



Observed substantial South Atlantic Upper-Ocean Water Mass changes over the past 45 years

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Abstract. Global climate change is driving widespread transformations in the ocean. The South Atlantic, a key component of the global meridional overturning circulation that regulates the climate, is thought to be undergoing large-scale and deep-reaching compound state changes in temperature and salinity throughout the basin, but the underlying physical processes remain unclear. In this study, leveraging advanced ocean observing networks, we investigate the observational evidence and the physical processes shaping both individual and compound (temperature–salinity coupled) changes from a water-mass perspective based on rigorous analyses of observational uncertainty. We find that there are substantial properties changes in all upper-ocean water masses in the South Atlantic over the past 45 years, in which the physical contributions of these changes are due to distinct dynamic and thermohaline processes across different water masses: Surface Water has warmed and become saltier since the 1980s, accompanied by a ~10% increase in volume driven. The South Atlantic Central Water has expanded vertically by ~1.15 times driven by vertical displacement. The Antarctic Intermediate Water has warmed by roughly 0.25 °C, leading to a density decrease dominated by thermal effects. The pronounced intermediate water warming signal represents a potential early fingerprint of anthropogenic change, now affecting ~70% of the basin. It has been mixed and advected near-isopycnally into the South Atlantic from the Subpolar Front, Southern Ocean and the Indian Ocean. This study advances basin-scale observational evidences and the physical mechanisms underlying regional individual and compound changes, providing new insight into regional climate variability and long-term changes in a warming climate.

1 Introduction

The ocean has absorbed more than 90% of the excess heat associated with Earth’s energy imbalance, thereby moderating the rate of global warming (Cheng et al. 2022; Minière et al. 2023; von Schuckmann et al. 2023). Concurrently, shifts in the global hydrological cycle are intensifying salinity contrasts and amplifying regional wet–dry differences under continued warming (Cheng et al. 2020; Lu et al. 2024). Ocean circulation is also evolving—from local to basin and global scales—redistributing heat and freshwater across basins and among water masses (Garcia-Soto et al. 2021; Pita et al. 2024; Talley 2011). In parallel, long-term compound state changes, including concurrent warming, salinification, acidification, and deoxygenation, as well as

compound extremes such as marine heatwaves accompanied by extreme deoxygenation events, are becoming evident from the surface to the deep ocean (Gruber et al. 2021; Tan et al. 2026; Wong et al. 2024). Together, these shifts illustrate the profound and interconnected physical changes underway in the ocean under ongoing climate change.

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The South Atlantic (SA) exhibits individual and compound changes similar to those observed across the global ocean and is a focal region for their expression (Fig. 1a; see also Tan et al. (2026)). The SA plays a central role in global heat and freshwater distribution (Evans et al. 2017; McDonagh et al. 2015; Talley 2008) and is tightly coupled to the Atlantic meridional overturning circulation (AMOC) (Dong et al. 2015; Manta et al. 2021; Zheng et al. 2024). Over the past six decades, sustained
 40 warming and salinization have been documented throughout the basin, with an ocean heat content (OHC) increase of $\sim 0.5\text{--}1.25 \text{ W m}^{-2}$ from 1958–2019 and long-term salinity trends of $\sim 0.01\text{--}0.06 \text{ g kg}^{-1}$ per century from 1960–2017 (Cheng et al. 2020; Cheng et al. 2022). Li et al. (2023b) further report accelerating warming in the mid-latitude South Atlantic, largely accumulated within mode and intermediate waters, with ensemble-mean OHC anomalies of $\sim 0.7\text{--}1.2 \times 10^9 \text{ J m}^{-2}$ during 2010–2020 relative to the 1980–2000 mean. As one motivation for this study, several regional hotspots of compound long-term
 45 change—such as the South Atlantic Current (SAC), the eastern tropical gyre, the Brazil Current (BC), and the Subantarctic Front (SAF) had been identified by the direct observations, from the epipelagic zone (0–200m) to the mesopelagic zone (200–1000 m; see Fig. 1a). The combined epipelagic and mesopelagic zone (0–1000m) is of particular importance because it is the habitat of diverse pelagic communities and contains important biogeochemical and ecological processes in the South Atlantic, in which the compound changes of temperature coupled with salinity may have substantial impacts on pelagic organisms (e.g.,
 50 an example of SA zooplankton (Lopes et al. 2018)). However, despite evidence of basin-scale individual and compound trends (Tan et al. 2026), major gaps remain in our understanding of the SA: (i) the physical processes driving these compound changes are not yet fully understood, and (ii) the responses of the various South Atlantic water masses remain debated—for example, the magnitude and variability of Antarctic Intermediate Water (AAIW) temperature and salinity trends (McCarthy et al. 2011; McCarthy et al. 2012; Rusciano et al. 2012; Yao et al. 2017). Addressing these gaps forms the basis of this study.

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The concept of a “water mass”, first introduced by Helland-Hansen (Helland-Hansen 1916), has been foundational to physical oceanography for more than a century. Water masses are typically identified by their temperature–salinity (T–S) characteristics (an example is shown in Fig. D1), which reflect the conditions in their formation regions. Their identification and classification rely on a combination of qualitative and quantitative approaches. Traditional qualitative methods use T–S diagrams together
 60 with phenomenological knowledge to categorize waters into surface, subsurface, intermediate, deep, and bottom classes (Talley 2011). Quantitative techniques include threshold-based definitions based on potential or neutral density (e.g., North Atlantic Deep Water (Talley 2013): $\gamma^n = 27.8\text{--}28.04 \text{ kg m}^{-3}$), as well as multivariate clustering (e.g., K-means, fuzzy C-means) and machine-learning approaches (Chen et al. 2022; Sato and Polito 2014). Recent advances further incorporate dynamical indicators such as potential vorticity and mixed-layer depth (MLD) to better resolve water-mass evolution (Li et al. 2023b).
 65 When water masses subduct, they retain the thermohaline “fingerprints” of the climate conditions prevailing during their formation and transformation (Fox-Kemper et al. 2021), thereby playing a central role in ocean heat uptake and the distribution



and structure of freshwater. The IPCC Sixth Assessment Report (AR6) underscores the importance of this framework: subtropical mode waters and Southern Ocean intermediate waters are assessed with high confidence and represent some of the most robust subsurface indicators of climate change (Fox-Kemper et al. 2021). In addition, the water mass is also modulated and transported by the (sub-)mesoscale eddy activities (Dong et al. 2025).

Analyses of water-mass property changes depend strongly on the coordinate framework used. Two common approaches are (i) fixed-depth coordinates and (ii) density (isopycnal) coordinates. Changes evaluated at fixed depth capture the full signal, combining true water-mass property changes due to air–sea flux perturbations with changes due to vertical displacements of water masses due to circulation variability. This perspective links naturally to Earth’s energy and freshwater budgets, which are typically evaluated within fixed-depth layers (e.g., 0–2000 m depth average (Cheng et al. 2022)). However, it can misrepresent water-mass distributions and introduce spurious diabatic circulations (Nurser and Lee 2004). In contrast, isopycnal analyses isolate intrinsic property changes by largely removing vertical displacement (i.e., “heave”) signals, thereby offering a clearer view of long-term water-mass modification (McIntosh and McDougall 1996; Nurser and Lee 2004). The ocean community commonly distinguishes between “heave”—changes in the depth of an isopycnal due to adiabatic redistribution—and “spice”—changes in properties such as temperature or salinity along an isopycnal due to diabatic transformation—a framework first introduced by Bindoff and McDougall (1994). Removing the heave component, which often dominated by natural variability such as wind-driven oscillations, helps reveal long-term property trends by reducing spurious mixing and diabatic-circulation artifacts (Bindoff and McDougall 1994; Portela et al. 2020). Although well-defined this decomposition is not perfect—for example, uniform warming generates both heave and spice components (Bindoff and McDougall 1994)—the isopycnal framework (especially the neutral density framework) is widely used in studies of long-term water-mass change (Iudicone et al. 2008; Manta et al. 2021; Rusciano et al. 2012; Tailleux 2021) because it can minimize such artifacts and emphasizes underlying property transformations.

Multiple studies have classified South Atlantic water masses using a variety of criteria (Chen et al. 2022; Li et al. 2023b; Manta et al. 2021), with three upper-ocean water masses consistently identified: Surface Water (SW), South Atlantic Central Water (SACW), and Antarctic Intermediate Water (AAIW). SW occupies the uppermost layer and is continuously modified by intense air–sea fluxes and ventilation (Manta et al. 2021), extending from the surface to ~150 dbar (Fig. 1c–d, Fig. D1). SACW forms the main thermocline water mass, and is characterized by the strongest vertical temperature and salinity gradients. It extends from ~150 dbar to ~750 dbar (Fig. 1c–d, Fig. D1) (Talley 1996; Talley 2011). AAIW is readily identified at intermediate depths by its pronounced low salinity and relatively high oxygen content, occupying depths from ~600 to 1200 dbar (Fig. 1c, Fig. D1), where the salinity minimum (S_{min}) is used to define the “AAIW core” (Rimaud et al. 2012; Rusciano et al. 2012; Talley 1996). Three types of AAIW are identified, distinguished by their source regions and properties: Atlantic/Pacific-origin AAIW (A-AAIW), Indian-origin AAIW (I-AAIW), and Indo-Atlantic AAIW (IA-AAIW) (Rusciano et al. 2012). These water masses are distinguished by robust modal structures in hydrographic properties, even though the exact placement of water-mass boundaries varies across studies (because of differences in diagnostic criteria and thresholds).



Despite major advances in documenting global-scale temperature and salinity changes (IPCC 2021), the South Atlantic has historically received less attention than the North Atlantic. Prior to the Argo era (pre-2005), hydrographic data primarily came from repeated transects conducted during WOCE and later GO-SHIP programs, most notably near 24°S and 30°S (Caínzos et al. 2022; Gouretski and Koltermann 2004; Katsumata et al. 2022; King et al. 2001) (see also Fig. 1b and Fig. D2-D3). Additional observational coverage was along boundaries, coastal regions, and continental margins, with some sampling across the central subtropical gyre and mid-ocean (20°–40°S; Fig. D2-D3). Most of these additional samples are OSD (Ocean Station Data), which includes Bottle, XCTD, and low resolution CTD, including a large contribution of observations from the Soviet Union (Zhang et al. 2024) (Fig. D2a-c). In addition, repeated XBT (Expendable Bathythermograph) trans-basin transects (e.g., AX97, AX18, AX08, AX15) by commercial ships supported by NOAA/AOML, Brazilian- and French- oceanic institutions also contributed to the advance of the regional observing system before the 2000s (Goes et al. 2026; Goni et al. 2019). Although the spatial and temporal coverage of observations is not as extensive as that available after the 2000s (Fig. D3), particularly at the mesoscale and on daily to seasonal timescales, these advances have made it possible to investigate decadal to longer-term changes before the 2000s.

Substantial improvement in observation coverage occurred after the global implementation of the Argo program in the mid-2000s, which progressively expanded profiling coverage throughout the basin (Argo 2000; Roemmich et al. 2019) (Fig. D2-D3). After 2009, sustained full-depth moored time series were established, including the SAMBA array at 34.5°S (Meinen et al. 2018). The South Atlantic Meridional Overturning Circulation (SAMOC) observing system, initiated in 2015, further enhanced long-term monitoring of circulation variability (Chidichimo et al. 2023; Kersalé et al. 2020; Pita et al. 2024). With the development of new capabilities within the global ocean observing system (Cheng 2023), recent advances in data recovery, quality control, and spatial gap-filling methodologies have substantially improved the availability of subsurface temperature (Cheng et al. 2024; Zhang et al. 2024) and salinity (Cheng et al. 2020; Li et al. 2023a; Tan et al. 2025), with continuous records extending back to the 1960s. These improvements, together with high-quality hydrographic data collected along key repeated transects (e.g., 24°S, 30°S), now enable the analysis of high-quality monthly gridded products in combination with targeted analyses of historical repeat ship-based hydrography. This integrated approach allows for rigorous assessments of product robustness in the South Atlantic and supports investigations of water-mass changes across the entire basin, including with reliable information extending back to the 1980s.

In this study, we focus on the observational evidence and the physical drivers of the long-term physical changes of upper-ocean water masses—temperature, salinity, density, and thickness—over the past 45 years, given their central role in shaping compound ocean state change and ocean circulation. Specifically, we address three scientific questions: (i) how have key physical properties of South Atlantic upper-ocean water masses changed in response to the large-scale and deep-reaching compound ocean state change over the past 45 years; (ii) which regions exhibit the strongest compound changes in physical

properties along isopycnal surfaces; and (iii) what physical processes underpin these observed individual and temperature-salinity coupled changes in the context of ongoing global warming?

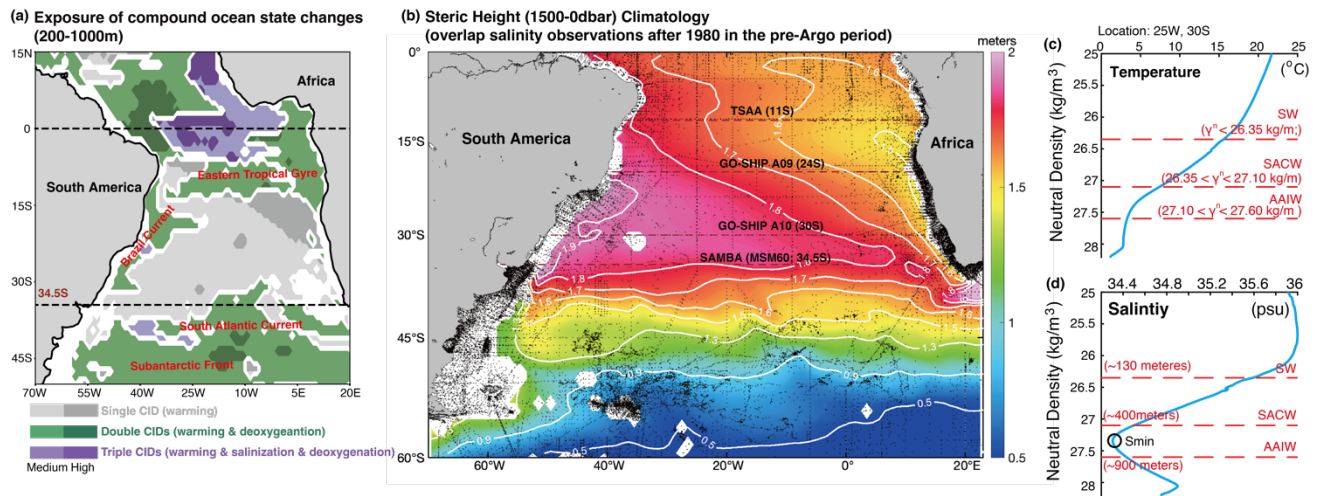


Fig. 1: Observational context and water-mass framework for the South Atlantic. (a) Spatial distribution of compound ocean state changes—concurrent variations in temperature, salinity, and dissolved oxygen—between 200 and 1000 m in the South Atlantic (used the data from Tan et al. (2026)). Here, the names of the primary surface currents are indicated. (b) Pan-South Atlantic map showing steric height climatology (1500–0 dbar; 2005–2024, white isolines with color patch), with superimposing black dots that indicate historical salinity observations collected during the pre-Argo period (1980–2005). Brown dashed lines show the locations of major hydrographic sections: Tropical South Atlantic Array (TSAA; at 11°S), GO-SHIP A09 (24°S), GO-SHIP A10 (30°S), and SAMBA/MSM60 (34.5°S). (c, d) Typical temperature and salinity climatology (2005-2024 mean) profiles (at 25°W, 30°S) plotted against neutral density. Horizontal red dashed lines denote the neutral-density bounds used to define the water masses examined in this study: Surface Water (SW), SACW, AAIW. The AAIW salinity minimum (Smin) and the basin mean depth of neutral density boundary are highlighted in (d). This study focuses on the core South Atlantic region (45°S–0°N).

2 Data and Methods

2.1 Datasets

In this study, we examine changes in upper-ocean water-mass properties, temperature, salinity, density, and thickness, over the past 45 years. Density and thickness are derived from temperature and salinity across pressure levels using observation-based gridded products, augmented by *in-situ* temperature–salinity profiles from cruise observations and Argo floats in the South Atlantic.

2.1.1 Observational-based gridded products



(1) IAP temperature and salinity gridded products

160 One of the primary datasets used in this study is the monthly IAP (Institute of Atmospheric Physics, Chinese Academy of Sciences) global temperature gridded product (Cheng et al. 2024) (Version 4.1) and the global salinity gridded product (Cheng et al. 2020) (Version 2.1), provided at $1^\circ \times 1^\circ$ horizontal resolution for the upper 6000 m across 119 standard depth levels from 1980 to 2024. The methodological strengths of the IAP products have been well documented: they include an enhanced automatic quality-control system that accounts for local temperature and salinity distribution skewness, topographic barriers, and climate-driven shifts in water-mass properties (Tan et al. 2023; Tan et al. 2025); bias corrections for in situ temperature measurements from XBTs, MBTs, and bottle data (Cheng et al. 2014; Gouretski and Cheng 2020; Zhang et al. 2024) and a gap-filling ensemble optimal interpolation (EnOI) approach that applies dynamic covariances to minimize sampling errors arising from heterogeneous temporal and spatial observation density (Cheng and Zhu 2016). A major advantage of the IAP product is its comprehensive uncertainty quantification (e.g., an example of temperature-OHC from Bao et al. (2026)). The EnOI system uses a dynamic ensemble to generate 30 temperature and 30 salinity realizations, and the spread of these ensemble members represents both instrumental uncertainty and reconstruction (sampling and mapping) uncertainty (Cheng and Zhu 2016).

(2) SIO Argo gridded product

To complement the IAP datasets and to assess the robustness of our water-mass property estimates, we also use an independent regional-scale observational product: the SIO/Scripps Argo monthly temperature and salinity fields (Roemmich–Gilson Argo Climatology; https://sio-argo.ucsd.edu/RG_Climatology.html) (Roemmich and Gilson 2009). This product is available at $1^\circ \times 1^\circ$ resolution from 2004 to 2024 and is based exclusively on Argo profiling floats. It provides global temperature and salinity fields from the surface to 2000 m depth using 58 standard levels. Importantly, the SIO-Argo product was not affected by the salinity drift that impacted certain Argo datasets after 2016 (Barnoud et al. 2021), making it an additional reliable source for evaluating the consistency of gridded water-mass property reconstructions.

2.1.2 In-situ temperature/salinity profiles

(1) MSM60 cruises

Hydrographic temperature and salinity data from the MSM60 cruise (R/V Maria S. Merian) are incorporated into this study (Karstensen et al. 2019). The measurements were collected along the nominal 34.5°S SAMBA line in January 2017 (Fig. 1b) (Meinen et al. 2018; Valla et al. 2018), with stations spaced at approximately 50 km intervals and extending from the surface to the seafloor. Each station recorded temperature, salinity, dissolved oxygen, and pressure using a pumped Sea-Bird SBE 9plus CTD system. The dataset has undergone expert manual quality control, including removal of outliers and calibration adjustments (Sloyan et al. 2019) and consists of 128 high-quality temperature and salinity profiles.

(2) GO-SHIP A09, A10 cruises

190 We further use temperature and salinity observations from two GO-SHIP sections downloaded from Katsumata et al. (2022): A09 at 24°S , occupied in March 2009, and A10 at 30°S , occupied in October 2011. Both transects are shown in Fig. 1b. The



profiles were obtained through the GO-SHIP Easy Ocean data product and subjected to rigorous manual quality control, including CTD calibration and removal of anomalous measurements. These sections represent the highest standard of oceanographic observations. In total, 95 full-depth profiles from A09 and 122 from A10 are used in this study.

195 (3) Individual Argo floats

To further test the robustness of our dataset choices, we also incorporate individual Argo temperature and salinity profiles for the South Atlantic. These profiles were obtained from the CAS–Oceanographic Data Center, Global Ocean Science Database (CODC) (Tan et al. 2025; Zhang et al. 2024), sourced from the Argo Global Data Assembly Center (GDAC) (Argo 2000; Le Traon et al. 2020). The dataset has undergone standard Argo quality-control and bias-correction procedures (Tan et al. 2025; 200 Zhang et al. 2024). A total of 120,518 temperature and salinity profiles collected between 2005 and 2023 within the domain 0–45°S and 65°W–20°E were compiled for analysis. Prior to use, profiles flagged (quality control’s flag) as unreliable by either GDAC (Wong et al. 2020) or CODC (Tan et al. 2023) were removed, as were all floats listed on the Argo grey list (Wong et al. 2020).

2.1.3 Heat Flux and Evaporation-Precipitation datasets

205 Monthly net surface heat flux data from 1990 onward are obtained from the ERA5 reanalysis (Hersbach et al. 2020), an atmospheric–ocean coupled product produced by ECMWF at 0.25° spatial resolution. ERA5 assimilates a wide range of multi-platform observations, including satellite radiance, radiosonde measurements, and in situ data from ships, buoys, and drifters. The net heat flux, defined as positive upward (ocean to atmosphere) and negative downward (atmosphere to ocean), includes contributions from shortwave radiation, longwave radiation, sensible heat, and latent heat (Hersbach et al. 2020).

210 Evaporation and precipitation fields are taken from the Objectively Analyzed Air–Sea Fluxes (OAFux) dataset (Yu and Weller 2007) and the Global Precipitation Climatology Project (GPCP) dataset (Adler et al. 2018), respectively. OAFux provides monthly evaporation estimates at 1° resolution beginning in 1958, with evaporation parameterized using flux-related variables such as wind speed, sea-surface temperature, and near-surface air humidity and temperature, primarily derived from satellite 215 observations (Yu and Weller 2007). GPCP supplies monthly precipitation fields at 2.5° resolution from 1980 onward, combining satellite retrievals with measurements from global rain-gauge and radiosonde networks (Adler et al. 2018).

2.2 Robustness of the dataset choices in the South Atlantic

In this study, the IAP temperature and salinity gridded products are adopted as the primary data source for assessing water-mass property changes over the past 45 years. A major advantage of the IAP product is the availability of ensemble-based 220 uncertainty estimates extending back to 1960—information not accessible in other global products such as Ishii or SIO-Argo. Although previous independent evaluations have demonstrated the robustness of the IAP framework for global and regional physical variability (Dong et al. 2021; Gulev et al. 2021; Kirchengast and Pichler 2025; Minobe et al. 2025), its performance for water-mass analyses in the South Atlantic remains insufficiently documented. A similar situation applies to the SIO-Argo



product. We therefore evaluate the skill of both datasets by comparing the salinity minimum ($S_{\min}$) of AAIW core along
225 selected transects with independent, high-quality in situ observations from GO-SHIP and MSM60 cruises, which serve as the
benchmark.

Comparison with ship-based CTD data indicates that both IAP and SIO-Argo capture the core thermohaline structure of AAIW
without producing spurious spikes or abrupt shifts (Fig. 2). Because gridded products are constructed from monthly atlases
230 that blend neighboring observations, transect-scale variability appears smoother relative to individual cruise measurements,
reflecting the influence of spatial and temporal averaging (Cheng et al. 2024). At the eastern boundary of the 24°S section
(A09), both products show a positive offset relative to GO-SHIP observations (Fig. 2a). This may reflect unresolved seasonal
variability not fully represented in the monthly gridded fields. Although IAP and SIO-Argo display broadly consistent patterns
and long-term tendencies, minor differences arise from the distinct observational inputs and processing methods used to
235 construct each dataset. IAP integrates XBT, CTD, and Argo observations, whereas SIO-Argo relies exclusively on Argo floats.
Despite these differences, the root-mean-square errors (RMSE) for both products relative to benchmark profiles are nearly
zero, indicating high fidelity in reproducing AAIW core properties. Additional validation against individual Argo profiles after
2005 (Fig. 2e–f) reveals strong positive correlations between IAP estimates and Argo-derived $S_{\min}$ and the temperature of $S_{\min}$,
with most comparisons aligning closely along the one-to-one line. Furthermore, the spatial patterns of $S_{\min}$, the pressure of
240 $S_{\min}$, and AAIW thickness in the western and southern African sectors, as derived from IAP, are consistent with those reported
by Rusciano et al. (2012).

These comparisons (Fig. 2) indicate that both IAP and SIO-Argo reliably capture the thermal and saline characteristics of
AAIW core in the South Atlantic without introducing systematic biases. Although smoothing is inherent in gridded products
245 due to climatological averaging, the key water-mass signals are preserved, supporting their use for basin-scale analyses,
particularly for SW and SACW, where observational density is greater than for AAIW. On this basis, we adopt the IAP and
SIO-Argo gridded products as the primary datasets for the analyses presented in this study.

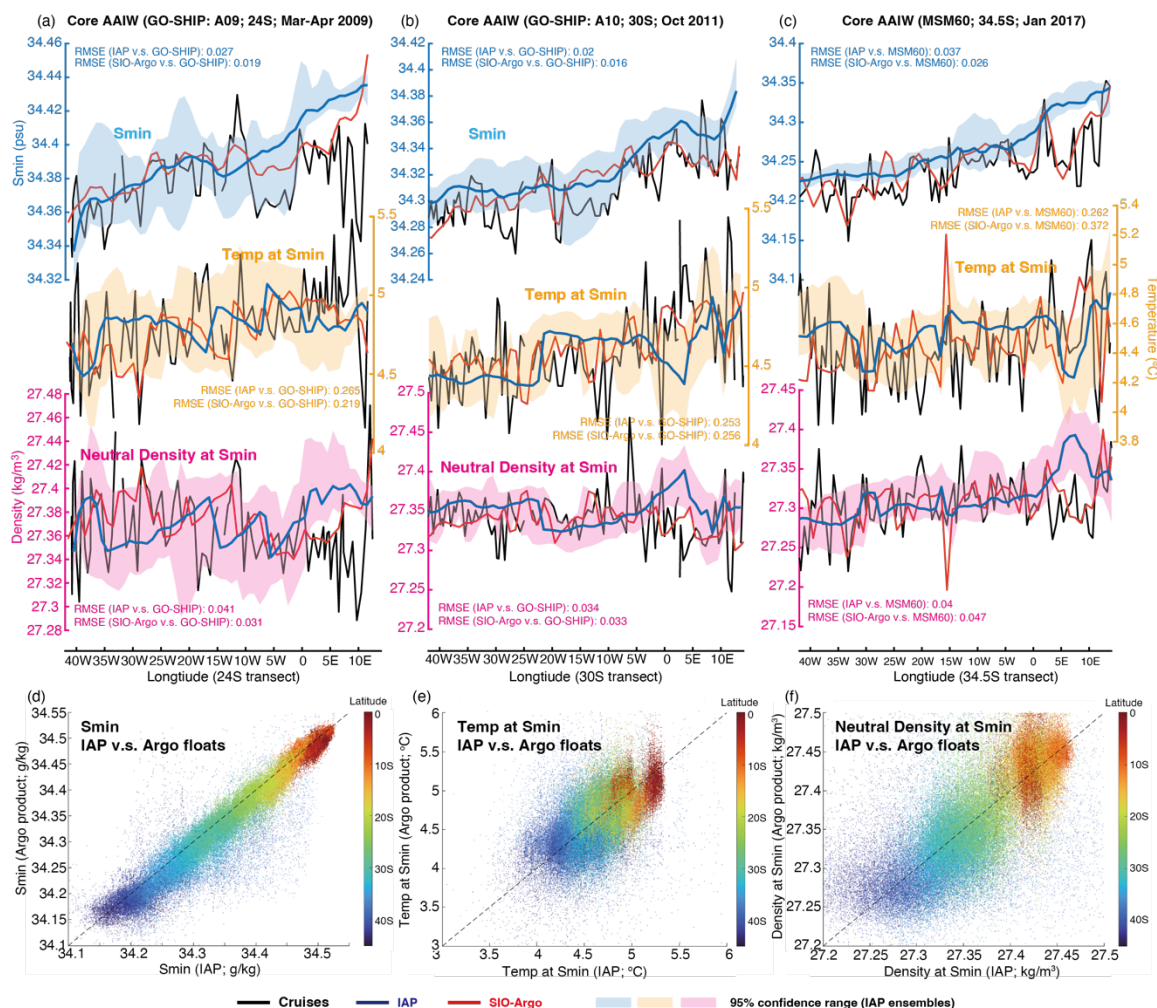


Fig. 2: Robustness of dataset choices for estimating AAIW core properties—salinity minimum (Smin), temperature at Smin, and neutral density at Smin—using atlas products and independent cruise and float observations. Two monthly gridded products are evaluated: the IAP temperature–salinity atlas and the SIO-Argo gridded product (see *Method* section). (a) Along-track comparison with GO-SHIP A09 (March–April 2009, 24°S). (b–c) Same as (a), but for GO-SHIP A10 (October 2011, 30°S) and MSM60 (January 2017, 34.5°S). Shaded envelopes denote the 95% confidence interval of the IAP estimates, incorporating instrumental, sampling, and mapping uncertainties. Root-mean-square errors (RMSEs) for each comparison are reported. (d–f) Scatterplot comparisons between IAP atlas estimates (x-axis) and individual Argo float measurements (y-axis), with point color indicating latitude.

2.3 Water mass framework refined and used in this study

Three upper-ocean water masses (SW, SACW, AAIW) discussed in this study exhibit distinct hydrographic characteristics: SW is characterized by high temperatures (mean conservative temperature ≈ 17.85 °C) and high absolute salinity ($S_a \approx 35.74$ g kg⁻¹) (Fig. D1), and is strongly modified by air–sea fluxes and ventilation (Manta et al. 2021). SACW (mean conservative temperature ≈ 11.63 °C; mean $S_a \approx 35.22$ g kg⁻¹; Fig. D1) is a generalized mode water class that encompasses the upper



permanent thermocline and pycnocline (Manta et al. 2021). It shares the overlap density range of the Subtropical Mode Water (STMW), as well as the northward transformation Subantarctic Mode Water (SAMW) across the Subantarctic Front (Talley 1996; Talley 2011). Here, the South Atlantic STMW, formed through deep winter convection with low potential vorticity, is subducted beneath the seasonal thermocline and progressively modified by lateral advection and mixing, thereby ventilating and contributing to the upper and middle layers of SACW. While, SAMW forms independently in deep winter mixed layers north of the Subantarctic Front, enters and subducts into the South Atlantic, and mixes with the denser lower SACW. In the eastern tropical South Atlantic, part of SACW (e.g., Eastern South Atlantic Central Water) also reflects admixture with waters entering from the Indian Ocean via Agulhas leakage, transported into the basin through rings, filaments, and intrusions (Chen et al. 2022; Laxenaire et al. 2019; Laxenaire et al. 2020; Sato and Polito 2014; Tim et al. 2018).

AAIW is identifiable on the temperature-salinity (T–S) diagrams by its low salinity (Fig. D1), with a fingerprint of salinity minimum ($S_{\min}$; Fig. 1c) that defines the “AAIW core” (Rimaud et al. 2012; Rusciano et al. 2012; Talley 1996). AAIW entering the South Atlantic originates primarily from two source regions: Atlantic AAIW (A-AAIW): Formed north of the Subantarctic Front through the transformation and subduction of SAMW, characterized by low practical salinity ($S_p \leq 34.2$) and associated with deep winter mixed layers and vigorous mode-water formation in the southeast Pacific and southwest Atlantic sectors of the Southern Ocean (Fig. 1b) (Sallée et al. 2010; Sloyan and Rintoul 2001; Suga and Talley 1995; Talley 1996; Talley 2013). A-AAIW spreads northward along both boundary and interior pathways, forming the classical low-salinity AAIW core (Emery and Meincke 1986; Rimaud et al. 2012; Rusciano et al. 2012). Indian-origin AAIW (I-AAIW): Formed from SAMW and AAIW in the southwest Indian Ocean, entering the South Atlantic via Agulhas leakage with higher salinity ($S_p \geq 34.3$) (McCarthy and Talley 1999; Sloyan and Rintoul 2001). This transport occurs primarily within Agulhas rings, filaments, and the Benguela Current system, with limited modification in the southern Cape Basin (Beal et al. 2011; Gordon 1986; Rimaud et al. 2012; Rusciano et al. 2012; Schmid et al. 2003). A third variety, Indo-Atlantic AAIW (IA-AAIW; $34.2 < S_p < 34.3$), has recently been identified as the product of isopycnal mixing between A-AAIW and I-AAIW in the Cape Basin (Rusciano et al. 2012).

In the literature, the boundary between water masses is defined using a range of approaches, and no consensus exists on either the methodologies or the thresholds applied. Some studies classify water masses using neutral density or potential density (Gao et al. 2022; Li et al. 2023b), while others impose density ranges combined with constraints on temperature, salinity, dissolved oxygen, thickness, or potential vorticity (Chen et al. 2022; Lv et al. 2023; Manta et al. 2021). Additional work has argued that fixed thresholds may themselves evolve in a warming climate, further complicating water-mass classification (Zhou et al. 2024). In the South Atlantic, however, the core or “modal” structures of the major water masses are sharply defined and readily distinguishable: AAIW is marked by a pronounced salinity minimum at intermediate depths (Fig. 1c–d; Fig. D1). SACW exhibits the strongest vertical gradients of temperature and salinity in the region (Fig. D1), coincident with the permanent pycnocline and a local maximum in Brunt–Väisälä frequency (N^2) (Feucher et al. 2019). These modal characteristics are robust; therefore, modest shifts in water-mass boundaries—whether induced by climate change or by alternative diagnostic criteria—



do not generally alter the fundamental interpretation of their core properties (Gao et al. 2022). In this study, our primary goal is to characterize the modal/core properties of each water mass. As such, our results are relatively insensitive to the precise placement of water-mass boundaries.

Following Manta et al. (2021) and Li et al. (2023b), we estimate water-mass properties using time-constant neutral density (γ^n) (Jackett and McDougall 1997)). Neutral density defines surfaces along which water parcels can move without experiencing a buoyant restoring force and thus more closely represents real-ocean pathways of interior mixing and advection than potential density (Jackett and McDougall 1997). This framework enables us to capture the true isopycnal evolution of water masses (i.e., “spice”) while isolating the effects of vertical displacements (“heave”). Neutral density is computed from *in-situ* temperature, practical salinity, and pressure following pre-TEOS-10 standards, after which practical salinity and *in-situ* temperature are linearly interpolated onto neutral-density surfaces (Jackett and McDougall 1997). These surfaces are defined at 0.005 kg m⁻³ intervals between $\gamma^n = 22$ and 28.30 kg m⁻³, consistent with the recommendations of Jackett and McDougall (1997).

To characterize basin-scale spatiotemporal trends over the past 45 years, we adopt a generalized water-mass classification based on neutral density (Manta et al. 2021). South Atlantic water masses are categorized as Surface Water (SW; $\gamma^n < 26.35$ kg m⁻³), South Atlantic Central Water (SACW; 26.35–27.10 kg m⁻³), and Antarctic Intermediate Water (AAIW; 27.10–27.60 kg m⁻³), as shown in Fig. 1b–c. To assess the robustness of these definitions, we conducted sensitivity experiments in which each boundary was perturbed by ± 0.03 kg m⁻³. Consistent with Gao et al. (2022), water-mass property trends remain insensitive to such small thresholds shifts (figures omitted). In this study, we therefore focus on SW, SACW, and AAIW. Water masses denser than AAIW, such as Upper and Lower Circumpolar Deep Water, North Atlantic Deep Water, and Antarctic Bottom Water, are not examined due to the rapid decline in observational coverage below ~ 1000 m, which precludes robust long-term assessment.

Water-mass-averaged temperature, salinity, neutral density, thickness, and volume are computed on isopycnal surfaces, following the approach of Foppert et al. (2021), to accurately represent adiabatic evolution along neutral trajectories without introducing artificial mixing or diabatic circulations (Nurser and Lee 2004). Specifically, a property $\bar{D}$ for a given water mass is calculated as:

$$\bar{D} = \frac{\iiint D(x,y,\gamma) dV}{\iiint dV} = \frac{\sum_{x,y} \sum_{\gamma=\gamma_1}^{\gamma_2} D(x,y,\gamma) \Delta H_{x,y,\gamma} A_{x,y}}{\sum_{x,y} \sum_{\gamma=\gamma_1}^{\gamma_2} \Delta H_{x,y,\gamma} A_{x,y}} \quad (1)$$

where $D(x,y,\gamma)$ is the water-mass property at horizontal location (x,y) on the neutral-density surface γ , $A_{x,y}$ is grid-cell horizontal area, and

$$\Delta H_{x,y,\gamma} = z(x,y,\gamma_2) - z(x,y,\gamma_1) \quad (2)$$



330 is the layer thickness between γ_1 and γ_2 . The denominator represents the full three-dimensional volume of the selected water mass.

For AAIW, we additionally examine its “core” properties, defined at the depth of the local salinity minimum ($S_{\min}$) (Manta et al. 2021; McCarthy et al. 2012). The core layer is identified as the minimum salinity between $27.20 < \gamma^n < 27.60 \text{ kg/m}^3$ at 335 depths deeper than 400 m. The variables evaluated at this level include $S_{\min}$, temperature at $S_{\min}$, and neutral density at $S_{\min}$.

To detect long-term changes in water-mass properties over the past 45 years, we apply multichannel singular spectrum analysis (M-SSA) diagnostics. M-SSA decomposes time series into trend, oscillatory variability, and noise, and when applied to multiple variables, it highlights synchronized modes and potential coupling mechanisms (Groth and Ghil 2015). With ~45 years of data, this method can resolve variability with periods up to ~15–20 years (Manta et al. 2024). Here we follow the 340 Monte-Carlo M-SSA protocol of Manta et al. (2024) to isolate trends longer than 20 years across the water-mass properties.

Because our study investigates changes occurring under global ocean warming, we also assess when and where long-term signals emerge from background natural variability. Time-of-emergence (ToE) is computed as the following: the long-term signal is obtained by applying a 25-year LOWESS filter to each time series (1960–2023), while the noise is quantified as 345 variability on timescales shorter than 25 years (Tan et al. 2026). A signal-to-noise ratio ($|SNR|$) greater than 1 is then computed to determine the ToE. Together, M-SSA and ToE provide complementary information on the magnitude, spatial coherence, and detectability of long-term physical changes, allowing us to link the underlying dynamical mechanisms to the observational emergence of potential anthropogenic warming in the South Atlantic.

2.4 Water mass properties uncertainty estimation

350 Uncertainty in observation-based estimates of water-mass properties arises primarily from limitations in data coverage, differences among mapping and interpolation methods (including climatological assumptions), and instrument-related errors such as systematic biases in XBT, MBT, or bottle data and imperfect quality control. To quantify these uncertainties, we used the 30-member IAP temperature ensemble and 30-member salinity ensemble (Cheng and Zhu 2016) to generate 900 ($=30 \times 30$) unique temperature–salinity realizations via the multiplication principle. For each ensemble pair, neutral density and other 355 derived variables such as thickness were calculated, producing 900 independent representations of the ocean state. From this ensemble, we derived the median estimate and the 95% confidence interval ($N=900$), defined by the 2.5% and 97.5% quantiles. This approach incorporates instrumental, sampling, and mapping uncertainties, as described in the subsection “Observation-based gridded products”.

360 For the ToE uncertainty estimation in Fig. 5f–g, we follow the method provided by Tan et al. (2026) and Lyu et al. (2014) to include the 95% uncertainty range of ToE with the IAP ensemble members. In short, the ToE is first estimated for each of the above 900 unique temperature–salinity realizations. Then, by excluding the outliers in other emergence directions that differ from the central estimate, the final estimate of ToE and its 95% data uncertainty range (the 2.5th percentile to the 97.5th



percentile) are defined. Finally, the ‘significant emergence at 95% confidence level’ is defined only if both the ToE and its 95% uncertainty range are earlier than 2024. Otherwise, it is defined as an ‘insignificant emergence’. More details about the definition could be found in the Extended Data Fig. 8 of Tan et al. (2026) and the Supplementary Figure S4 of Lyu et al. (2014).

2.5 Estimation of the steric height of the water mass

Steric height, defined as the vertical displacement of an isobaric surface relative to a deep reference level, reflects density variations integrated over depth that arise from temperature- and salinity-driven changes (Roquet et al. 2015). It serves as a key metric of water-mass heat and freshwater content, as its thermal and haline components quantify volumetric expansion due to heat uptake and contraction associated with salinity changes. Steric height was calculated using the GSW toolbox (McDougall and Barker 2011), with anomalies integrated from 1500 to 200 dbar for SACW and from 1500 to 600 dbar for AAIW (see the Appendix B for sensitivity tests with integration from 600 to 200 dbar for SACW).

3 Results

3.1 Trends of South Atlantic Surface Water (SW) since the 1980s

Since 1980, the South Atlantic Surface Water (SW) has undergone pronounced increases in temperature and salinity (Fig. 3a). Over the past 45 years, regional-mean temperature has risen by ~ 0.355 [0.177 – 0.534] °C and salinity by ~ 0.123 [0.061 – 0.1765] g kg⁻¹ (2020–2024 relative to 1980–1985; 95% ensemble uncertainty range). Multichannel Singular Spectrum Analysis (M-SSA) (Ghil et al. 2002; Manta et al. 2024) applied to both variables confirms that these changes reflect significant multi-decadal trends, with the trend mode (first principal component) explaining 19.8% of the variance in temperature and 20.5% in salinity (Fig. D4). These property changes are largely density-compensating: neutral density exhibits no notable long-term trend, although it shows pronounced interannual-to-decadal variability (Fig. 3a).

Spatially, the strongest warming and salinification occur within the South Atlantic Subtropical Current (Fig. 3b–c), with peak linear trends in the southern Brazil Current (Falkland/Malvinas Confluence) near 36–40°S reaching ~ 0.4 °C per decade and ~ 0.015 g kg⁻¹ per decade. These surface-intensified signals coincide with significant ($p < 0.05$) increases in net downward surface heat flux and evaporation-minus-precipitation (E–P) over the past three decades (Fig. 3e–f). Pronounced increases in temperature and salinity also emerge in many tropical regions (north of 15S), especially for the Angola Dome, where significant trends in both the increasing uptake of heat by the ocean and the increasing evaporation rate are present. The spatial coherence across subtropical and tropical regions indicates that strengthened air–sea fluxes are a dominant driver. The above-discussed mechanisms fill the physical knowledge gap of Tan et al. (2026) from the water mass perspective, who identified compound warming–salinization emergence signals in the epipelagic zone (0–200 m)—corresponding to SW—across the Benguela Current, Angola Dome, and Brazil Current since the 1980s. Over the same period, SW has thickened, leading to a



395 volume increase of $3.26 [1.37\text{--}4.87] \times 10^5 \text{ km}^3$ (Fig. 3a,d; Fig. D5a), which is concentrated in the Cape Basin and along
downstream pathways of the South Equatorial Current (Fig. 3d), and that most regions of increased thickness coincide with
enhanced downward surface heat flux (Fig. 3e). This coincidence is suggested by Donners et al. (2005) that air-sea heat uptake
is a potentially important factor contributing to SW volume expansion. Previous observational and tracer-based studies show
that Agulhas leakage and mesoscale or submesoscale intrusions transport warm, saline water through the Cape Basin and
400 across the South Atlantic, where they can depress isopycnals and redistribute upper-ocean volume (Barabinot et al. 2024; Beal
et al. 2011; Chen et al. 2022; Großelindemann et al. 2025; Gunn and Beal 2026; Laxenaire et al. 2020; Schubert et al. 2021).

Physically, the observed SW warming and salinization over the past 45 years arise from two primary mechanisms. First,
enhanced air-sea fluxes and vertical exchange—stronger downward heat flux and increased evaporation—have amplified
405 surface forcing (Fig. 3e–f), followed by vertical mixing and subduction into the upper thermocline (Leyba et al. 2019;
McDonagh and King 2005). These processes modify both heave (thickness/volume) and spice (T/S) components, consistent
with earlier work showing that surface warming is dominated by surface heat and freshwater flux perturbations at global and
regional scales (Cheng et al. 2022; Li et al. 2023b; von Schuckmann et al. 2023). Second, lateral advection from the Indian
and Southern Oceans contributes additional warming and salinization through the intrusion of Agulhas Rings and mesoscale
410 eddy mixing (Chen et al. 2022; Drouin and Lozier 2019).

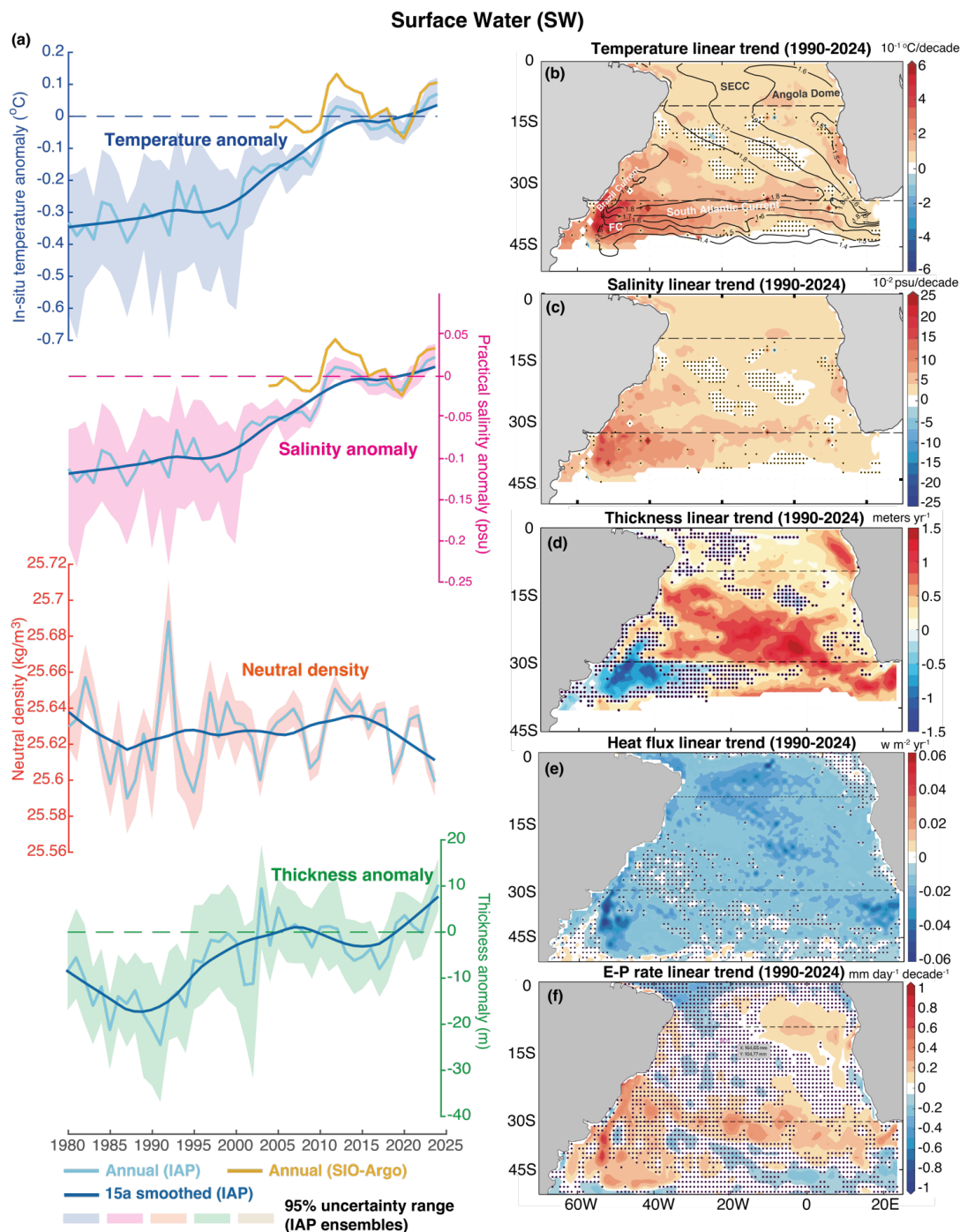


Fig. 3: Temporal evolution and spatial trends of South Atlantic Surface Water (SW; 45°S – 0°N ; $\gamma^n < 26.35 \text{ kg m}^{-3}$), based on IAP (Institute of Atmospheric Physics, CAS) and SIO-Argo monthly gridded products. (a) Regional-mean time series of SW temperature anomaly, practical salinity anomaly, neutral density, and layer-thickness anomaly. Shading represents the 95% uncertainty range of the IAP ensemble estimates, incorporating instrumental, sampling,



and mapping uncertainties (see “*Water-mass property uncertainty estimation*” in *Methods*). Dark blue and orange curves show the IAP (15a LOWESS-smoothed) and the SIO-Argo annual estimates (12-month average), respectively. All anomalies are relative to the 2005–2024 baseline. (b–f) Spatial linear trends for the period 1990–2024 in temperature, practical salinity, layer thickness, net surface heat flux (negative values denote the uptake of heat by the ocean), and evaporation-minus-precipitation (E–P) rate. Here, black dots indicate the insignificant trends at 95% confidence level ($p\text{-value} > 0.05$). The black contoured lines in Panel B are surface steric height climatology (2005–2024; 1500–0 dbar). FC: Falkland/Malvinas Confluence, SECC: South Equatorial Countercurrent.

3.2 Trends of South Atlantic Central Water since 1980s (SACW)

Figure 4 summarizes the temporal and spatial evolution of South Atlantic Central Water (SACW; $26.35 < \gamma^n < 27.10 \text{ kg m}^{-3}$) from 1980 to 2024. Prior to the 2000s, SACW exhibits a much larger observational uncertainty in temperature and salinity, with instrumental, sampling, and mapping errors exceeding the amplitude of internal variability (e.g., salinity variability of $\sim 0.01 \text{ g kg}^{-1}$ versus uncertainty of $\sim 0.08 \text{ g kg}^{-1}$; Fig. 4a). As a result, the apparent cooling and freshening before 2000 are smaller than, and not statistically distinguishable from zero because of large observational uncertainty. This reflects both the limited spatial coverage during the pre-Argo era and the inherently large T–S variability of this water mass (see Section 1). After 2005, when Argo substantially improved observational coverage and data quality, a weak but detectable warming and salinization signal emerges (Fig. 4a), although its magnitude remains modest relative to that observed in SW. Over the full 1980–2024 period, neutral density shows no notable trend.

In contrast to the limited and uncertain temperature and salinity changes, our ensemble-based analysis indicates a coherent positive trend in SACW thickness (Fig. 4a). The layer has thickened by 35.93 [17.83–53.37] m, corresponding to a 13.69% [7.31%–20.33%] expansion in volume (from $\sim 9 \times 10^6 \text{ km}^3$ in 1980–1985 to $\sim 10.4 \times 10^6 \text{ km}^3$ in 2020–2024; Fig. 4a; Fig. D5b), with the strongest increases south of 30°S , within the subtropical gyre and South Atlantic Current (Fig. 4b). These thickening signals are in a higher confidence than the T/S trends: for temperature and salinity, the uncertainty bounds are large enough that trend estimates can differ in sign, whereas thickness and volume trends remain consistently positive across the full uncertainty range. This is because the uncertainty of the boundary depth of the water mass is linked to the high stratification that can effectively ‘anchor’ the interface, leading to the minor vertical displacements that are less sensitive to the temperature and salinity observation errors (McDougall and Krzysik 2015). Spatially, significant positive trends ($p < 0.05$) in thickness occur across most of the basin, with maximum increases ($\sim 30 \text{ m per decade}$) in the southern subtropical gyre and near the Subantarctic Front, close to the formation region of the Subantarctic mode water (SAMW) (Fig. 4b). A meridionally oriented band of variability reflects strong interannual–decadal fluctuations (Fig. 4e). Consistently, the steric height at 200 dbar (representative of the approximate SACW upper-bound depth) relative to 1500 dbar displays an almost identical long-term increase in the M-SSA reconstruction (Fig. 4c–d). Because layer-thickness changes reflect vertical displacements of neutral-density surfaces (see Section 1), these results indicate a dominant heave signal, with the lower SACW boundary shifting downward through time (Fig. 4f).



To further separate heave and spice contributions, we examined geostrophic volume transport across 34.5°S referenced to the observed mean velocity at 1000 dbar, estimated from Argo float parking pressure (Michel et al. 2026). We find no substantial long-term increase in northward SACW transport (or associated heat or freshwater transport) since the 1980s at 95% confidence level, although pronounced interannual–decadal variability persists (Fig. D6). This result agrees with Caínzos et al. (2022), who report no substantial long-term trends in mass transport at 30°S based on repeat cruises. The absence of transport trends arises because insignificant trend in meridional geostrophic velocities offsets the increased cross-sectional area resulting from layer thickening (Fig. D6). Combined with the negligible long-term trends in temperature and salinity on neutral surfaces, this indicates that neither isopycnic component of advection nor isopycnic mixing has made a substantial contribution to SACW “spice” changes over the study period.

In the upstream formation area of SACW (e.g., east of the Drake Passage and within the ACC/Polar Front of the Southern Ocean related to SAMW), the vertical subduction associated with meridional excursions of the ACC is one of the main drivers of the volume increase (Downes et al. 2017; Portela et al. 2020). The enhanced subduction and the resulting heave adjustment in the upstream mode-water formation region can ultimately lead to the stronger downstream heave signal observed in the South Atlantic Basin. In addition, an increasing number of eddies injected into the basin from the Polar Front mode water formation region resulting to the dynamic height anomaly might be another reason (Beal et al. 2011; Gunn et al. 2020). Such heave signals are also consistent with wind-stress forcing and climate-driven changes in Ekman downwelling (vertical advection) and large-scale circulation over the past ~30 years (Drouin et al. 2021; Häkkinen et al. 2016; Kolodziejczyk et al. 2019). In short, the greater confidence long-term change in SACW is a heave-driven thickening, whereas long-term property changes arising from isopycnic processes remain minimal.

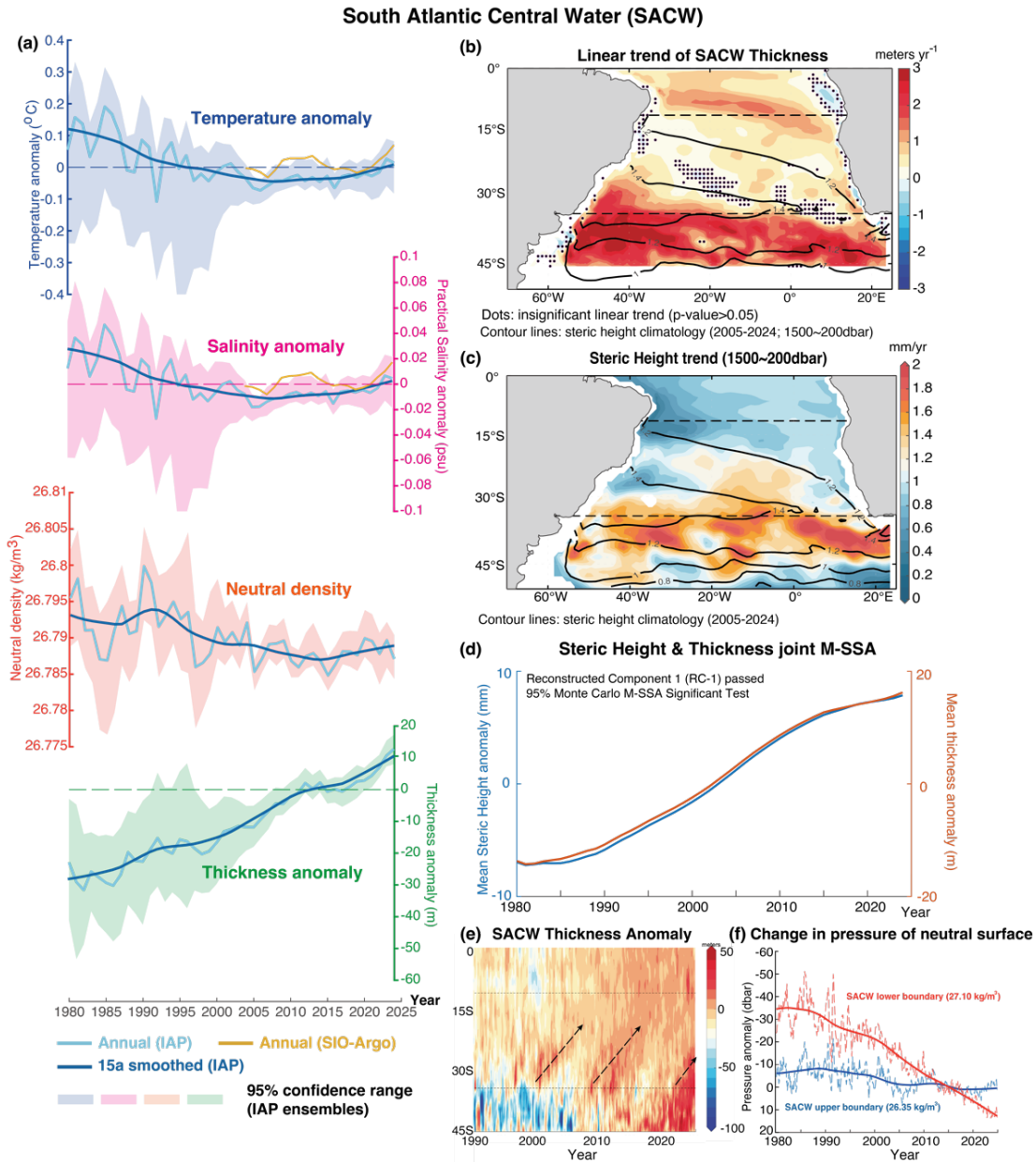


Fig. 4: Temporal evolution and spatial trends of properties of South Atlantic Central Water (SACW; 45°S–0°N; 26.35 < γ^n < 27.10 kg m⁻³) based on the IAP monthly gridded atlas and SIO-Argo monthly gridded products. (a) Same as Fig. 3a, but for SACW: regional-mean anomalies of temperature anomaly, practical salinity anomaly, neutral density, and layer thickness anomaly. All anomalies are referenced to the 2005–2024 baseline. (b) Spatial linear trend (1990–2024) of SACW layer-thickness anomaly. The black dots indicate insignificant trends at 95% confidence level (p-value > 0.05), with steric height climatology (2005–2024) contoured. (c) Spatial linear trend (1990–2024) of steric height integrated from 1500 to 200 dbar (representative of the approximate SACW upper-bound depth), with steric height climatology (2005–2024) contoured. (d) Basin-mean reconstructed trends of steric height and layer thickness obtained using the multichannel singular spectrum analysis (M-SSA). Shown is the first reconstructed component (RC1; long-term



change). (e) Zonal–temporal evolution (1980–2024) of SACW thickness. (f) South Atlantic mean changes in the pressure (depth) of SACW isopycnals, represented by the neutral-density surfaces $\gamma^n = 26.35 \text{ kg m}^{-3}$ (upper bound) and $\gamma^n = 27.10 \text{ kg m}^{-3}$ (lower bound).

3.3 Trends of core Antarctic Intermediate Water (AAIW) since the 1980s

Over the past 45 years, the core of AAIW has undergone pronounced changes in its physical properties (Fig. 5). The salinity minimum (Smin) shows an apparent freshening prior to the 2000s, consistent with earlier cruise-based analyses reporting freshening between 1980 and 2000 (McCarthy et al. 2011). However, pre-Argo observational uncertainties (combined instrumental, sampling, and mapping errors) are substantially larger than those of the post-Argo era, making the early freshening and its variability statistically indistinguishable from noise. After the 2000s, when observational coverage improves markedly, a weak but robust salinification emerges, with Smin increasing by $0.003 [-0.005-0.010] \text{ g kg}^{-1}$ over decadal scales (2020–2024 relative to 2000–2005). In contrast, temperature changes at the Smin level are much more substantial than salinity changes. The AAIW core has warmed by $0.26 [0.17-0.35] ^\circ\text{C}$ since the 1980s (2020–2024 relative to 1980–1985; Fig. 5a), with the strongest increases occurring along the pathways of the subtropical gyre, South Atlantic Current, and southern South Equatorial Current (Fig. 6a). This warming drives a marked decrease in neutral density dominated by thermal expansion (Fig. 5a,c), with the thermal contribution accounting for more than 80% of the total density change (Fig. A1).

Notably, early indications of anthropogenic warming emerge in the core properties of AAIW. Time-of-emergence (ToE) estimates, defined as the time at when the long-term signal rises above short-term variability, show that 70.29% [39.06%–90.69%] (95% confidence range) of the South Atlantic exhibits detectable AAIW warming since the 1990s relative to a 1960–1979 baseline (Fig. 5d,f,g). A similarly early emergence occurs in decreasing neutral density (Fig. 5e). The earliest ToE values appear near $\sim 30^\circ\text{S}$ within the subtropical gyre (Fig. 5f), providing a strong basin-scale fingerprint of anthropogenic influence on intermediate waters. Latitude–time evolution further reveals a south-to-north progression of the warming signal from roughly 35°S to 10°S over the past ~ 30 years (Fig. 5b). In addition to these property changes, the AAIW layer has thinned by $15.11 [1.03-35.46] \text{ m}$ since the 1980s, indicating a contraction of the water-mass core (Fig. 5a). This thinning corresponds to a volume decrease of approximately $4.60 \times 10^5 \text{ km}^3$ (a 2.5% reduction relative to the 1980s) across the South Atlantic (Fig. 5a).

A central question concerns the origin and subsequent northward propagation of the potential anthropogenic warming signal detected in the core of AAIW. As noted in the Section 2.3, AAIW is formed in three primary regions: (i) Indian-origin I-AAIW entering the basin via Agulhas leakage, and (ii) the A-AAIW is sourced from the Southern Ocean/Pacific Ocean via the Drake Passage, and (iii) also sourced from the transformation from the Subantarctic mode water (SAMW) in the north of the Subantarctic Front (e.g., along the Falkland/Malvinas Current and South Atlantic Current) (Rusciano et al. 2012). In the Southern Ocean, the formation of A-AAIW is ventilated within the upper branch of the meridional overturning circulation,



where surface buoyancy forcing (sea-ice melt and sea-ice-driven freshwater) dominates the AAIW density variation
 515 (Abernathy et al. 2016).

To begin with, regarding the AAIW source regions, the warming is the combined effect of basin-external source pathways and a basin-internal formation region via isopycnic advection. Specifically, from the perspective of all these three source regions, we found that: (1) For the Indian-sourced regions where I-AAIW is formed and subsequently enters the South Atlantic through
 520 Agulhas rings, there is a pronounced imprint warming signal from the South Indian Ocean subtropical gyre and Mozambique Channel (Fig. 6a), implying that inter-basin transport plays an important role in importing upstream warming into the South Atlantic along isopycnic advection. One possible reason is due to the strengthened and southward-shifting westerlies associated with the increasing positive phase of the Southern Annular Mode (SAM) (Großelindemann et al. 2025) in the Southern Ocean, the multidecadal increase in Agulhas leakage was detected by the observations and model simulations, finally leading to an
 525 increasing mesoscale eddy transport with the warming signal across the GoodHope Section (Durgadoo et al. 2013; Loveday et al. 2015). (2) For the Pacific-sourced Atlantic AAIW branch (A-AAIW) entering via the Drake Passage, which is largely steered by the ACC, however, exhibits cooling trends (although it is not significant at 95% confidence level; Fig. 6a) in the advection pathway, suggesting that advection of southeast Pacific AAIW into the Atlantic Southern Ocean does not seem to make a dominant contribution to the observed northward warming propagation of AAIW in the South Atlantic basin. (3) From
 530 the perspective of the basin-internal formation region in the southwestern South Atlantic (near the Falkland/Malvinas Current and the South Atlantic Current), we found that the locally ventilated A-AAIW has experienced substantial warming (Fig. 6a), with the maximum magnitude of a positive temperature linear trend reaching ~ 0.15 °C/decade. This is likely due to changing air–sea exchanges, which could be indicated by the strengthened downward surface heat fluxes and mixed-layer warming over recent decades, despite regional differences in statistical significance (Purple box in the Fig. 6b–c). Further downstream, in the
 535 Cape Basin and adjacent southeastern South Atlantic, intense isopycnic mixing between A-AAIW and I-AAIW blends these two warming signals, forming Indo-Atlantic AAIW (IA-AAIW), which then contributes to the basin-scale propagation of the AAIW warming signal with isopycnic advection and isopycnic mixing (Fig. 6a).

Then, from these two regions, warming signals from the SAMW water mass transformation (related to the A-AAIW) and
 540 warming signals from the Indian Ocean (related to the I-AAIW) together subsequently drive the warmer and less dense waters advected into the South Atlantic basin (cross 34.5S) along the neutral density surface. The associated subduction and northward propagation of anomalies can be traced by the steric height isolines aligned with the subtropical circulation pathways (Fig. 6a). More specifically, the warming signal associated with fresher A-AAIW follows the South Atlantic Current and the subtropical gyre into the Brazil Current system, where it is carried northward along the western boundary by the isopycnal advection
 545 (Stramma and England 1999). Meanwhile, the warming signal associated with saltier I-AAIW propagates from the Agulhas leakage region—augmented by eddy-driven mixing—into the Benguela Current and its northward extension (Laxenaire et al. 2018; Stramma and England 1999). These anomalies subsequently feed the Southern South Equatorial Current (SSEC), which flows northwestward between $\sim 20^\circ\text{S}$ and 30°S and mediates the bifurcation between northward- and southward-flowing



boundary currents at the western margin (Schmid 2014; Stramma and England 1999), thereby transporting the anomalies
550 toward the equatorial current systems. In addition, isopycnal mixing driven by mesoscale eddy turbulence, especially along
the pathways of the Agulhas System and the Falkland/Malvinas Current (FC), is another key process that contributes to the
northward spread of the AAIW anthropogenic warming signal, as shown by both observations (Sévellec et al. 2025) and model
simulations (Capuano et al. 2018). In contrast, vertical diapycnal mixing plays only a negligible role in shaping these changes:
diapycnal heat flux contributes less than 1% of the total AAIW heat-content change ($\sim 0.55 \text{ ZJ yr}^{-1}$ versus $\sim 360 \text{ ZJ yr}^{-1}$; Fig.
555 C1; consistent with Portela et al. (2020)). Taken together, the alignment of steric height with these circulation routes highlights
the along-isopycnal processes (both isopycnal advection and isopycnal mixing) as the dominant mechanism transporting
potential anthropogenic heat from high to low latitudes within South Atlantic intermediate waters.

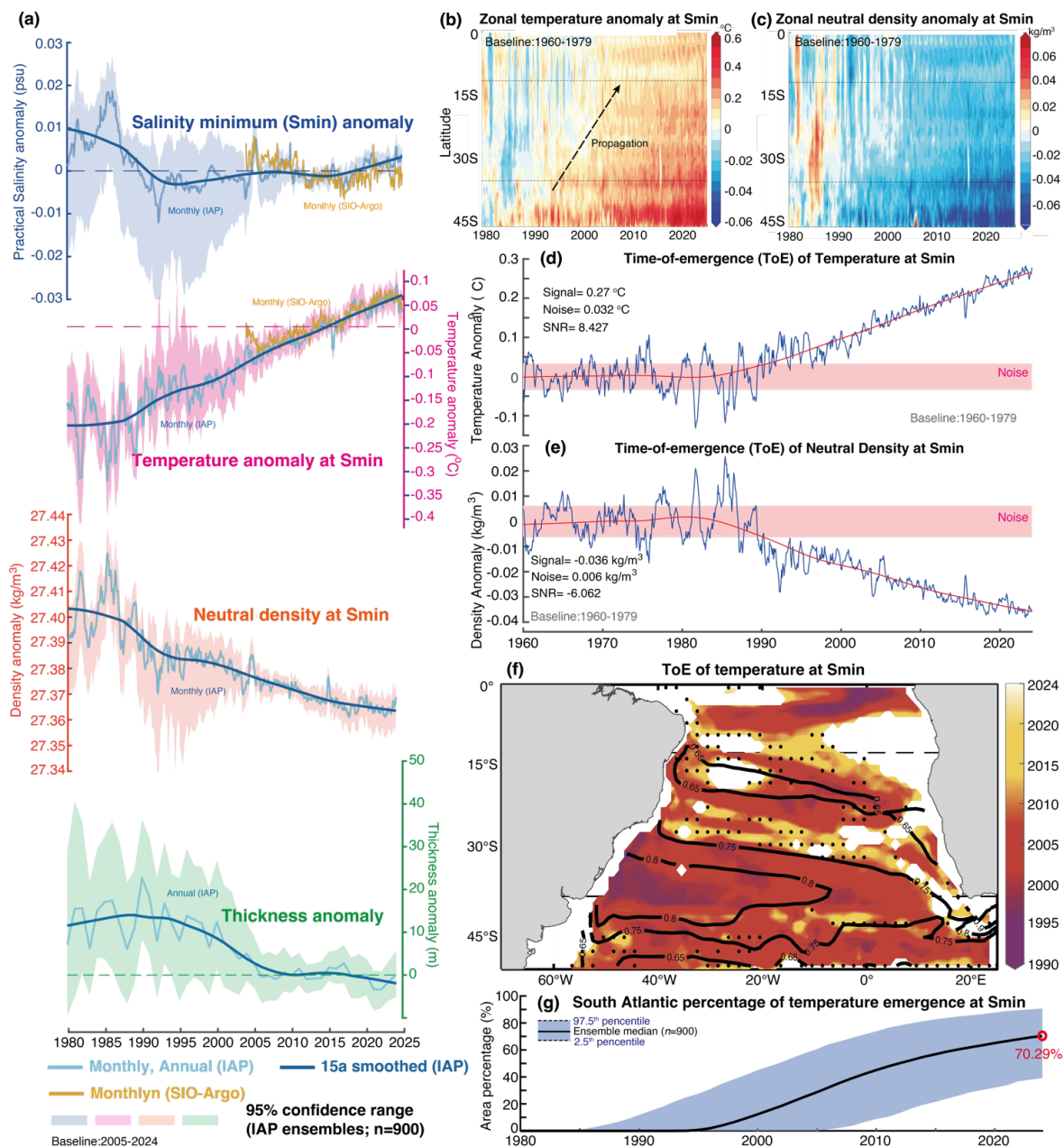


Fig. 5: Evolution of Antarctic Intermediate Water (AAIW) properties (45°S – 0°N ; $27.10 < \gamma^n < 27.60 \text{ kg m}^{-3}$; the core of AAIW is defined at the salinity-minimum depth) based on IAP monthly gridded products. (a) Same format as Fig. 4a, but showing the time series of practical salinity minimum (S_{min}) anomaly, temperature anomaly at the S_{min} level, neutral-density at the S_{min} level, and layer-thickness anomaly. All anomalies are referenced to the 2005–2024 baseline. (b–c) Latitude–time evolution of anomalies (baseline: 1960–1979) in temperature and neutral density at the S_{min} level. (d–e) Time-of-emergence (ToE) estimates for the regional-mean temperature and neutral density at the S_{min} level over 1960–2024. Red lines show the long-term (anthropogenic) signal (25-year LOWESS-smoothed); shaded red bands denote the noise envelope. Signal-to-noise ratios (SNRs) are indicated (see *Materials and Methods* for signal and noise



definitions). (f) Spatial distribution of ToE for warming at the S_{min} level, overlaid with steric height climatology (600–1500 dbar) shown as black contours. The white color indicates no temperature emergence before 2024. Areas with insignificant emergence before 2024 (definitions see the *Method* section) are marked as black dots. (g) is the same as (f) but shows the South Atlantic basin percentage area (45S–0N; 60W–25E) of emergence as a function of the year with the shaded areas representing the data uncertainty (sampling error, instrument error and mapping error) arising from ensemble members of the observational-based data products.

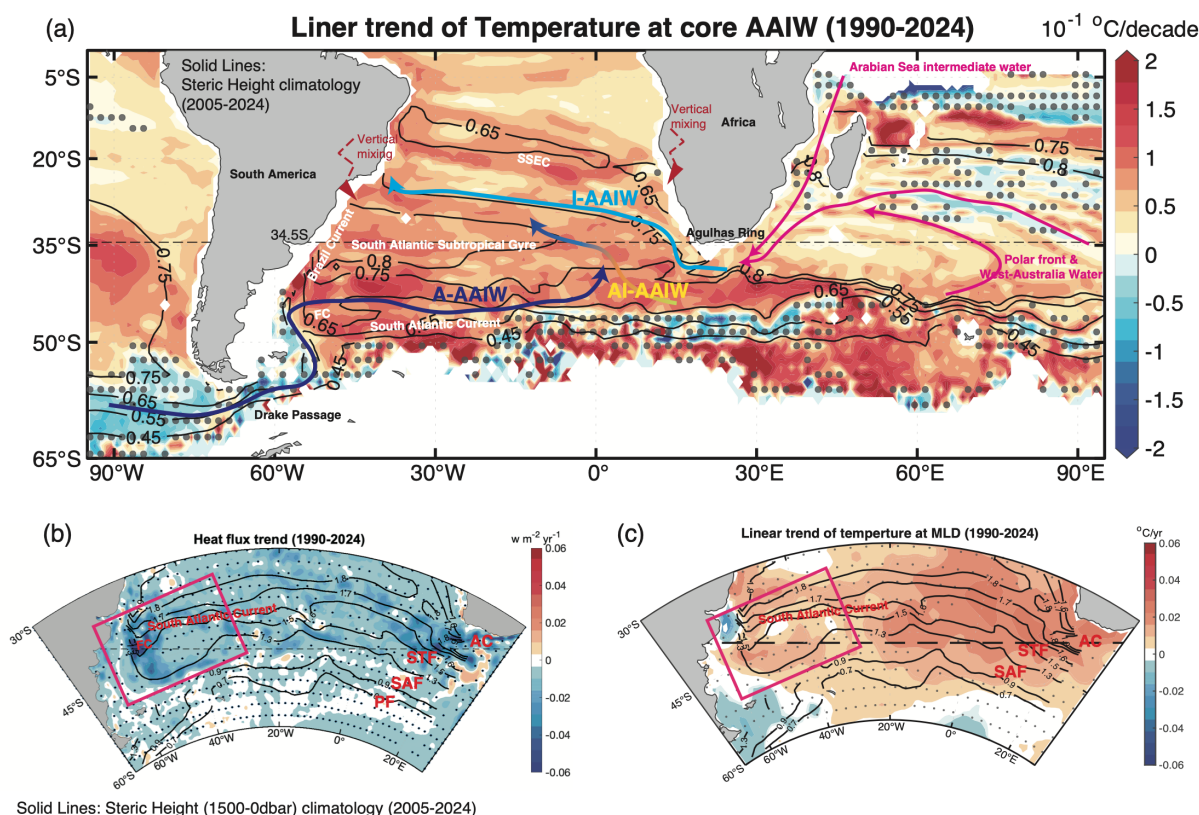


Fig. 6: The origin and propagation of the early indications of potential anthropogenic warming signatures into the South Atlantic basin interior in the intermediate layer. (a) Spatial linear trend (1990–2024) of temperature at the AAIW salinity-minimum depth (S_{min}), overlaid with steric height climatology (600–1500 dbar) shown as black contours, depict the principal AAIW circulation pathways south of 11°S, including the Brazil Current, Falkland/Malvinas Current (FC), South Atlantic Current, Benguela Current, Agulhas Current (AC), and the Southern South Equatorial Current (SSEC). (b) is the same as (a), but showing spatial linear trends (1990–2024) in net surface heat flux (ERA5) over AAIW outcrop and subduction regions (AC: Agulhas Current; STF: Subtropical Front; SAF: Subantarctic Front; PF: Polar Front). Solid black contours show steric height climatology integrated from 0–1500 dbar. Black dots indicate the insignificant trends at 95% confidence level ($p\text{-value} > 0.05$). (c) is the same as (b), but for spatial linear trends (1990–2024) in mixed-layer temperature. Mixed-layer depth (MLD) is defined as the depth at which potential density exceeds its value at 10 m by 0.03 kg m^{-3} .



4 Conclusions and discussions

In the pre-Argo era, the South Atlantic benefited from high-quality ship-based observations along repeated hydrographic sections. Over the past decade, the development of a coordinated regional ocean observing system—together with advances
 590 in the recovery, quality control, and processing of historical data—has enabled the creation of more spatially uniform and temporally continuous gridded products extending back to the 1980s. Integrating repeated ship-based transects with these observation-derived gridded datasets now allows systematic, basin-scale, and high-resolution tracking of water-mass evolution since the 1980s. Building on Tan et al. (2026), this study provides the first basin-wide assessment of 45 years of individual and compound climatic impact-drivers (temperature coupled with salinity, and the coupling effects on ocean physical state) in
 595 upper-ocean water masses of the South Atlantic (SW, SACW, and AAIW) using multiple state-of-the-art gridded products, complemented by ship-based data and rigorous uncertainty estimates (Fig. 7). In conclusion, our comprehensive investigation of the joint evolution of temperature, salinity, and derived variables yields three main results:

(1) Long-term substantial physical changes of upper-ocean water masses. All major upper-ocean water masses exhibit
 600 substantial ~45 years changes. The SW has warmed and markedly increased in salinity since the 1980s, accompanied by column thickening and volume expansion. SACW has similarly thickened and increased in volume, consistent with the steric height isolines. AAIW has become warmer, lighter, and thinner, with its density decrease dominated by thermal expansion. We further identify early indications of potential anthropogenic warming in intermediate waters, with about 70.29% [39.06%–90.69% at 95% confidence] of the basin displaying “unusual and unfamiliar” changes since the 1990s (in the
 605 terminology of Frame et al. (2017)).

(2) Key regions driving the changing water-mass structure. The strongest multi-decadal trends since the 1980s occur in two principal regions: the South Atlantic subtropical gyre (e.g., near 34.5°S) and the eastern tropical gyre (e.g., the Angola Dome). Together with the influence of the Agulhas System, which facilitate eddy-driven mixing and property exchange with Indian
 610 Ocean waters (Chen et al. 2022), these regions emerge as critical nodes for reorganizing the distribution of heat and freshwater across water masses along the pathway of the upper limb of AMOC. Their role as climate-change “hotspots” carries important implications for ecosystem resilience, biogeochemical functioning, and regional climate risk assessments.

(3) Physical processes governing long-term individual and compound (temperature coupled with salinity) state changes. Our
 615 analysis identifies the dominant mechanisms driving individual and compound (temperature-salinity coupled) ocean state changes in the South Atlantic, firstly documented in Tan et al. (2026): In the epipelagic zone (0–200 m), compound warming–salinization is controlled primarily by intensified air–sea fluxes and enhanced evaporation affecting SW, followed by the strengthened inflow of warmer and saltier waters transported from the Agulhas system. In the mesopelagic zone (200–1000 m), the drivers differ between water masses: SACW changes are dominated by heave-driven vertical displacement
 620 (thickening), a signal that is detected with greater confidence than the corresponding property (spice) changes across the full



uncertainty range, whereas AAIW changes arise largely from spice-driven processes associated with the northward propagation of potential anthropogenic heat from the mode water transformation region around the Falkland/Malvinas Current, the formation regions in the Southern Ocean and as well the Indian Ocean via isopycnal advection and isopycnal mixing into the basin interior (Fig. 6a).

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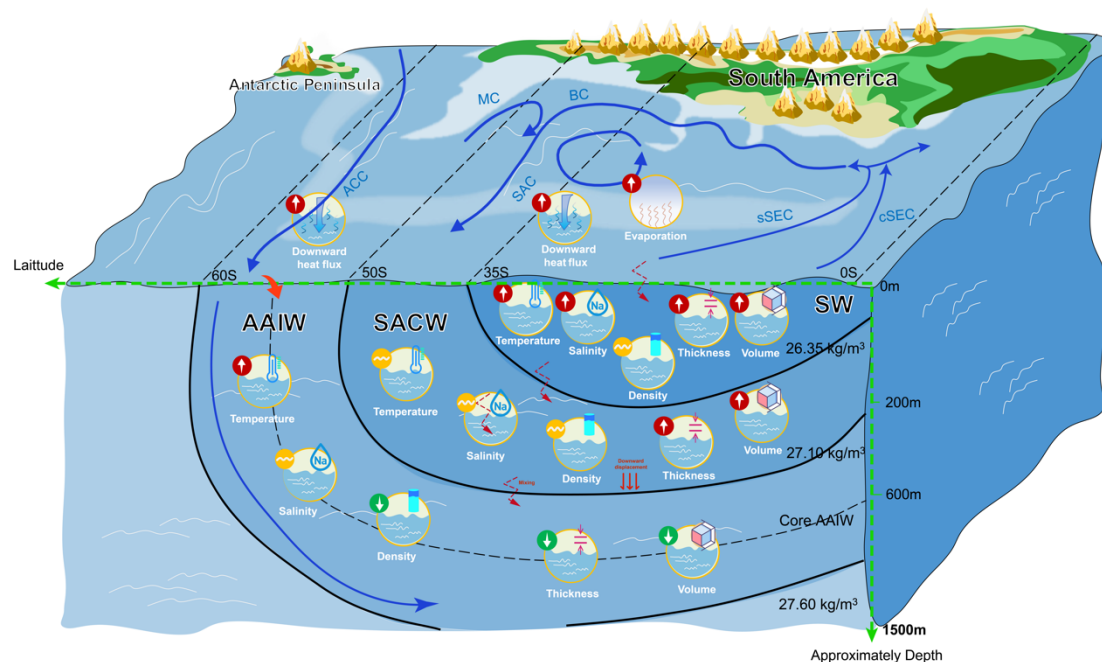


Fig. 7: Schematic summary of the observed changes in South Atlantic upper-ocean water masses (SW, SACW, AAIW) over the past 45 years. Red upward arrows indicate regional increases in the corresponding properties or processes; green downward arrows indicate regional decreases; yellow wavy lines denote no robust or no statistically confident long-term change. Major surface currents are shown (sSEC: Southern South Equatorial Current; cSEC: Central South Equatorial Current; BC: Brazil Current; SAC: South Atlantic Current; MC: Malvinas Current; ACC: Antarctic Circumpolar Current). Neutral-density climatology contours delineate the boundaries of the water masses in the depth–latitude section. Coordinates are approximate and not intended to represent exact geographical positions.

We acknowledge that using a ~45-year analysis window may, in principle, introduce contamination from internal variability operating on similar timescales (e.g., the Atlantic multidecadal oscillation-AMO or Southern Annular Mode-SAM). However Tan et al. (2026) demonstrated that such effects have negligible influence on long-term temperature and salinity trends, on ToE estimates, and on the fundamental conclusions of this study. With the increasing availability of high-quality observations, future work can extend the analysis to deeper and denser water masses—including Circumpolar Deep Water (CDW), North Atlantic Deep Water (NADW), and Antarctic Bottom Water (AABW)—across broader spatial and temporal scales (Johnson and Purkey 2024; Palmer et al. 2019), as well as can extend to other basins, such as the North Atlantic or the North Pacific —



which are also regarded as the hotspot regions that are the most impacted by the compound state changes (Tan et al. 2026). While we report early potential indications of anthropogenic warming signatures in AAIW core, we did not conduct a formal attribution analysis. Future studies should address this gap using CMIP6 or upcoming CMIP7 simulations (Kwiatkowski et al. 2020).

645

Despite substantial progress in data recovery and processing, large observational uncertainties remain in some water-mass property estimates, particularly in the pre-Argo period (e.g., temperature and salinity time series in Fig. 4a, salinity time series in Fig. 5a). A key priority is to continue improving spatial and temporal sampling. Progress will rely on strengthening and expanding existing international observing efforts, including GCOS (Zemp et al. 2022), GOOS (Moltmann et al. 2019), and

650 SAMOC.

Since the relevance of temperature and salinity of a water mass is within the frameworks of the Climatic Impact-Drivers (Ruane et al. 2022; Tan et al. 2026) that provides a link between water-mass evolution and broader questions of ocean–climate adaptation and sustainability in the ocean ecosystem, the coupling of warming with salinity change may strongly affect marine ecosystems in the South Atlantic, particularly in the two hotspot regions where physical changes are most pronounced. Elevated temperature can stress marine ecosystems, intensify marine heatwaves, and increase the mortality of marine organisms (Cheng et al. 2022). Shifts in salinity, whether freshening or salinization, can displace species distributions, modify nutrient fluxes, and disrupt biological functioning (Röthig et al. 2023). Additive and synergistic effects of temperature and salinity coupling (Tan et al. 2026) may influence primary production and propagate through food webs, affecting fish communities and fisheries across the region, including the Argentine Basin and the oligotrophic subtropical gyre (Pinho et al. 2021; Sant’Ana et al. 2025). One illustrative example is the mangrove habitat in the subtropical southwestern Atlantic near the Brazil Current, where combined warming and salinity changes are expected to reduce propagule density (e.g., mangrove seeds or seedlings) and alter dispersal trajectories, potentially reducing mangrove forest (Van der Stocken et al. 2022). Future studies can investigate such compound impact comprehensively, both at the level of individual species and across entire ecosystems.

665

Previous studies establish that Southern Ocean surface forcing and water-mass transformation ventilate SAMW and AAIW, while the Agulhas system supplies Indian Ocean thermocline and intermediate waters to the South Atlantic; these waters contribute to the northward upper limb of the overturning circulation (Bronse laer et al. 2020; Li et al. 2023b). Our observational analysis shows that waters between 65S and 40S within the SACW and AAIW density ranges have become warmer, fresher, and lighter, and that the spatial pattern of AAIW-core warming is aligned with climatological circulation pathways (Fig. 6). They are consistent with source-region change followed by northward export. The Indian Ocean enters the South Atlantic through the Agulhas system. However, it remains unclear how the influences of the Indian Ocean and/or the Southern Ocean on the South Atlantic may change under a warming climate, for example, what is the percentage of the Indian-origin water, Atlantic-origin water, Pacific-origin water, respectively, and how do these percentage change over the past decades. Quantifying how these source fractions and their thermohaline properties evolve will be necessary to attribute basin-scale

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AAIW change to individual pathways, which is a priority for future research. Recently, the coupling between dissolved oxygen and salinity within South Atlantic AAIW has been used as a fingerprint of climate-driven physical-biogeochemical co-variability linked to Indian Ocean inflow (McCarthy et al. 2011), which can be one of the potential tools for this priority.

680 Compound ocean-state changes also encompass dissolved oxygen (Tan et al. 2026), an essential variable linked to ecosystem functioning, biological carbon storage, water-mass age, and ocean ventilation (Davila et al. 2023). The analysis presented here did not extend to include dissolved oxygen yet. However, as temperature, salinity, and dissolved oxygen are tightly coupled through joint, causal, and composite relationship (Tan et al. 2026) (e.g., via remineralization, warming, and ventilation changes), integrating high-quality oxygen measurements into water-mass frameworks is a critical next step (Sanders et al. 685 2025). Ongoing community efforts are substantially improving the quality, availability, and accessibility of dissolved oxygen observations (Du et al. 2026; Gouretski et al. 2024; Ito et al. 2026; Ito et al. 2025; Wang et al. 2025; Zhou et al. 2025), opening the door to more comprehensive assessments in the near future.



690 Appendix A

Method to quantify the contribution of temperature and salinity to the neutral density changes

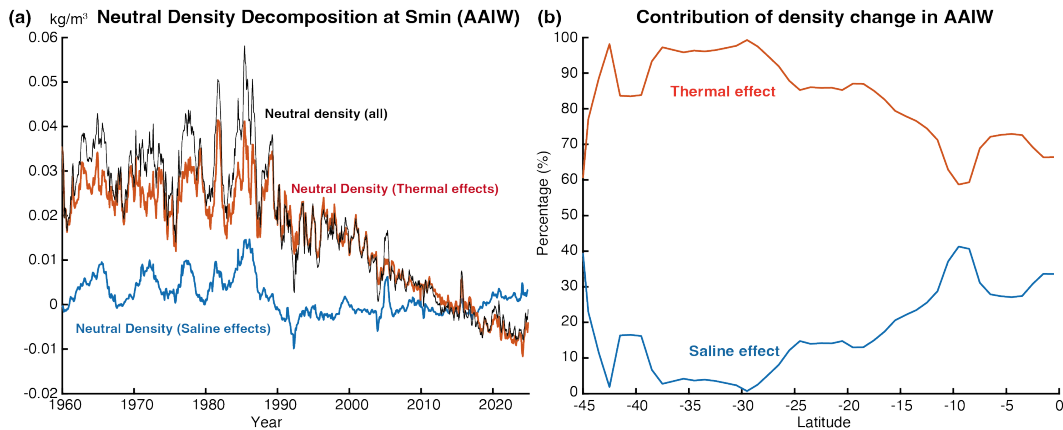
Owing to the nonlinearity of the neutral density equation, the contributions of temperature and salinity to the density change cannot be fully separated. Therefore, similar to Li et al. (2020), we compute the contributions of the in-situ temperature (Θ) and absolute salinity (S_A) to the change in neutral density as follows:

$$T: \gamma_\theta = \gamma(\bar{\theta} + \theta', S_A)$$

$$S: \gamma_s = \gamma(\bar{\theta}, \bar{S}_A + S_A')$$

where, γ_θ and γ_s represents the neutral density change related to thermal and saline effects, respectively. Here, overbars and primes denote climatological mean and deviations from the mean. In details, we used monthly conservative temperature and baseline climatology (2005-2024 in this study) absolute salinity to calculate γ_θ , and then calculate the linear trend of γ_θ .

700 Similarly, the monthly absolute salinity and baseline climatology conservative temperature, to calculate γ_s , and then calculate the linear trend of γ_s . The result is shown in Fig. A1.



705 **Figure A1: The decomposition of neutral density changes at core AAIW (Smin). (a) is the temporal decomposition. The black line denotes the anomaly timeseries (relative to 2005-2024) without decomposition (same as the orange line in Fig. 6a), the red line denotes the anomaly timeseries only take into account of the thermal effects (only temperature contribution), while the blue line denotes the one that only take into account of the saline effects (only salinity contribution). (b) is the percentage contribution of density change as a function of latitude in the South Atlantic. The decomposition methodology is shown below.**

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Appendix B

Sensitivity tests of the threshold choice for the estimation of the steric height

In this study, we used steric height at 200 dbar relative to 1500 dbar, that is, integrated over the 1500 to 200 dbar layer, to represent the large-scale geostrophic circulation associated with SACW. We acknowledge that integration over the 1500 dbar to 200 dbar layer will inevitably include some contribution from the geostrophic circulation signal of AAIW. Here, the



sensitivity tests shows that the results are not sensitive to whether the reference level is set at 600 dbar or 1500 dbar. In detail, figure B1 shows that the relative geostrophic circulation at 200 dbar is dominated primarily by the 200 to 600 dbar layer, which corresponds to the mean depth range of SACW. Therefore, following the recommendations of Talley (2011), we adopted steric height at 200 dbar relative to 1500 dbar in the main-text discussion of the large-scale geostrophic circulation of SACW. The same consideration was applied in the discussion of AAIW, for which steric height at 600 dbar relative to 1500 dbar was used.

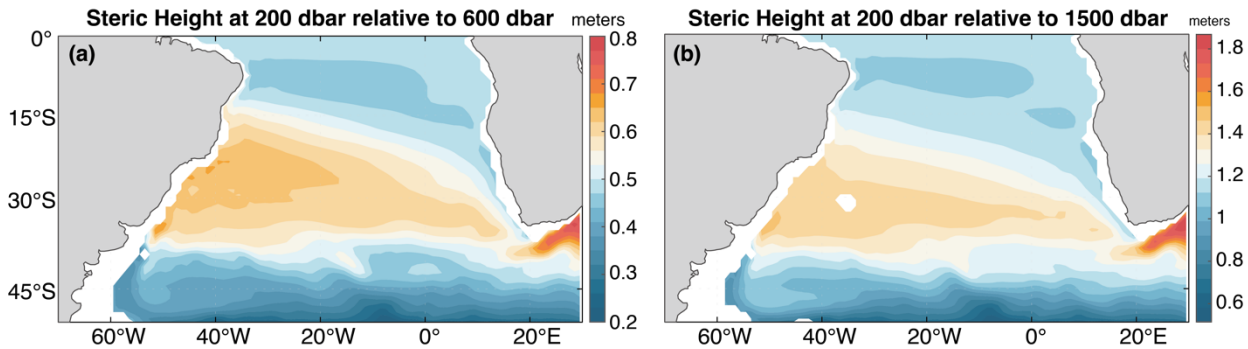


Figure B1. Sensitivity tests of the threshold choice for the steric height. (a) steric-height climatology (2005–2024) at 200 dbar relative to 600 dbar (integrating over 600–200 dbar); (b) is the same as (a), but relative to 1500 dbar. Here, the IAP temperature and salinity gridded product is used.

Appendix C.

730 Method to calculate the diapycnal heat flux and vertical transport (Turbulent Mixing)

The method to calculate the diapycnal heat flux and vertical transport (Turbulent Mixing) is followed by Ferrari and Wunsch (2009); Osborn (1980). In details, the turbulent (eddy) heat fluxes are parameterized as:

$$F = -K_T \frac{\partial T}{\partial Z}$$

Where K_T is the turbulent eddy diffusivity (thermocline), with the value ranging from 10^{-4} (open ocean interior, boundary zones) to 10^{-6} m^2/s (deep ocean)(Osborn 1980). In our study, we use $K_T = 10^{-5}$ m^2/s . With the heat flux, we then estimate its vertical transport by multiple the C_p (3985J/(kg*°C)) and sea water potential density, and then integrate (sum) in the whole South Atlantic with full time within the AAIW.

Consuming the total heat in AAIW (OHC) is the sum of (i) advection heat transport along the density surface and (ii) diapycnal (vertical) heat transport by turbulent mixing, The Fig. C1 demonstrates that the contribution of diapycnal heat transport is only 0.15% (the magnitude of annual diapycnal heat transport to the AAIW is about 0.55ZJ, while AAIW annual heat content is about 360ZJ in average). If $K_T = 10^{-4}$ m^2/s , the percentage will be 1.5%, and if $K_T = 10^{-6}$ m^2/s , the percentage will be 0.015%. Therefore, regardless of the value of K_T , it doesn't change the narrative of this study.

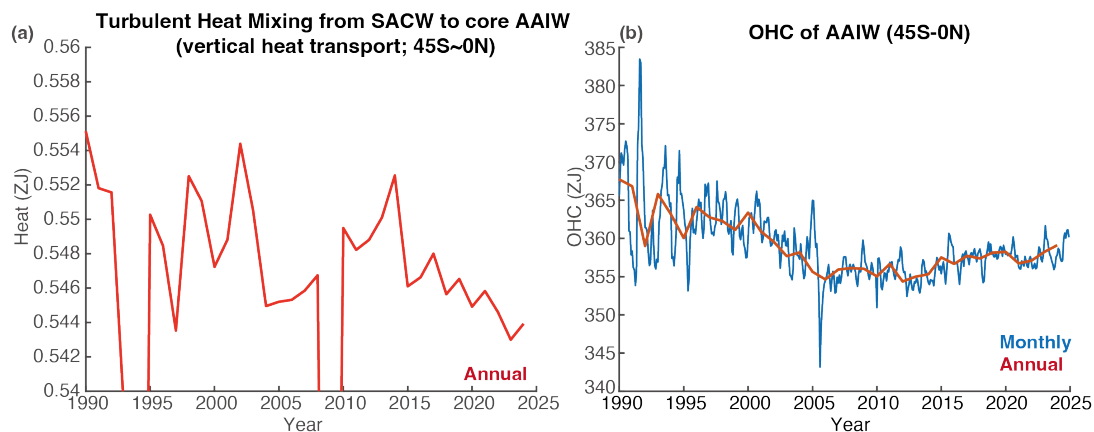


Figure C1: The diapycnal heat flux and vertical transport (Turbulent Mixing). (a) Diapycnal annual heat transport down to the AAIW (from SACW) represented by the turbulent heat mixing. (b) the annual OHC in AAIW.

Appendix D

Supplementary Figures

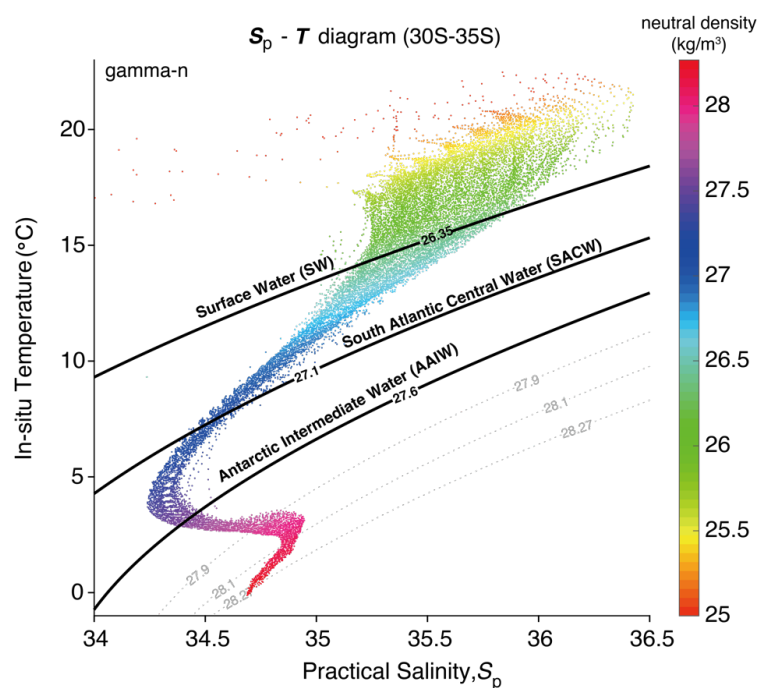


Figure D1: Typical T- S_p diagram in the South Atlantic (30S-35S) based on the IAP T/S monthly grid products (climatology plot; 1960-1979). Upper-ocean water masses acronyms and the neutral density (γ^n) boundaries are indicated. The color denotes the neutral density. For references to the water masses see Fig. 1c-d.

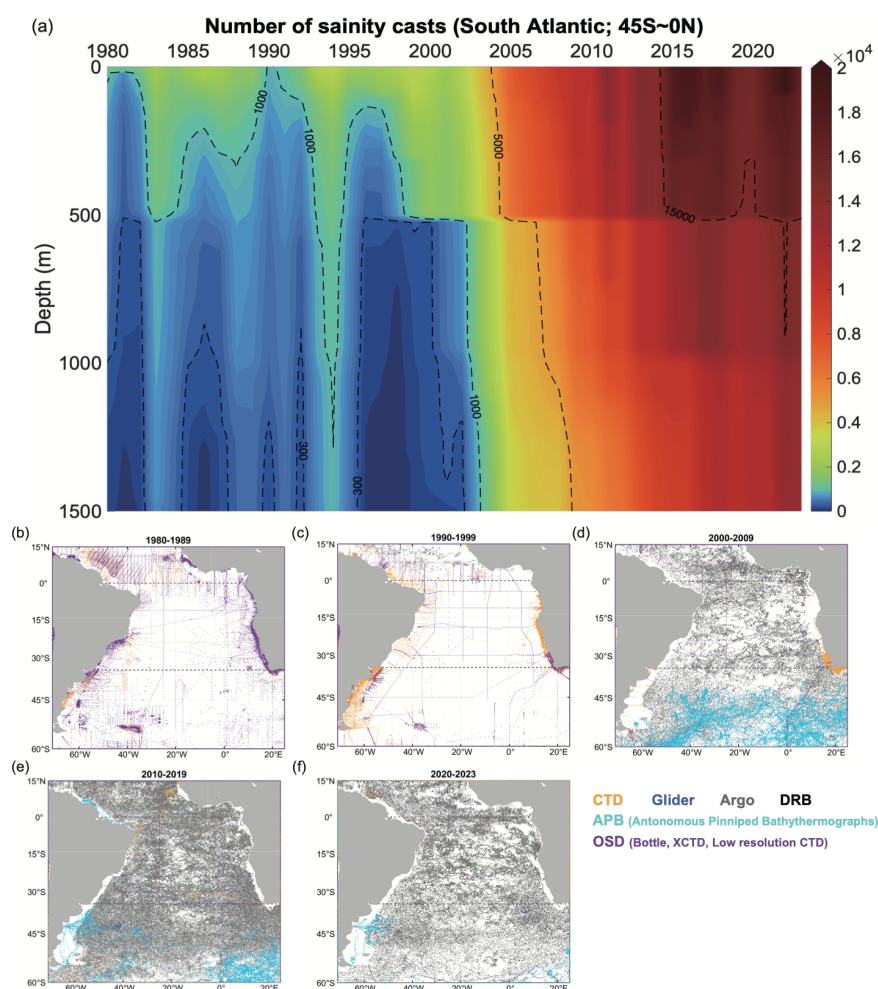


Figure D2: The evolution of the South Atlantic observation capacity. (a) Yearly number of salinity observations at depth levels (0-1500m) in the South Atlantic. (b-e) Geographical distribution of all salinity profiles from different instruments 1980 for decadal periods in the South Atlantic: CTD (conductivity–temperature–depth), Glider, Argo, drifting buoys (DRB), APB (animal mounted pinniped-borne), OSD (Ocean Station Data).

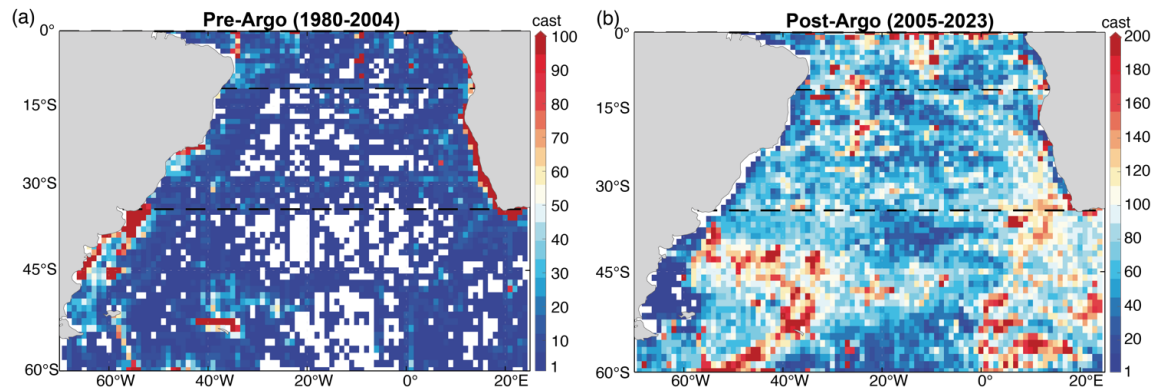


Figure D3: Observational density in the South Atlantic (Number of subsurface salinity casts) at 1-degree box. (a) pre-Argo era (1980-2004); (b) post-Argo era (2005-onward). Note that the scale of the colorbar is not the same in these two panels.

Surface Water (SW) M-SSA analysis
Reconstructed Component 1 (RC-1): Long-term trend of temperature and salinity
19.83% (temperature) and 20.49% (salinity) of explained variance

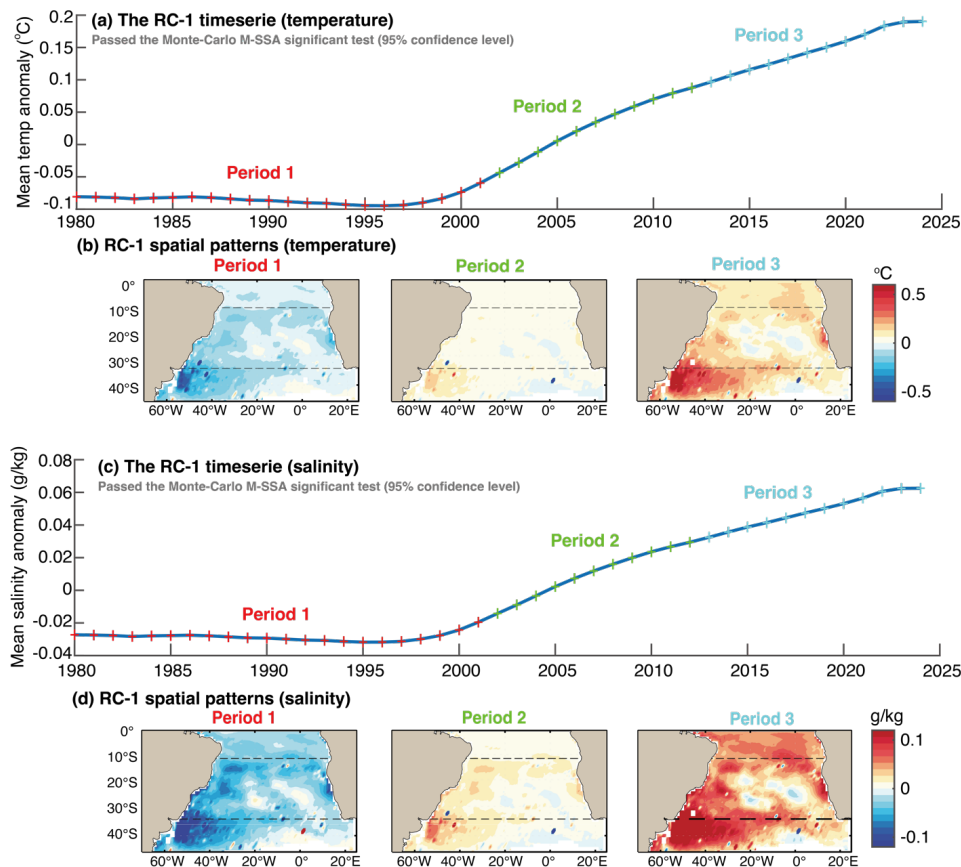


Figure D4: The multichannel singular spectrum analysis (M-SSA) reconstructed long-term trend captured by reconstructed component 1 (RC1; trend mode). (a) Spatially averaged surface water (SW) temperature anomalies for the domain of study. The line with colored plus signs shows the three periods of the mode (composites of the anomalies of each period), which are plotted in (b), which show in contours the mean temperature anomaly field associated with the long-term trend. (c) and (d) are the same as (a) and (b), but for the surface water (SW) salinity trend. Note that all the RC1 long-term trend is passed the Monte-Carlo significance analysis at 95% confidence level.

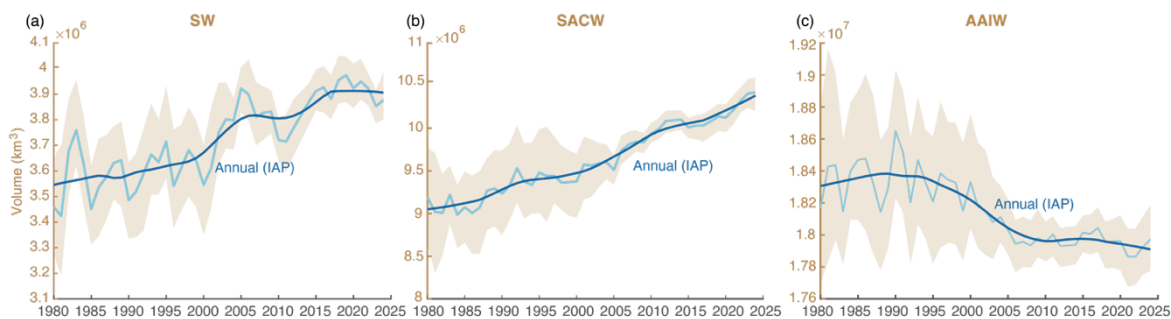


Figure D5: The monthly temporal evolution of water masses thickness in the South Atlantic (regional mean; 45°S–0°N) based on the IAP (Institute of Atmospheric Physics, CAS) monthly gridded products. (a) Surface Water (SW), (b) South Atlantic Central Water (SACW), and (c) Antarctic Intermediate Water (AAIW). Shading represents the 95% uncertainty range of the IAP ensemble estimates, incorporating instrumental, sampling, and mapping uncertainties. Dark blue curves show the 15-year LOWESS-smoothed version.

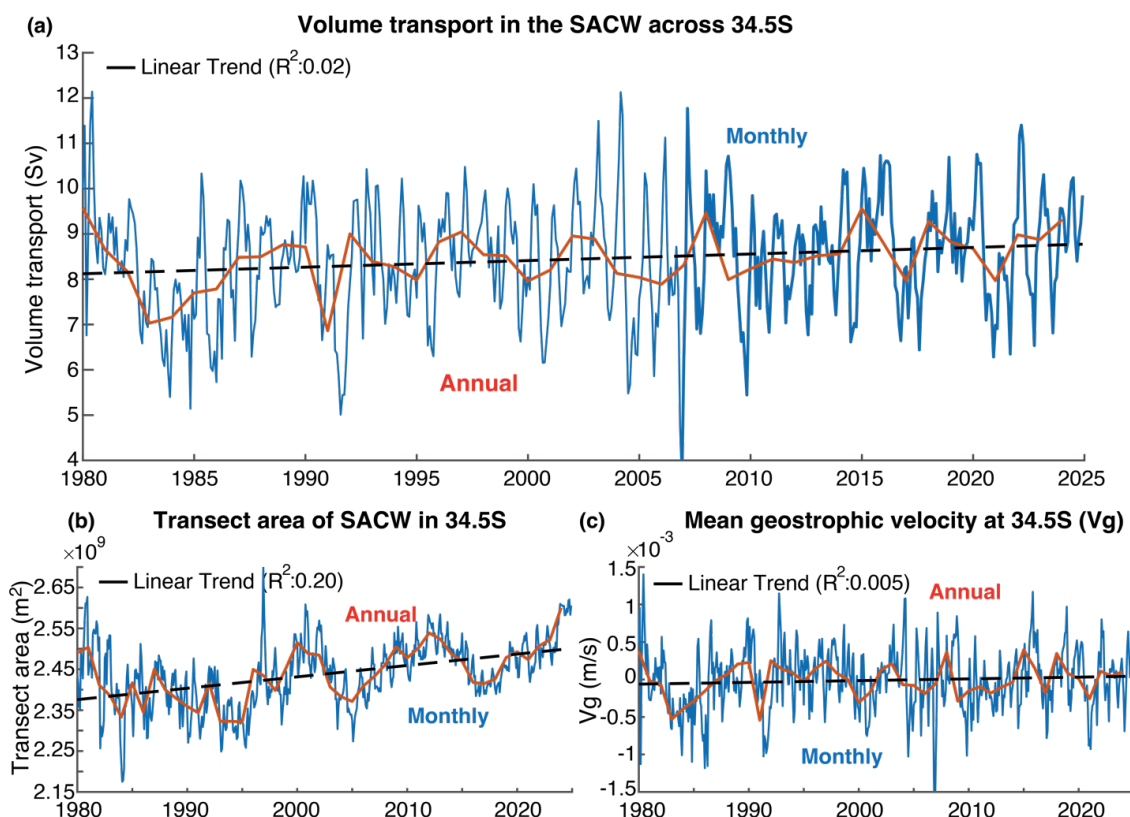


Figure D6: Volume transport and its associated metrics in the SACW at 34.5S transect. (a) Mean volume transport (units: Sverdrup; 1 Sv = 10^6 m³/s) of SACW at 34.5S transect as a function of year based on geostrophic velocity. (b) is the transect area of SACW in 34.5S transect; (c) is the meridional geostrophic velocity (V_g) estimating by temperature and salinity with referenced to the observed mean velocity at 1000 dbar, estimated from Argo float parking pressure. Incorporating time-varying deep reference velocities from Argo trajectory products could be a key priority for future research.

Data availability

The data and materials underlying this article have been organized, compiled, and deposited in Zenodo (<https://zenodo.org/records/20268065>). The availability of the following products used in this study is also appreciated: (i) the IAPv4.2 temperature gridded product is available at http://www.ocean.iap.ac.cn/ftp/cheng/IAPv4_IAP_Temperature_gridded_1month_netcdf/. (ii) the IAPv2.2 salinity gridded product is available at http://www.ocean.iap.ac.cn/ftp/cheng/IAP_Salinity_v2.2_gridded_1month_netcdf/. (iii) the CODCv2.1 temperature and salinity *in-situ* profiles are available at http://www.ocean.iap.ac.cn/ftp/cheng/CODCv2.1_Insitu_T_S_database/ (iv) the MSM60 hydrographic data can be downloaded at <https://doi.org/10.1594/PANGAEA.915898> (v) the GO-SHIP Easy Ocean product is available at



805 <https://doi.org/10.7942/GOSHIP-EasyOcean>. (vi) the "ANDRO argo floats displacements product is made freely available by SNO Argo France at LOPS Laboratory (supported by UBO/CNRS/Ifremer/IRD) and IUEM Observatory (OSU IUEM/CNRS/INSU), and was funded by Ifremer, Coriolis, SOERE-CTDO2 and SNO Argo France at doi: <https://doi.org/10.17882/47077>. The official Matlab codes of the neutral density can be found at https://www.teos-10.org/preteos10/software/neutral_density.html.

810 Author contributions

Z.T., S.S., and E.M. conceptualize and design the work; Z.T. analyzes and visualizes the data; Z.T. drafts the first manuscript. Z.T. A.Z.R., G.M., and S.S. contributed to software development. All authors contribute to the methodology, interpretation of results, and review of the manuscript.

Competing interests

815 The authors declare no competing interests.

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