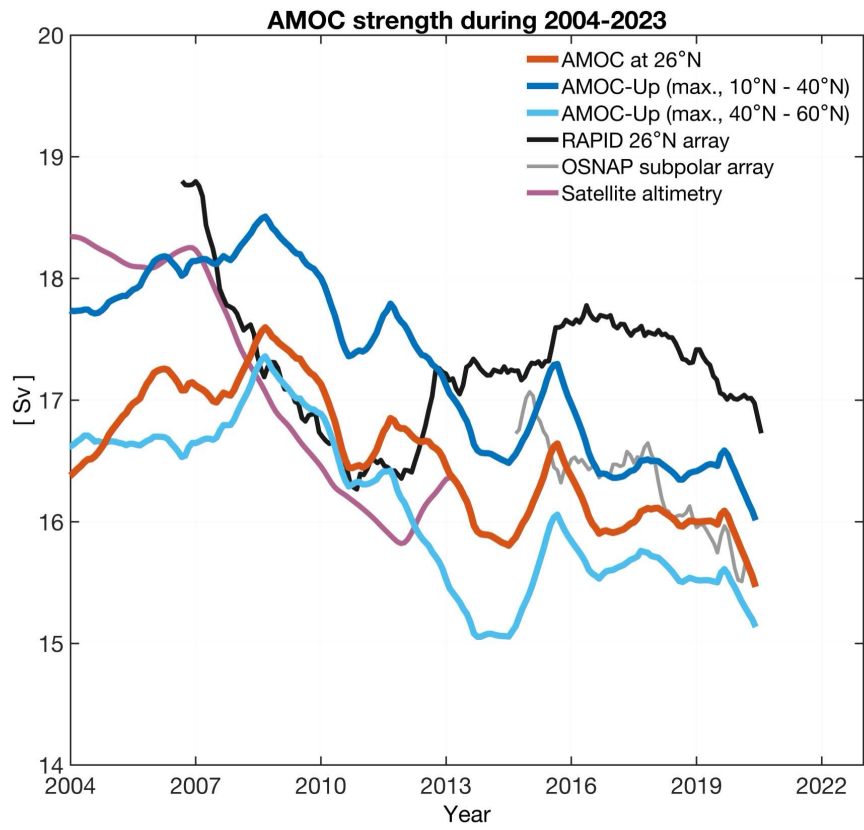
Extended Data Figure 3 | AMOC streamfunction and its components. **a–d**, Annual mean AMOC streamfunction (Sv) reconstructed from **(a)** IAP, **(b)** EN4-ESM, **(c)** SIO RG-Argo, and **(d)** Ishii datasets for the period 2004–2023. **e–h**, Ensemble-averaged annual mean AMOC streamfunction (Sv), decomposed into contributions from **(e)** air-sea buoyancy fluxes and shortwave radiation penetration, **(f)** mesoscale neutral mixing and horizontal mixing, **(g)** small-scale vertical mixing, and **(h)** volume tendency. Superimposed contours in all panels show streamfunction values of -10 , -2 , 2 , 10 , and 15 Sv; dashed grey lines in **(c)** indicate the neutral density layers below $\gamma^n = 27.8 \text{ kg m}^{-3}$ that are not fully sampled by traditional Argo float profiles.



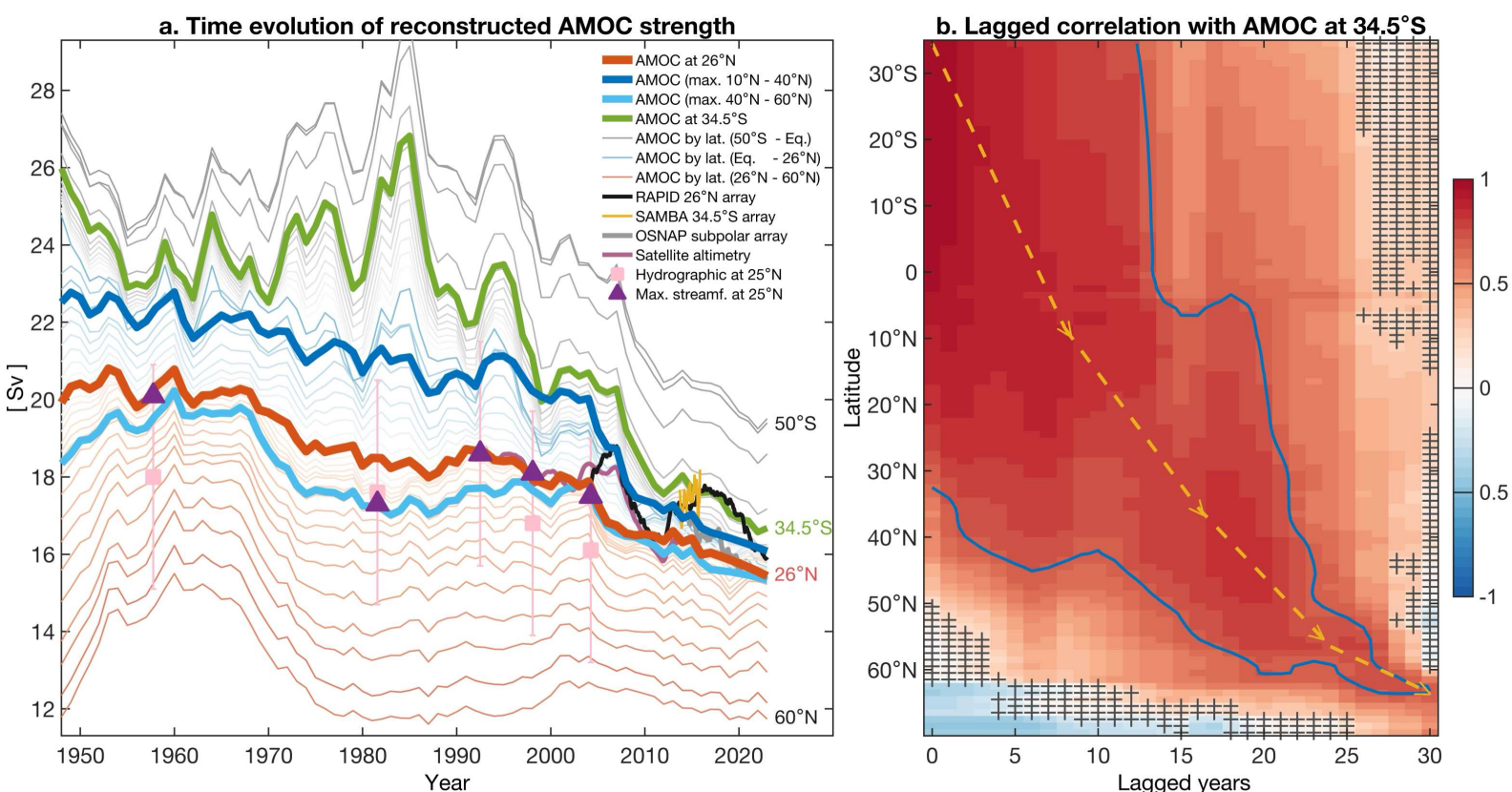
915 **Extended Data Figure 4 | Strength of the subtropical overturning cell. a**, Annual mean
 916 zonally averaged neutral density (kg m^{-3}) in the North Atlantic, based on IAPv4.2 data for the
 917 period 2004–2023. Superimposed contours represent neutral density surfaces $\gamma^n = 24.0, 26.5,$
 918 $26.6, 27.6, 27.9,$ and 28.0 kg m^{-3} ; abbreviations indicate Subtropical Mode Water (STMW),
 919 Labrador Sea Water (LSW), and North Atlantic Deep Water (NADW). Yellow contour
 920 highlights the North Atlantic subtropical cell; the thick red contour (9 Sv of reconstructed
 921 streamfunction) illustrates the northward pathway of intermediate waters and the southward
 922 return flow of LSW and NADW. **b**, Ensemble-averaged strength of the subtropical cell (Sv;
 923 dark-blue line) and associated one standard deviation across datasets (blue shading), based on
 924 IAPv4.2 (red), EN4-ESM (grey), Ishii (green), and SIO RG-Argo (purple) datasets. All
 925 reconstructed lines represent 15-year running averages. The subtropical cell strength is defined
 926 as the maximum streamfunction within 10°N – 40°N and above the neutral density surface $\gamma^n =$
 927 26.6 kg m^{-3} (yellow box in **Fig. 3a**).
 928



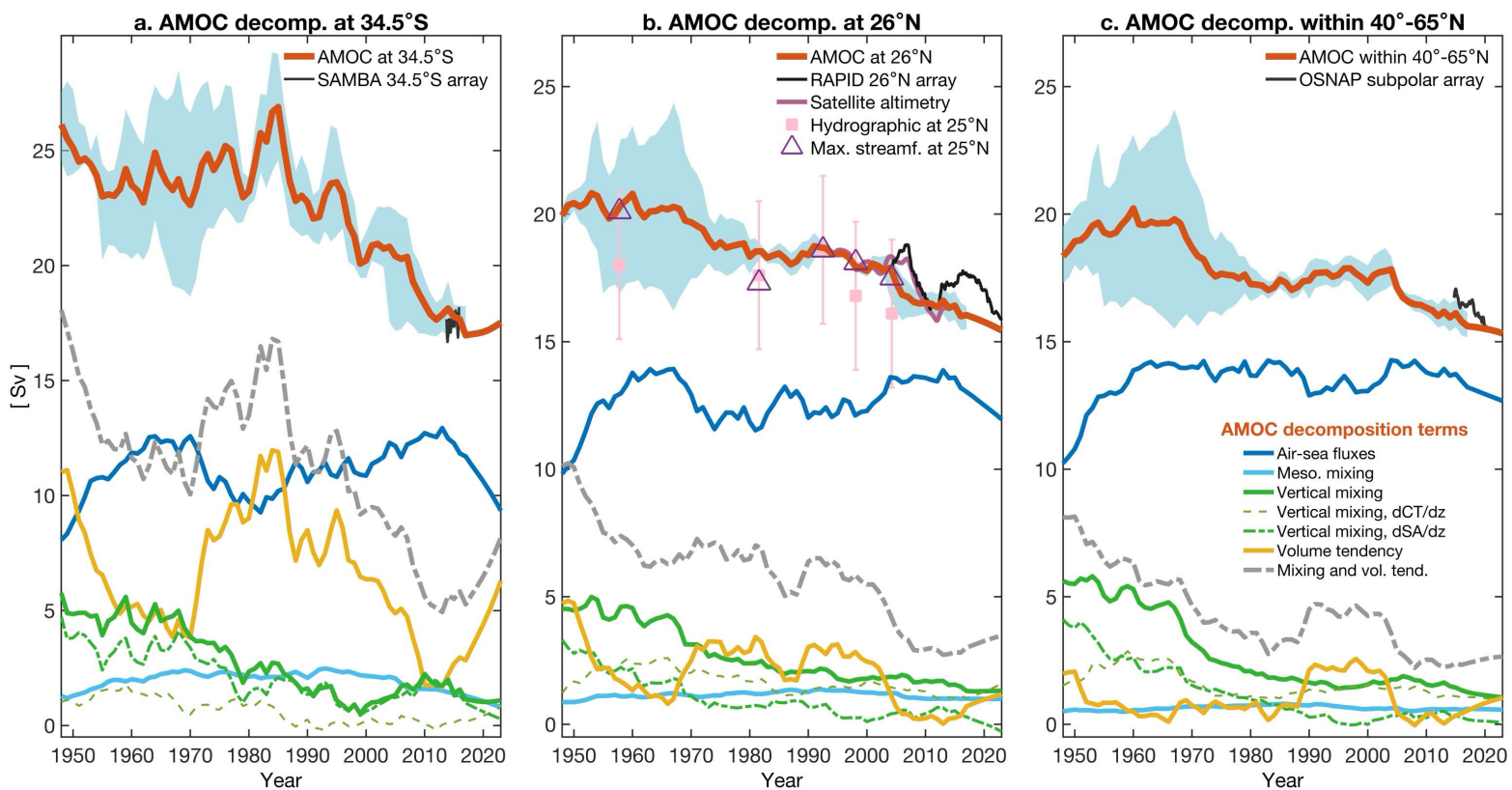
Extended Data Figure 5 | AMOC strength observed by the RAPID array at 26°N. a, 10-day averaged overturning streamfunction (Sv) as a function of density. The blue line indicates the density layer with the maximum streamfunction across the full profile; green line indicates the layer with the maximum intermediate and deep-layer streamfunction below $\sigma_0 = 26.6 \text{ kg m}^{-3}$. **(b)** Same as **Fig. 4c**, but showing the observed AMOC strength at 26°N (Sv), defined by the maximum streamfunction in depth coordinates (black line), in density coordinates (red line), and in the intermediate and deep-water layer below $\sigma_0 = 26.6 \text{ kg m}^{-3}$ (green line). Thick lines represent 5-year smoothed monthly values; thin grey lines show the original 10-day averages.



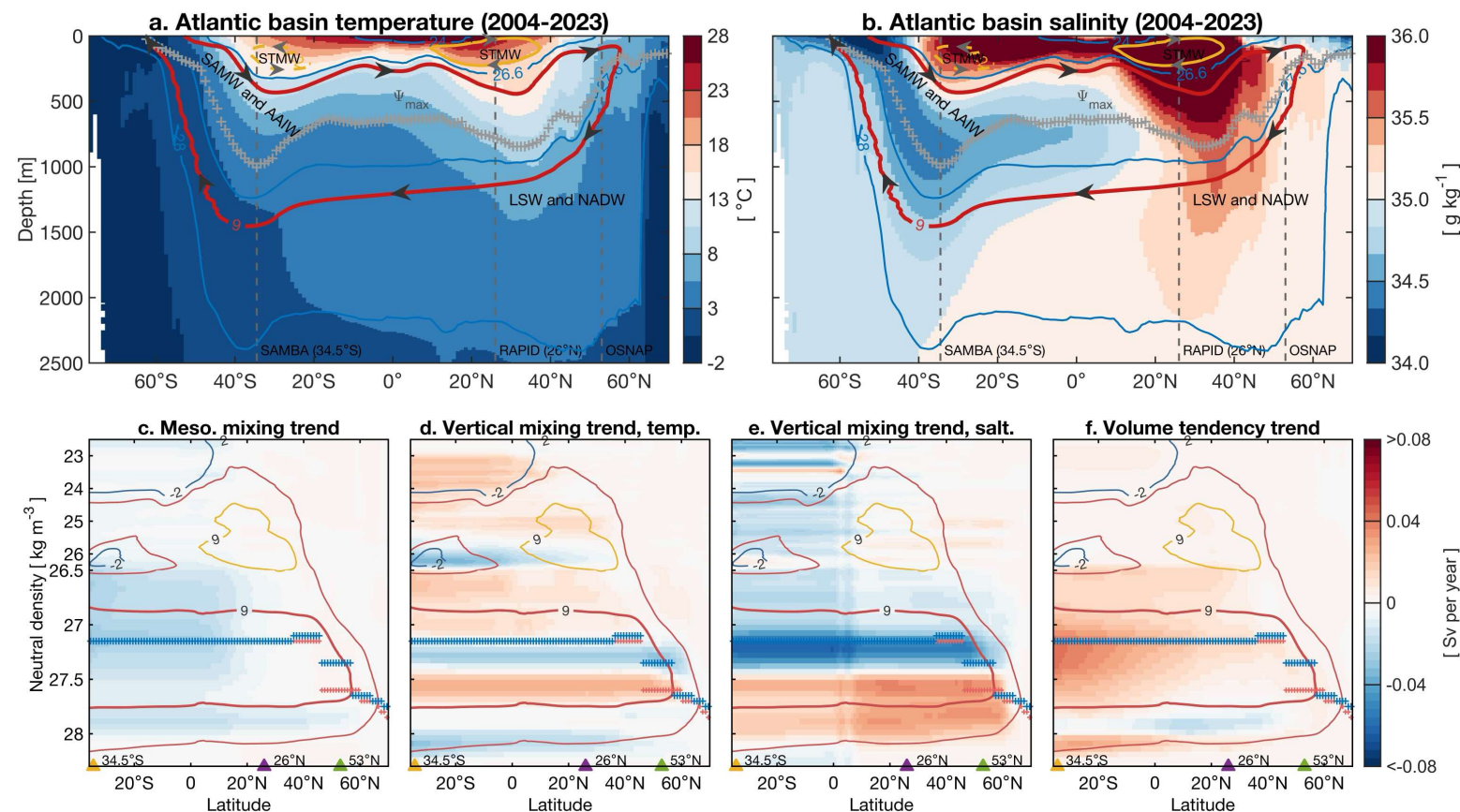
Extended Data Figure 6 | Time evolution of the reconstructed AMOC. Same as Fig. 3b, but showing the ensemble-averaged monthly AMOC strength (Sv) for the period 2004–2023. All time series are smoothed using a 5-year running mean.



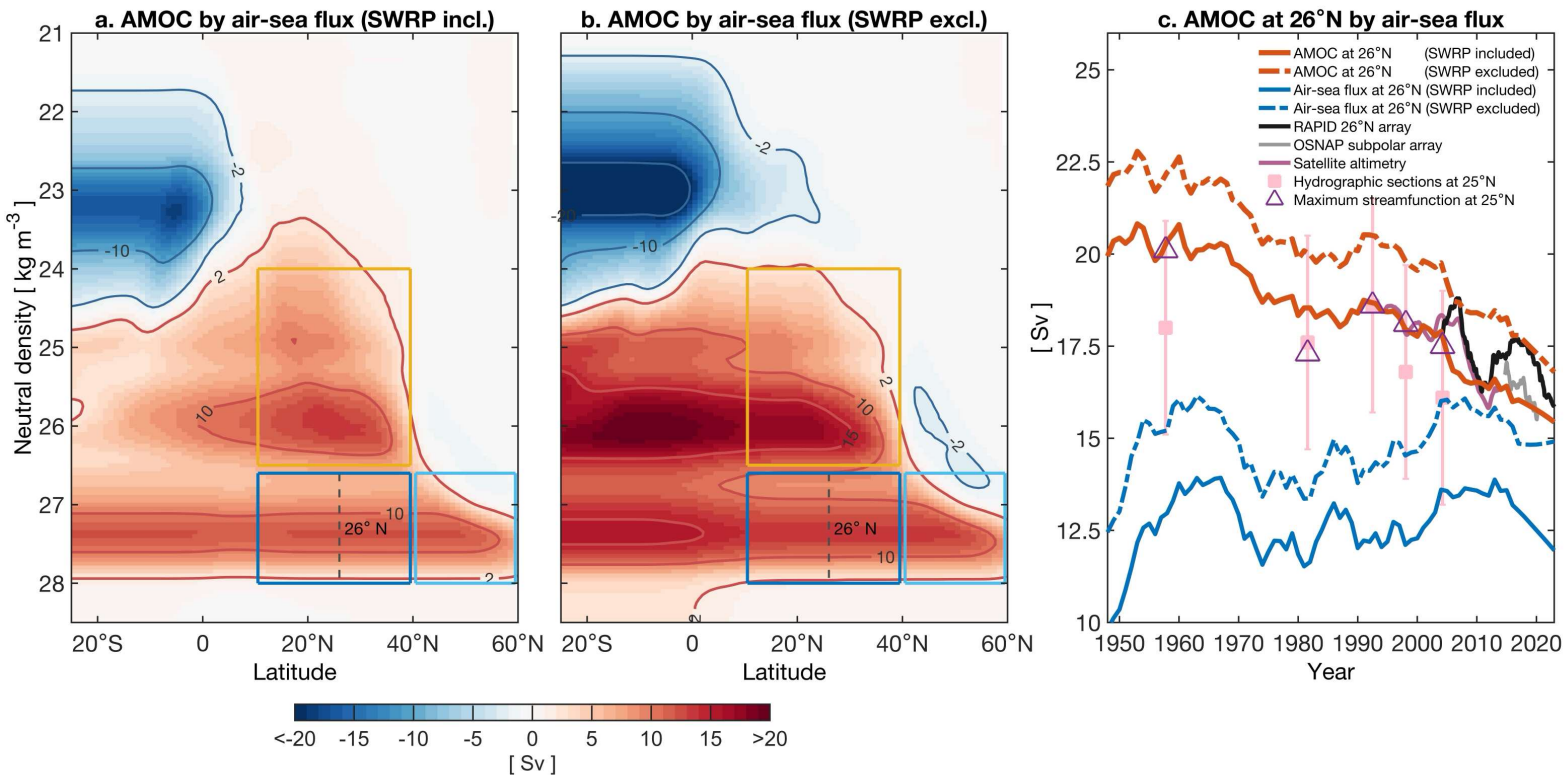
Extended Data Figure 7 | Reconstructed AMOC strength and interhemispheric overturning connectivity. **a**, Same as **Fig. 3b**, but also showing the ensemble-averaged annual mean AMOC strength (Sv) at each latitude (thin lines), defined as the maximum intermediate and deep-layer streamfunction. Time series are shown for the latitude bands 50°S–0° (thin grey lines), 0°–26°N (thin blue lines), and 26°N–60°N (thin red lines), as well as for 34.5°S (green line), based on IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets. All reconstructed lines represent 15-year running averages, with the last 7 years smoothed using a locally weighted linear regression (‘lowess’) approach. Observational estimates of AMOC strength are also shown as in **Fig. 3b**; the yellow line indicates the SAMOC strength observed by the SAMBA array at 34.5°S (3-year running mean). **b**, Lagged correlation between the reconstructed AMOC strength at each latitude and the AMOC strength at 34.5°S. All time series are smoothed using a 15-year running window before computing correlations. The blue contour represents the 0.7 correlation coefficient, and the yellow line indicates the northward propagation of changes observed at 34.5°S into the North Atlantic. Grey crosses indicate the grids where correlations are not statistically significant ($p > 0.05$).



958 **Extended Data Figure 8 | Decomposition of reconstructed AMOC at 34.5°S, 26°N and in**
 959 **the subpolar North Atlantic.** Same as Fig. 4c, but showing the decomposition of
 960 reconstructed AMOC strength (Sv) at (a) 34.5°S, (b) 26°N, and (c) in the subpolar North
 961 Atlantic (40°-65°N) into contributions from air-sea fluxes (dark blue), mesoscale mixing (light
 962 blue), vertical mixing (green), and upper-limb volume tendency (yellow). The vertical mixing
 963 component is further separated into contributions from temperature (thin dashed green) and
 964 salinity (thick dashed green) gradients. The dashed grey line indicates the combined
 965 contribution from all subsurface processes (mesoscale mixing, vertical mixing, and upper-limb
 966 volume tendency). Observational estimates include hydrographic section transports (pink
 967 squares; ref.²⁰), hydrographic streamfunction maxima (purple triangles; ref.²²), satellite
 968 altimetry (purple line; ref.²⁴), and continuous observations from SAMBA (black line in a),
 969 RAPID (black line in b), and OSNAP (back line in c), as in Fig. 3b.
 970



Extended Data Figure 9 | Changes in AMOC streamfunction components. **a-b**, Annual mean zonally averaged **(a)** temperature (°C) and **(b)** salinity (g kg⁻¹) in the Atlantic Ocean during 2004–2023 from IAPv4.2. Blue contours represent neutral density surfaces $\gamma^n = 24.0, 26.6, 27.6,$ and 28.0 kg m^{-3} . Yellow cells highlight the subtropical cells; the thick red contour (9 Sv of reconstructed streamfunction) illustrates the northward pathway of intermediate waters and the southward return flow of LSW and NADW. **c–f**, Decomposition of the linear trend in reconstructed AMOC streamfunction from 1960 to 2023 into contributions from **(c)** mesoscale neutral mixing and horizontal mixing, **(d)** the temperature component of vertical mixing, **(e)** the salinity component of vertical mixing, and **(f)** volume tendency. Crosses mark the layer of maximum streamfunction for 1980–2000 (red) and 2004–2023 (blue).



Extended Data Figure 10 | Overturning streamfunction driven by air-sea buoyancy fluxes.

a, Ensemble-averaged annual mean AMOC streamfunction (Sv) driven by air-sea buoyancy fluxes, including the effects of shortwave radiation (SWR) penetration, as shown in **Extended Data Fig. 3e**. **b**, Same as (**a**), but assuming no SWR penetration into the ocean interior. **c**, As in **Fig. 3b**, but showing the AMOC strength at 26°N (red lines) and the corresponding component driven by air-sea fluxes (blue lines). Solid lines represent estimates that include SWR penetration, while dashed lines show estimates without SWR penetration. All ensemble means are based on IAPv4.2, EN4-ESM, Ishii, and SIO RG-Argo datasets.