

Upper ocean warming in the western subtropical Atlantic drives AMOC weakening at 26.5°N

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Abstract

Observations at about 26.5°N over the past two decades reveal a significant weakening of the Atlantic Meridional Overturning Circulation (AMOC), a central component of Earth's climate system. Here we show that this decline is dominated by density changes in the upper ocean, which account for most of the observed weakening (-0.7 ± 0.4 Sv/decade), while contributions from the deep ocean are comparatively small and not statistically significant. Density anomalies at the western boundary alone would imply a much stronger decline (-1.1 ± 0.3 Sv/decade), but this signal is partially offset by opposing density changes at the eastern boundary at both upper and deep ocean. Basin-wide observations further reveal a pronounced lightening of the upper ocean west of 50°W, primarily driven by warming and associated with a tripole-like pattern in sea surface height. Our findings emphasize the importance of upper-ocean heat convergence and boundary-specific processes in shaping interannual-to-decadal AMOC variability. Together, these results identify upper-ocean western boundary temperature changes as one of the main drivers of the observed AMOC weakening, while highlighting the modulating role of the eastern boundary.

Introduction

The Atlantic Meridional Overturning Circulation (AMOC) is a component of the global overturning circulation and a key regulator of heat, freshwater, and nutrient transport within the Atlantic Ocean^{1,2}. Observations over the past two decades indicate a weakening of the AMOC at about 26.5°N, although the extent to which this decline reflects anthropogenic climate forcing remains uncertain^{3,4}. At the same time, climate projections and proxy-based reconstructions consistently show a substantial AMOC decline by mid- to late-century under anthropogenic forcing with implications for global and regional climate systems^{5–9}. Variations in the AMOC strength play a critical role in influencing European weather^{6,10}, Arctic sea ice variability¹¹, and regional sea levels across the North Atlantic^{12–14}. A weakened AMOC is expected to reduce the North Atlantic CO₂ uptake, which could lead to greater atmospheric CO₂ accumulation and further climate change intensification through a positive feedback loop^{15–18}.

The AMOC at 26.5°N has been measured since 2004 by the joint RAPID-MOCHA-WBTS (Rapid Climate Change - Meridional Overturning Circulation and Heatflux Array - Western Boundary Time Series) program (hereafter RAPID array; Fig. 1). The RAPID array combines estimates of the wind-driven Ekman transport, the Florida Current transport, and the mid-ocean geostrophic transport derived from tall moorings to estimate the AMOC's strength and variability^{19,20}. The RAPID observations have revealed that the AMOC weakened between 2004 and 2012, followed by a strengthening period up to 2018, and weakening through 2024^{4,20–22}. Overall, the AMOC has exhibited a weakening trend since observations have started in 2004²³. However, studies have not found a weakening trend in the Ekman transport^{4,24}, and the Florida Current has remained remarkably stable over the past four decades²³. The AMOC decline is thus mainly attributable to changes in the mid-ocean region, where the southward Lower North

Atlantic Deep Water transport has decreased and the upper ocean southward (upper 1000 m) return flow has simultaneously increased^{4,24}. The question of what has caused the observed volume transport changes in the mid-ocean region remains unresolved²⁴.

Recent studies have identified substantial changes in the upper 1000 m of the North Atlantic, particularly in the western subtropical and tropical regions, that may have important implications for the AMOC^{13,14,26–29}. Steinberg et al.²⁶ show that the North Atlantic subtropical gyre has undergone significant warming and freshening, leading to increased sea levels, along the U.S. East Coast. Huang et al.²⁸ identify a similar pattern of rapid subsurface warming in the subtropical western North Atlantic since 2011, with warming rates exceeding 0.2°C per decade at 200–300 m depth. They find that this warming is primarily associated with the salinity maximum waters in the upper thermocline of the gyre and that their downstream impacts are spread via the Florida Current. Warming of the upper ocean near the western boundary can cause an increase in the zonal gradient of dynamic height (DH) across the mid-ocean region and strengthen the southward interior geostrophic flow in the upper ocean, potentially causing a weakening of the AMOC. The AMOC-relevant implications of these changes are further supported by Gradone et al.²⁷, who document sustained upper ocean warming (~ 0.2°C per decade), surface freshening, and subsurface salinification in the Caribbean Through-Flow since 1960, resulting in stronger stratification and altered heat and freshwater transports. Finally, Volkov et al.¹³ demonstrate that the interannual variability of ocean heat content (OHC) in the subtropical North Atlantic is primarily driven by AMOC-related heat convergence between 26°N and 41°N, and that this convergence leads to large-scale upper ocean warming and sea level rise. These processes contribute to the North Atlantic sea surface height (SSH) tripole, in which SSH anomalies in the subtropical gyre vary inversely with those in the subpolar gyre and the tropics^{14,29} and have increased the frequency of coastal flooding in the southeastern U.S. by up to 50% since 2015¹³. Collectively, these studies point to a robust upper ocean signal of warming and reduced ocean density across the subtropical North Atlantic, with an amplification in the western sector, that may modulate AMOC variability and potentially contribute to its long-term decline.

Changes in density at deeper levels in the water column can also impact the AMOC strength by modifying the southward transport of North Atlantic Deep Water (NADW) within the Deep Western Boundary Current (DWBC)^{30–33}. The DWBC, which transports both upper and lower NADW (UNADW and LNADW, respectively), is the fastest southward pathway for deep water masses to move from the subpolar North Atlantic to the subtropical North Atlantic^{34,35}. Model studies have shown that variability in buoyancy forcing over the subpolar North Atlantic exerts a strong control on the AMOC, particularly on decadal timescales³⁶, and that it can influence lower latitude overturning through the southward propagations of density anomalies along the DWBC^{31,32}. Reduced convection in the Labrador Sea since the early 2000s has been proposed as a driver of AMOC weakening at 26.5°N, through the southward propagation of density anomalies that reach the subtropics within 7–8 years^{30,33}. Consistent with this view, observations reveal declining densities at the western boundary at depths of 2000–3500 m,

associated with a reduced southward transport of NADW and a weakening of the AMOC during 2004–2014³⁷. More recent sustained hydrographic measurements document continued freshening of deep water masses below 2000 m at 26.5°N, linking ongoing property changes in the subtropical North Atlantic to multi-decadal variability originating in the subpolar region and propagating southward along the DWBC³⁸. Together, these findings identify the subpolar North Atlantic as a key source region shaping the DWBC variability, with potentially important implications for the downstream evolution of the AMOC.

Understanding and predicting future changes in the AMOC requires identifying the processes that control its variability. A key unresolved question is whether the recent weakening of the subtropical AMOC is driven primarily by upper ocean density changes or by deep density anomalies propagating from the subpolar North Atlantic along the western boundary. Using observations from the RAPID array at 26.5°N, we show that changes in upper ocean density are the dominant contributor to the observed AMOC decline, with smaller contributions from the deep ocean. We further demonstrate that density changes at the western boundary act to weaken the AMOC, while opposing changes at the eastern boundary partially offset this trend.

Results

a. The impact of Upper and Deep Ocean layers on AMOC weakening at 26.5°N

The AMOC time series at 26.5°N reveals a statistically significant long-term weakening of -0.9 ± 0.6 Sv/decade over the 2004–2024 period (Fig. 2a). All trend estimates presented here are accompanied by their respective 95% confidence intervals, calculated by accounting for the effective number of degrees of freedom in each time series, which is based on the decorrelation timescale of the anomalies. To assess the contribution of density changes at different depths to the AMOC variability, we recompute the RAPID AMOC time series by selectively replacing temperature and salinity profiles over specific depth ranges of the water column by their monthly climatological values while the full variability over the rest of the water column is retained. The resulting geostrophic transport profiles are referenced to 4820 dbar and adjusted by applying a mass compensation term to ensure zero net volume transport across the section (see Methods).

Density changes in the upper ocean (defined here as above 940 m, see Methods) at the eastern and western boundaries collectively accounts for -0.7 ± 0.4 Sv/decade of the overall AMOC trend. The Western Boundary Wedge²⁰ (Fig. 1) changes in transport are formally included in the upper ocean contribution, but its overall trend is negligible (Figure S1). In contrast, density changes in the deep ocean (below 940 m) cause a weaker and statistically insignificant trend contribution of -0.2 ± 0.3 Sv/decade. The Ekman transport and the Florida Current present non-significant long-term influence on the AMOC weakening, with trends of 0.1 ± 0.4 Sv and 0.0 ± 0.3 Sv/decade, respectively (not shown).

When decomposing the upper and deep ocean contributions into boundary components, a clear asymmetry between the western and eastern boundaries emerges. The western boundary DH anomalies caused by both upper and deep layer density changes exert an equally strong and significant weakening influence on the AMOC, (-1.1 ± 0.3 Sv/decade each). This strong western boundary signal is the primary driver of the overall AMOC weakening trend. However, the contribution from the eastern boundary acts to partially oppose this decline: DH anomalies caused by the deep ocean eastern boundary density anomalies contribute to a positive AMOC trend of $+0.9 \pm 0.2$ Sv/decade, while anomalies from the upper ocean exerts a smaller compensating trend of $+0.4 \pm 0.2$ Sv/decade. These results demonstrate that zonal pressure gradients, particularly the differential trends between the western and eastern boundaries in the upper part of the ocean interior, are central to understanding the AMOC's recent weakening.

b. Density trends at the western and eastern boundaries

We now investigate in more detail the long-term density changes throughout the water column at the western and eastern boundaries at 26.5°N (Fig. 3). Boundary density profiles were constructed from merged temperature and salinity records from moored instruments spanning 2004–2024 (see Fig. 1b-c). Both boundaries exhibit general tendencies toward lighter densities (i.e., negative trends), but with markedly different magnitudes and vertical structures. The largest east-west density trend differences occur in the upper 800 m and in the deep ocean below ~ 3000 m, identifying these layers as key contributors to the observed overturning changes (Fig. 3).

At the western boundary, the most pronounced lightening occurs in the upper 800 m, where density trends are both strong and statistically significant (Fig. 3a). In the upper 200 m, the average density trend reaches -0.16 ± 0.05 $\text{kg m}^{-3}/\text{decade}$, decreasing to -0.06 ± 0.01 $\text{kg m}^{-3}/\text{decade}$ between 200 and 400 m, while the layer between 400 and 800 m shows a slightly weaker but still significant trend of -0.019 ± 0.009 $\text{kg m}^{-3}/\text{decade}$. In contrast, the deep ocean at the western boundary exhibits much smaller density changes (Fig. 3b), with density trends generally an order of magnitude weaker than the upper ocean values (average of -0.004 ± 0.0008 $\text{kg m}^{-3}/\text{decade}$).

The eastern boundary shows a distinct vertical structure of density trends compared to the western boundary. The strongest density decrease is confined to the upper 200 m, with an average trend of -0.07 ± 0.02 $\text{kg m}^{-3}/\text{decade}$ (Fig. 3a). Between 200 and 800 m, density trends are weak and heterogeneous, with some layers small, statistically insignificant increases and others modest but significant decreases (ranging between 0.006 ± 0.008 and -0.011 ± 0.003 $\text{kg m}^{-3}/\text{decade}$). In the deep ocean, density trends remain negative but are weaker in magnitude compared to the upper layers (Fig. 3b). The signal becomes damped below 3500 m and approaches near-uniform values of about -0.0007 ± 0.0002 $\text{kg m}^{-3}/\text{decade}$. At both boundaries, density changes are driven primarily by temperature rather than salinity variations (Figs. 3 and S2). In the upper ocean, the density decrease at both boundaries is consistent with the observed warming trends, although localized cooling at the eastern boundary between 200 and 450 m reduces density trends to statistically insignificant levels. Salinity changes generally act as a

secondary, compensating influence (Figs. 3 and S2). Below 940 meters, density trends along the western boundary continue to reflect warming, whereas eastern boundary temperature remains the dominant control on density down to about 3000 meters, with salinity becoming increasingly important at greater depths.

The contrasting vertical structure of density trends at both boundaries leads to distinct changes in the zonal density gradient, with direct implications for the geostrophic components of the AMOC. Two depth ranges emerge as particularly influential. First, in the upper 800 m, the strong lightening near the western boundary drives most of the east-west difference, with only a minor contribution from the upper 200 m of the eastern boundary (Fig. 3a). Secondly, below 3000 m, density trends at the western boundary exceed those at the eastern boundary, again shifting the zonal gradient and potentially contributing to changes in the deep ocean circulation (Fig. 3b). From 800 to 3000 m, the density trends at both boundaries are nearly the same, compensating each other and resulting in minimal net change in the zonal pressure gradient across the basin at these depths. These spatial patterns provide important context for understanding how boundary-driven signals in specific depth ranges influence the long-term AMOC trend at 26.5°N.

c. Impact of density on total volume transport trend

To evaluate how boundary density anomalies shape the AMOC structure in more detail, we examined trends in the DH difference between the eastern and western boundaries and the associated transport per unit depth trend profile across the mid-ocean region, after applying the mass compensation term (See Methods; Fig. 4). The east-west DH shows a significant negative trend through most of the water column. Although density trend differences between the boundaries are close to zero between 800 and 3000 m (Fig. 3), DH integrates specific volume anomalies over depth and therefore remains sensitive to density changes below this layer. As a result, a relatively uniform negative trend in the DH difference is observed between 800 and 3000 m (Fig. 4a; approximately -0.05 ± 0.02 cm/decade), despite the weak local density trend contrasts.

The trend in transport per unit depth shows a distinct vertical pattern: a negative trend in the upper 800 m and a positive trend below 3000 m (Fig. 4b), suggesting a reorganization of the overturning structure in response to changing boundary conditions. The maximum negative transport trend occurs in the upper 200 m (-0.003 ± 0.001 Sv/decade), coinciding with the region of greater changes in the DH gradient. The DH difference trend reaches a maximum of -0.35 ± 0.10 cm/decade, with approximately 85% of this signal arising from the upper 800 m (~ 30 cm/decade) and 15% from depths below 3000 m (~ 5 cm/decade). These vertical patterns in DH and transport per unit depth trends indicate that the observed AMOC weakening is dominated by warming-induced density changes at the western boundary, with a significant contribution from density anomalies at the eastern boundary.

In summary, the -0.9 Sv/decade weakening of the AMOC (Fig. 2) is reflected in a reduced upper overturning and is balanced by the increased transport in the deep layers, primarily below 3000 m

(Fig. 4b). The depths at which transport per unit depth trends are strongest (Fig. 4b; the upper 800 m and below 3000 m) closely correspond to the layers where zonal density gradients (Fig. 3) and DH differences (Fig. 4a) exhibit their largest trends. This provides further evidence that boundary density changes govern the structure of the observed AMOC weakening.

d. Spatial structure of Upper Ocean density trends in the North Atlantic

We next examine how upper ocean density trends at 26.5°N relate to basin-scale patterns of temperature, salinity and density across the North Atlantic. Basin-wide density trends derived from Argo observations³⁹ reveal spatial structures in the upper 0–200 m and 200–940 m layers that are consistent with the RAPID boundary measurements at 26.5°N (Figs. 3a and 5; see Methods). Both layers exhibit a pronounced lightening along the western boundary, whereas at the eastern boundary, a significant density decrease is confined to the upper 200 m, with weak or insignificant trends below. The upper 200 m density trends form a tripole-like meridional structure, with alternating trend polarity north of 45°N, between 25 and 45°N, and south of 25°N, closely resembling the SSH tripole pattern reported previously^{13,14}.

In the 200–940 m layer, statistically significant negative density trends are mostly confined to the western subtropical North Atlantic (Fig. 5c). Along the eastern boundary, trends in this layer are weak or insignificant, consistent with the small and vertically heterogeneous density changes observed in the mooring records at 26.5°N (Fig. 3a). The time series of spatially averaged density anomalies near each boundary further highlight this contrast (Figs. 5b and 5d). At the western boundary, the upper 200 m layer shows a pronounced density decrease beginning around 2011, with large negative anomalies persisting through 2016. This intense lightening west of 50°W is driven by strong warming anomalies in the upper 200 m (Figure S2). Despite the magnitude of the 2014–2016 anomaly at the western boundary, the long-term density trend remains robust and is not dominated by this episodic event. In contrast, the eastern boundary exhibits a more gradual and sustained density decrease limited to the upper 200 m, with no trend in the 200–940 m layer.

Temperature trends dominate the upper ocean density signal across the North Atlantic, while salinity trends present a compensating influence (Figures S3 and S4). In both the upper 200 m and the 200–940 m layers, spatial patterns of density change closely follow those of temperature, indicating that warming-driven buoyancy changes are the dominant driver of the observed density trends. Consistent with this, the spatial distribution of the upper ocean density trends closely mirrors previously observed SSH tripole patterns and corresponding OHC anomalies^{13,14}. Specifically, significant warming is evident near the western boundary in both the upper 200 m and 200–940 m layers, while near the eastern boundary, warming is confined to the upper 200 m, with no significant trend between 200 and 940 m (Figures S2, S3 and S4). Together, these results indicate that the density trends estimated from mooring records at 26.5°N reflect the large-scale basin structure of upper ocean temperature change.

Discussion and Conclusions

a. Upper ocean control of the AMOC decline

The observed weakening of the AMOC at 26.5°N over the past two decades is primarily driven by upper ocean subtropical Atlantic warming, rather than by deep-water density anomalies propagating from the subpolar North Atlantic^{30,33,36}. Although deep western boundary density changes are detectable, and consistent with previous analyses highlighting the importance of RAPID array western boundary moorings in resolving long-term overturning variability⁴⁰, their net contribution to the AMOC trend is largely offset by deep eastern boundary adjustments. The strong asymmetry between western and eastern boundary density trends highlights the importance of zonal pressure gradients in controlling overturning strength on decadal timescales.

Density changes over 2004–2024 exhibit a clear vertical and zonal structure that governs the overturning response. The dominant signal arises from pronounced lightening at the western boundary in the upper ocean, whereas eastern boundary trends are shallower and weaker. Below ~ 3000 m, eastern boundary trends are also smaller than the lightening trend observed at the same levels at the western boundary. In contrast, density trends between roughly 800 and 3000 m are nearly equal at both boundaries. Although the mechanisms underlying the changes at the eastern boundary are not addressed here, further investigation of the causes of those changes would be valuable. The overturning response therefore reflects vertically localized boundary asymmetries rather than a uniform basin-wide density decline.

Zonal DH gradients and transport-per-unit-depth trends agree with the observed boundary density trend differences. The transport trends feature a strengthened southward interior flow in the upper ocean and a compensating adjustment at depth. However, only the upper-ocean density changes contribute significantly to the overall long-term AMOC trend. The observed ~ 0.9 Sv/decade upper ocean transport weakening therefore reflects a reorganization of the interior geostrophic transport rather than changes in the Florida Current or Ekman transport, which have remained stable^{23,24}. In the deep ocean, a reduction in southward NADW transport^{4,41} is consistent with mass conservation, whereby enhanced southward flow in the upper ocean requires a compensating northward transport anomaly in the deep ocean, leading to a reduced southward NADW transport. The trends in deep boundary densities partly contribute to this declining NADW transport, while the remainder is due to barotropic compensation that is required to maintain a net mass balance across the entire section.

Different generations of climate models show broadly consistent AMOC weakening over the first two decades on the 21st century at 26.5°N, partially overlapping the RAPID array observational period. The multi-model means of the Coupled Model Intercomparison Project Phases 5 and 6 (CMIP5 and CMIP6) simulate AMOC declines of approximately – 0.7 and – 0.6 Sv/decade, respectively, over 2000–2020⁴³. These values fall within the uncertainty range of the RAPID estimate over 2004–2024, indicating consistency in sign and comparable magnitude between climate models and observations during the recent period.

However, it is worth emphasizing that the AMOC trend observed thus far at 26.5°N – related to changes in the ocean transport within the ocean interior – is in contrast with outcomes from the Intergovernmental Panel on Climate Change (IPCC) models. IPCC projections suggest that the 21st-century AMOC decline is associated with a weakening of the Florida Current, or more broadly, the surface western boundary current system⁴².

Many models simulate a gradual AMOC decline under anthropogenic forcing, typically unfolding over multi-decadal to centennial timescales^{5–7}. Because the RAPID array record spans only 2 decades, it is not long enough to isolate multi-decadal to centennial trends associated with anthropogenic forcing. The apparent differences in mechanisms of AMOC weakening may therefore reflect differences in the timescales considered rather than a fundamental inconsistency between observations and model projections. Our results suggest that distinct processes dominate AMOC variability at different time scales, with upper ocean subtropical density adjustments governing the observed trend thus far at 26.5°N.

b. Consistency with upper ocean heat content patterns

Upper-ocean density trends across the subtropical gyre are dynamically consistent with the RAPID boundary observations at 26.5°N (Fig. 5). Pronounced lightening west of 50°W in the upper 200 m aligns with the North Atlantic SSH tripole pattern and reflects basin-scale heat convergence within the subtropical gyre. The SSH tripole is consistent with the observed density trends at both eastern and western boundaries, with especially strong lightening near the western boundary during the 2010–2016 period, consistent with anomalous OHC accumulation in the western subtropical gyre during the tripole's positive phase.

In contrast, upper ocean density changes along the eastern boundary are weaker and confined to the upper 200 m. This boundary asymmetry further confirms the western boundary's dominant role in altering the pressure gradients that drive the AMOC weakening. The upper ocean signals are thus both temporally and spatially consistent with interannual-to-decadal variability in OHC and sea level observed across the North Atlantic^{13,14}. By increasing DH near the western boundary, the recent AMOC weakening also contributes to elevated coastal sea level and enhanced flood risk along the southeastern United States¹³. Together, these patterns indicate a dynamically coupled system linking AMOC trends, regional heat content changes, and boundary density anomalies.

In summary, our results provide observational evidence that recent changes in the overturning circulation at 26.5°N are predominantly governed by upper-ocean processes on the western side of the basin, where density anomalies drive DH changes that translate into basin-scale adjustments of overturning transport. These findings reinforce the mechanistic links between the 20-year AMOC trends and basin-wide climate variability, including the North Atlantic SSH tripole, while also highlighting local impacts such as regional dynamic sea level rise and increased coastal flood risk.

Methods

a. Temperature and Salinity data

Temperature and salinity measurements at about 26.5°N were obtained from the moored instruments of the RAPID array near the western and eastern boundaries of the Atlantic basin (Fig. 1). Data from multiple moorings at each boundary were combined to form representative western and eastern boundary profiles²⁰. The configuration of the eastern boundary moorings has changed over time to minimize operational costs while maintaining data quality and accuracy⁴⁰. Both eastern and western boundary T and S data were vertically interpolated to a 20 dbar grid using previously established methodology^{20,44,45}. The 12-hour resolution T and S profiles were averaged to monthly values. The density (ρ) was estimated using the Gibbs Sea Water package (<https://www.teos-10.org/software.htm>). The monthly climatological signal was obtained by fitting a two-harmonic model to the detrended timeseries using least squares. The anomaly signal was obtained by removing the monthly climatological signal from the original timeseries.

b. AMOC estimates at 26.5°N

The AMOC observing system is a collaborative effort between the U.K. RAPID Climate Change program, the U.S. Meridional Overturning Circulation and Heat-Flux Array (MOCHA), and NOAA's Western Boundary Time Series (WBTS) program. The RAPID array estimates semi-daily AMOC transports since 2004, using three key components: the Florida Current transport, Ekman transport, and mid-ocean transport.

The Florida Current transport is measured using a subsea electromagnetic cable spanning the Florida Straits, calibrated with direct velocity observations during regular hydrographic cruises²³. The Ekman transport is calculated from surface wind stress data provided by ERA5 reanalysis⁴⁶. Mid-ocean transport between the Bahamas and Western Africa, is measured through DH moorings positioned at both the eastern and western boundaries (Fig. 1). The resulting geostrophic transport profiles are referenced to 4820 dbar and adjusted by applying a mass compensation term to ensure zero net volume transport across the section²⁰. Mid-Atlantic ridge tall moorings were recently discontinued⁴⁰, as experiments have shown that most of the AMOC signal can be retrieved by using only the moorings located near the eastern and western boundaries⁴⁷. Direct transport measurements in the Antilles Current and the DWBC region immediately offshore of the Bahamas, referred to as the Western Boundary Wedge, complement the poleward mid-ocean transport estimates²⁰.

The AMOC transport is defined as the maximum streamfunction value, typically found at around 1100 m depth, separating the upper limb (comprising the Florida Current, Ekman transport, and upper mid-ocean transport) from the lower limb (comprising the Upper and Lower North Atlantic Deep Water).

c. Trend estimates

The trend estimates are derived by linear regressions applied to time series. The 95% confidence intervals for the trend account for the effective number of degrees of freedom, which is adjusted based on the decorrelation timescale of the timeseries. This approach ensures that the uncertainty in the trend estimate properly reflects the autocorrelation structure of the data. For temperature, salinity and density data, the upper 100m data were excluded from trend analysis, as the shallowest moored instrument can occasionally be outside this range due to mooring tilt caused by local conditions.

d. Influence of different regions on the AMOC weakening

To separate the upper and deep ocean layers, we defined the $\sigma_2 = 36.5 \text{ kg m}^{-3}$ isopycnal (density referenced to 2000 dbar) as the upper boundary of the NADW, consistent with widely used definitions of deep water mass structure in the North Atlantic Ocean^{48,49}. The mean depth of this isopycnal, calculated from the time-averaged mooring data at both the western and eastern boundaries, is approximately 940 m, which was adopted as the fixed interface between the upper ocean (0–940 m) and the deep ocean (940–4820 m). This depth differs slightly from the level used operationally by the RAPID array, but it more accurately reflects the observed isopycnal structure at the boundaries in our dataset.

The influence of a given area anomalies on the AMOC transport is assessed by computing the difference between the RAPID AMOC estimate²⁰, and the AMOC estimated from eight experiments:

1. Ekman transport contribution to AMOC trend: monthly values of the Ekman transport profile are replaced by their corresponding monthly climatological values.
2. Florida Current contribution to AMOC trend: monthly values of the Florida Current transport profile are replaced by their corresponding monthly climatological values.
3. Western boundary contribution to AMOC trend: Western boundary wedge transport per unit depth profiles were set to their monthly climatological mean and western boundary DH profiles were calculated from monthly climatological profiles of temperature and salinity at the western boundary.
4. Upper ocean western boundary contribution to AMOC trend: Western boundary wedge transport per unit depth profiles were set to their monthly climatological mean and western boundary DH profiles were calculated from merged upper ocean (upper 940 m) monthly climatological profiles of temperature and salinity and original temperature and salinity deep ocean (between 940 and 4820 m) data at the western boundary.
5. Deep ocean western boundary contribution to AMOC trend: Western boundary DH profiles were calculated from merged upper ocean (upper 940 m) original temperature and salinity data, and deep ocean (between 940 and 4820 m) monthly climatological profiles of temperature and salinity at the western boundary.
6. Eastern boundary contribution to AMOC trend: Eastern boundary DH profiles were calculated from monthly climatological profiles of temperature and salinity at the eastern boundary.

7. Upper ocean eastern boundary contribution to AMOC trend: Eastern boundary DH profiles were calculated from merged upper ocean (upper 940 m) monthly climatological profiles of temperature and salinity and original temperature and salinity deep ocean (between 940 and 4820 m) data at the eastern boundary.
8. Deep ocean eastern boundary contribution to AMOC trend: Western boundary DH profiles were calculated from merged upper ocean (upper 940 m) original temperature and salinity data, and deep ocean (between 940 and 4820 m) monthly climatological profiles of temperature and salinity at the eastern boundary.

Although the western boundary wedge extends to depths of 4000 m, it is excluded from the deep western boundary experiment because it does not capture the full transport of the DWBC and is dominated by variability in the upper ocean. Moreover, no significant trends in transport per unit depth are observed below 1000 m (Figure S1).

e. Gridded upper ocean data product

To examine large-scale temperature and density trends in the upper ocean, we used the Roemmich-Gilson Argo Climatology³⁹, obtained from https://sio-argo.ucsd.edu/RG_Climatology.html. This gridded product provides monthly temperature and salinity fields from 2004 to 2023 at a $1/4^\circ$ horizontal resolution, based on optimal interpolation of Argo profiling float data. The vertical resolution varies from 10 m in the upper 200 m to 100 m at maximum depth of 2000 m. Temperature and salinity fields were used to compute density, and linear trends were estimated in two depth layers: the upper 200 m and the 200–940 m layer. The upper 200 meters layer was analyzed separately, as this layer exhibits significant density variations along both the eastern and western boundaries (see Fig. 3). At depths between 200 and 940 m, density trends at the eastern boundary are close to zero. These data were used to map temperature and density trends across the North Atlantic ($0-70^\circ\text{N}$) and to derive regional averages representative of the eastern and western boundary regions at 26.5°N .

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Figures

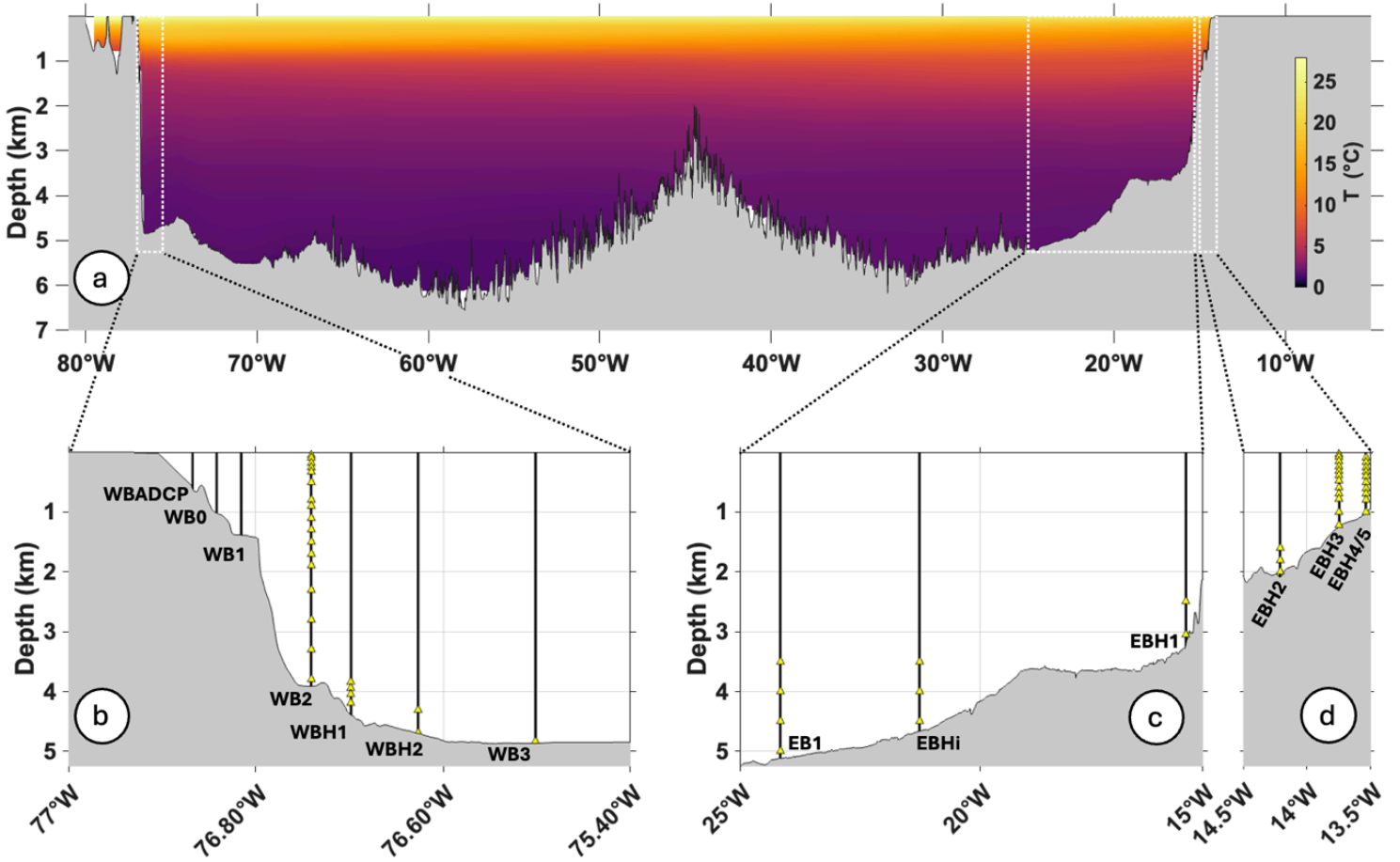


Figure 1

Mooring configuration of the RAPID array at 26.5°N. (a) Time-mean temperature section (2004–2024) from the RAPID-2D product, with bathymetry from the GEBCO dataset²⁵. White dashed boxes indicate zoomed-in regions at the western and eastern boundaries. (b) Western boundary mooring array, including the western boundary wedge moorings (WBADCP, WB0 and WB1). (c) Deep eastern boundary moorings array. (d) Shallow eastern boundary moorings near the continental shelf. Yellow triangles mark the locations of instruments measuring temperature and salinity across all panels used in this work.

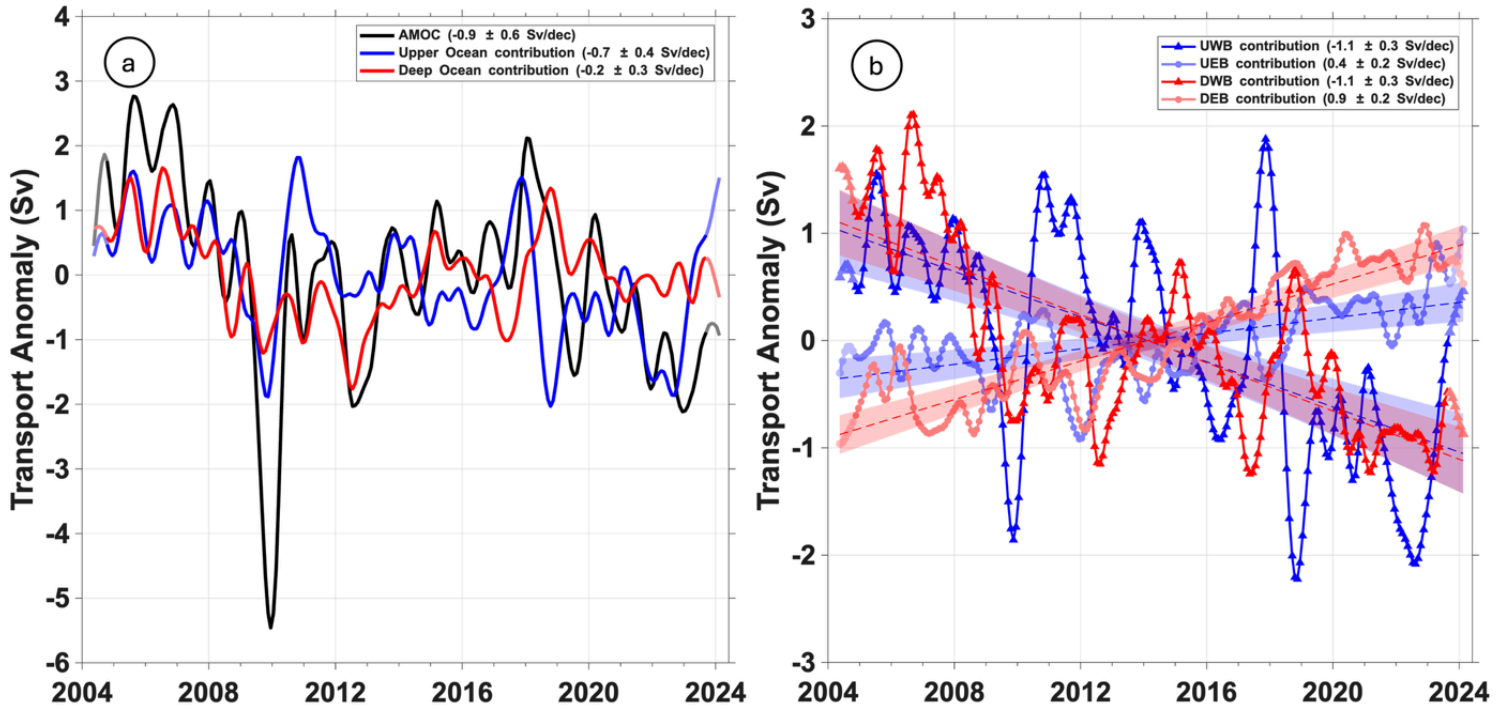


Figure 2

The 2004-2024 AMOC time series and the contribution of different layers to the AMOC trend. (a) Time series of the AMOC transport anomaly at 26.5°N (black curve), along with its upper ocean (0-940 m; blue curve) and deep ocean (940-4820 m; red curve) contributions. (b) Upper and deep ocean contributions further separated into western (UWB and DWB, respectively; triangles) and eastern (UEB and DEB, respectively; circles) boundary components, with blue and red tones indicating upper and deep layers, respectively. Dashed lines denote linear trends for each time series, and shaded areas represent 95% confidence intervals. Time series are smoothed using a 13-month low-pass Gaussian filter for visualization. Linear trends and associated 95% confidence intervals (in Sv/decade) are calculated from the original monthly anomalies (unfiltered) and are displayed next to each legend entry. The first and last 6 months of each series are shown with lighter shading to indicate boundary effects of the filter.

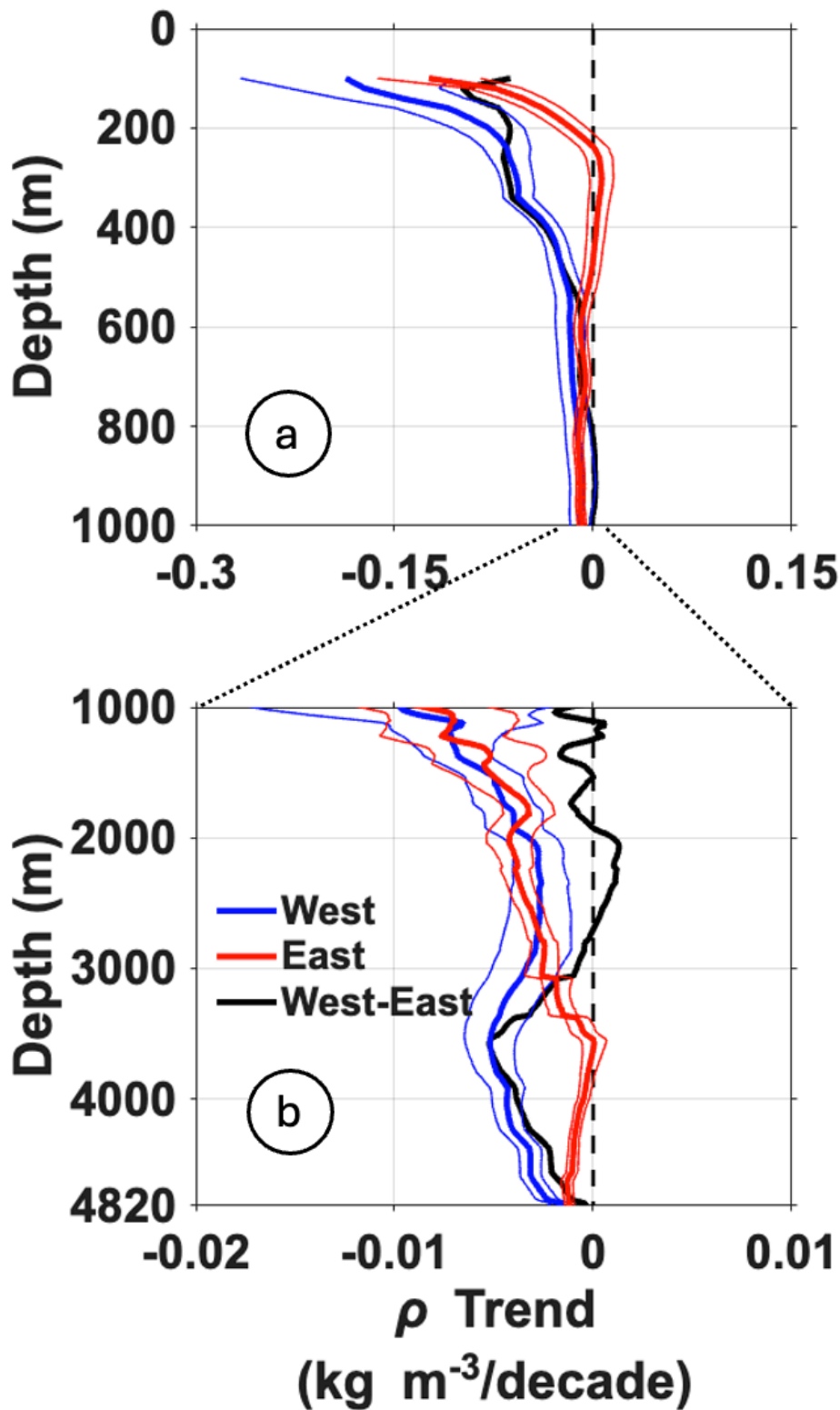


Figure 3

Vertical profiles of 2004-2024 linear density trends at the 26.5°N section. (a) Density trend profiles for the upper 1000 m at the western (thick blue curve) and eastern (thick red curve) boundaries. Thin curves of the same color denote the 95% confidence intervals. (b) Same as panel (a), but for the 1000–4820 m depth range, showing deep ocean density trend profiles for both boundaries. Vertical profiles of the difference in density trends between western and eastern boundaries (i.e., western minus eastern) are

shown in black curves. Dashed vertical lines indicate zero trend for reference. Note the difference in x-axis scale between upper and lower panels, reflecting the larger trend magnitudes in the upper ocean (dotted lines mark the x-axis limits of the lower panel for direct visual comparison).

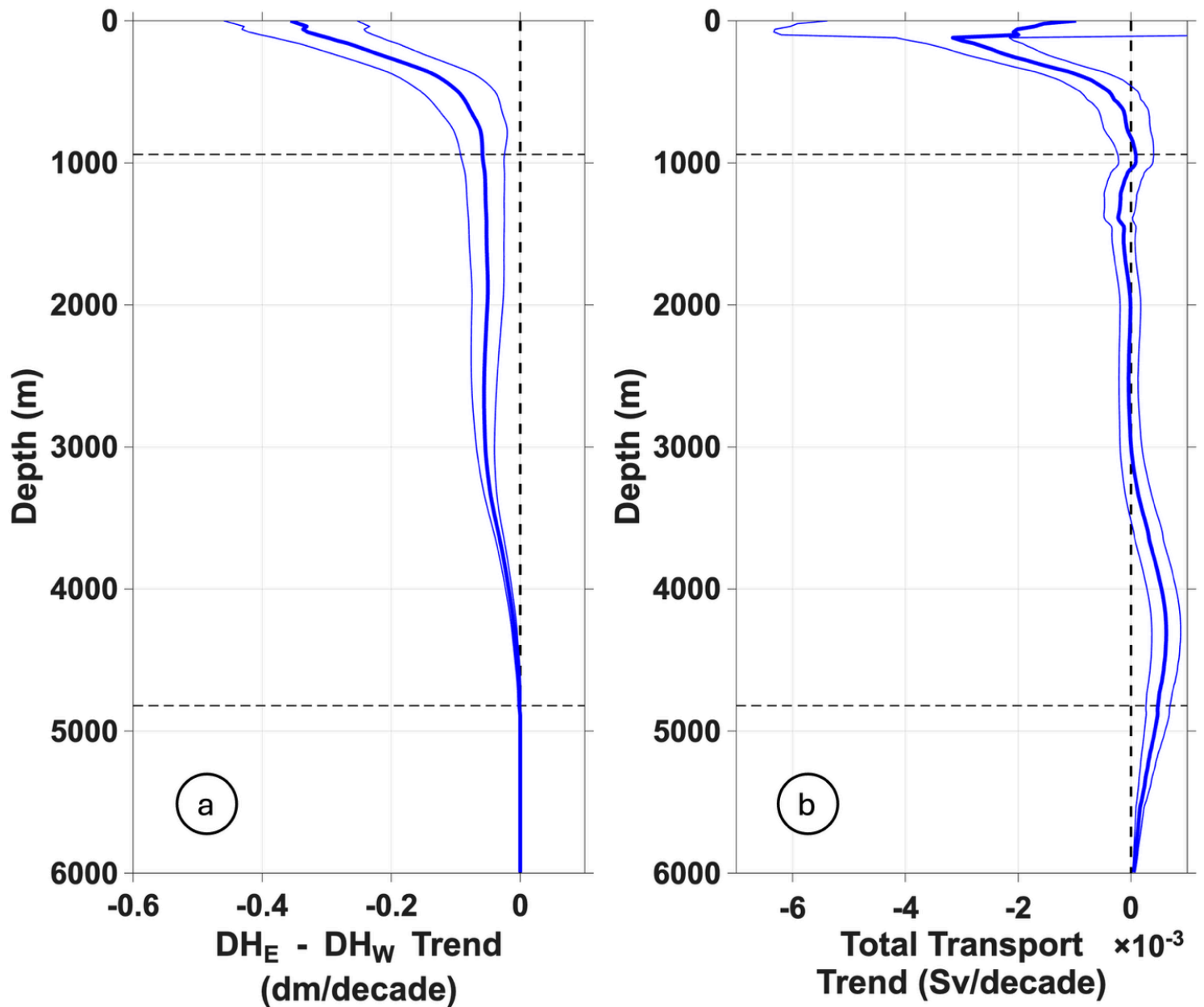


Figure 4

Vertical profiles of the long-term trends in zonal DH difference and corresponding total transport per unit depth at 26.5°N. (a) Trend in DH difference (eastern minus western boundary) across the water column (thick blue curve). (b) Trend in total transport per unit depth across the full section at 26.5°N (thick blue curve). In both panels, thin blue lines represent the 95% confidence intervals. The vertical dashed line indicates zero trend, while horizontal dashed lines mark the 940 m and 4820 m levels, which define the boundaries of the upper ocean (upper 940 m) and deep ocean (between 940 and 4820 m), respectively.

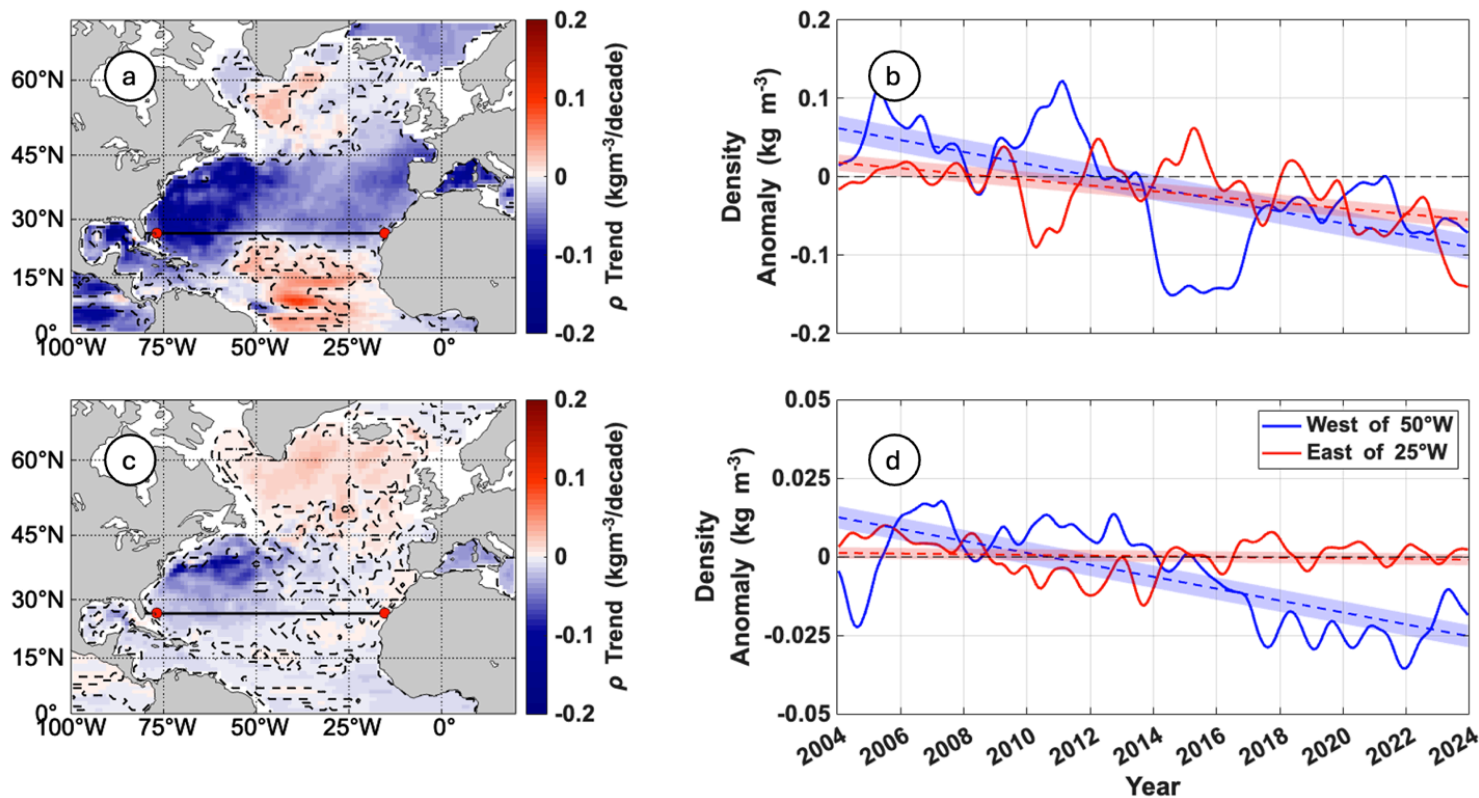


Figure 5

Upper ocean density trends for the North Atlantic (0–70°N) from 2004 to 2023 based on Argo data (see Methods). (a) Spatial distribution of density trends in the upper 200 m, with red tones indicating positive trends (densification) and blue tones indicating negative trends (lightening). The solid black line marks the latitude of the RAPID array at 26.5°N, and red dots indicate the approximate positions of the eastern and western boundary moorings (see Figure 1 for details). (b) Time series of 13-month low-pass filtered density anomalies averaged over the upper 200 m west of 50°W (blue curve) and east of 25°W (red curve) along 26.5°N. (c) Same as panel (a), but for the 200–940 m layer. (d) Same as panel (b), but for the 200–940 m layer. Dashed lines denote linear trends for each time series, and shaded areas represent 95% confidence intervals. Time series are smoothed using a 13-month low-pass Gaussian filter for visualization. Note that panels (b) and (d) have different y-axes.

Supplementary Files

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