



Stratification control regimes in observations and their representation in CMIP6 models

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Abstract. Temperature and salinity can reinforce or compensate each other in setting ocean stratification, with important consequences for winter mixing and dense-water formation. Here, we use the Stratification Control Index (SCI), which distinguishes temperature-stabilised alpha regimes ($SCI > 1$), salinity-stabilised beta regimes ($SCI < -1$), and transition regimes ($-1 < SCI < 1$), to diagnose the thermohaline structure below the mixed layer. We construct SCI and mixed-layer depth (MLD) climatologies for 2004–2024 based on the EN.4.2.2 observation dataset and compare them with 26 CMIP6 historical simulations processed with the same definitions. In EN.4.2.2, winter SCI separates subtropical alpha oceans from polar beta oceans, with deep mixed layers occurring mainly for $SCI > 1$. Summer SCI is smoother and shifted toward more alpha-like values because of the seasonal thermocline. CMIP6 shows much larger discrepancies in winter than in summer, with a systematic shift toward stronger haline control at high latitudes. In the North Atlantic north of 40°N , the winter CMIP6 mean product has $SCI = -1.11$ and a beta-regime area of $8.6 \times 10^6 \text{ km}^2$, compared with $SCI = 0.40$ and $4.3 \times 10^6 \text{ km}^2$ in EN.4.2.2. Across the models with available Atlantic Meridional Overturning Circulation (AMOC) estimates, North Atlantic SCI and beta-regime area covary with overturning strength. These systematic winter biases indicate that many CMIP6 models misrepresent the thermohaline preconditioning of regions central to dense-water formation. Since this preconditioning shapes the response of Atlantic and Southern Ocean overturning to warming and freshwater forcing, such mean-state errors limit confidence in projections of future overturning changes.

1 Introduction

Ocean stratification constrains vertical motions and turbulent diapycnal exchange, thereby regulating the vertical transport of heat, salt, and biogeochemical tracers (Ivey et al., 2008; Maffioli et al., 2016). Stratification is also dynamically coupled to the global overturning circulation, whose structure and variability emerge from interacting controls by surface buoyancy forcing, convection, mechanical mixing, and wind pumping (Roquet et al., 2025). Its climatic importance extends beyond the upper ocean: stronger stratification can reduce the downward transport and storage of anthropogenic carbon (Heinze et al., 2015), while intermodel differences in background stratification contribute to the spread of ocean heat and carbon uptake in Climate Model Intercomparison Project phase 6 (CMIP6) (Liu et al., 2023; Bourgeois et al., 2022). Stratification is set by the vertical density gradient, with temperature and salinity contributions that may reinforce or compensate one another.

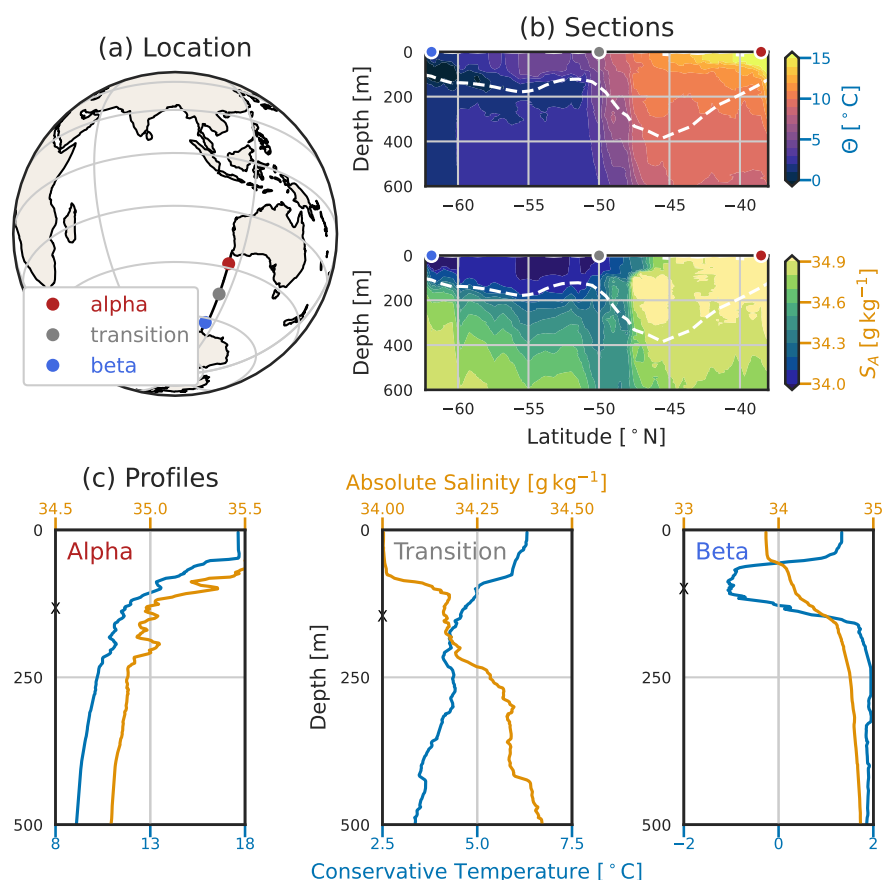


Figure 1. Illustration of alpha, transition, and beta stratification regimes using temperature and salinity observations from the IO9S hydrographic section south of Australia. (a) Location of the section and of the three profiles shown in panel (c). (b) Conservative Temperature and Absolute Salinity along the IO9S section from hydrographic cruise 09AR20120105 (CCHDO Hydrographic Data Office, 2023). The dashed line shows the climatological winter MLD from de Boyer Montégut (2023). (c) Conservative Temperature and Absolute Salinity profiles representative of alpha, transition, and beta regimes. The "X" cross on the y-axes represents the depth of the winter mixed layer. In the alpha regime, temperature is the main stabilising contribution below the mixed layer; in the beta regime, salinity is the main stabilising contribution; in the transition regime, both temperature and salinity contribute to stratification. The selected profiles highlight the meridional change from subtropical alpha stratification to polar beta stratification through an intermediate transition regime.



25 A recent synthesis estimates that stratification in the global upper 2000 m has increased by $0.8 \pm 0.1\%$ per decade since the 1960s and projects continued strengthening under future warming (Cheng et al., 2025). An independent centre-of-mass diagnostic applied to the ECCO ocean state estimate yields a similar global increase in stratification of about 1 % per decade over 1992–2017, driven primarily by upper-ocean warming (Roquet et al., 2026). The response is nevertheless strongly heterogeneous in space, with a growing contribution from salinity at high latitudes and a markedly different behaviour in the North Atlantic, where stratification shows little overall strengthening and slight weakening in some regions. These measures quantify how strong the density stratification is, but focus less on whether temperature or salinity controls it. That distinction is especially important near alpha–beta transition zones, where comparatively small changes in either contribution can shift the response to winter buoyancy loss and alter where mode, intermediate, deep, and bottom waters form. Understanding climatic changes in stratification therefore also requires diagnosing their thermohaline control.

35 A hydrographic section south of Australia illustrates the contrast between thermally and haline controlled stratification (Fig. 1). In subtropical waters, temperature and salinity generally both decrease with depth below the mixed layer, so temperature stabilises the water column while salinity partially compensates this effect. In polar waters, both increase with depth, and stratification is instead maintained by salinity against a destabilising temperature gradient. These regimes are referred to as alpha and beta oceans, respectively, with transition zones where both temperature and salinity contribute to stratification (Pollard et al., 2002; Carmack, 2007; Caneill et al., 2022). Tropical oceans are also often transition-like (Clément et al., 2020). The transition from subtropical alpha to polar beta regimes occurs in both hemispheres, but its latitude, width, and regional structure vary between ocean basins (You, 2002; Stewart and Haine, 2016).

The alpha–beta distinction reveals fundamental differences in stratification. The halocline present in a beta ocean favours sea-ice formation and isolates a subsurface heat reservoir that can limit ice growth, promote melt, and contribute to deep-convection events (Carmack, 2007; Martinson et al., 1981; Pierce et al., 1995; Martin et al., 2013; Polyakov et al., 2013). Reduced sea ice under warmer conditions lowers ocean albedo, increasing heat absorption and further amplifying ocean warming (Curry et al., 1995; Perovich and Polashenski, 2012). Winter mixed layers in salinity-stabilised regions are generally shallower than in temperature-stabilised ones, while double-diffusive effects arise in alpha and beta oceans but not in transition zones where both temperature and salinity stabilise the water column (Stern, 1960; Schmitt, 1994; You, 2002).

50 The existence of a polar transition zone (PTZ) in all subpolar oceans points to a role for the nonlinear properties of seawater. The thermal expansion coefficient becomes small in cold water, making the density gradient comparatively insensitive to temperature while freshwater input can maintain a polar halocline (Roquet et al., 2022; Caneill et al., 2022). If thermal expansion remained fixed at its global-ocean mean value, winter heat loss would erode this halocline much more readily (Caneill et al., 2024). The characteristics of the PTZ vary, however, between the different basins. In the North Pacific and North Atlantic Oceans, the PTZ is associated with a surface density maximum (Roden, 1970; Stewart and Haine, 2016; Johnson et al., 2012; Swift, 1986). This density maximum is absent from the Southern Ocean (SO) PTZ, where it is instead located close to Antarctica in a beta ocean. In the Southern Ocean, the transition from subtropical alpha stratification to polar beta stratification is closely related to the large-scale frontal structure of the Antarctic Circumpolar Current (ACC). Although an exact correspondence is not expected because dynamical fronts are filamentary and time-dependent (Chapman et al., 2020), major



transitions in stratification control broadly coincide with classical hydrographic fronts such as the Subantarctic Front and Polar Front (Pollard et al., 2002; Pauthenet et al., 2017).

The control of stratification by temperature or salinity below the mixed layer varies with season and location (Helber et al., 2012; DuVivier et al., 2018). Stratification below the winter mixed layer characterises the permanent pycnocline that links the surface ocean to the interior. In the SO, for example, subsurface salinity maxima are misrepresented in Earth System Models and are associated with MLD biases (DuVivier et al., 2018). The thermal and haline contributions to stratification therefore provide information on the processes controlling the exchange of heat, salt, and carbon between the surface and the ocean interior (Heinze et al., 2015; Liu et al., 2023). In climate models, such biases are particularly important in the subpolar North Atlantic and Southern Ocean, where stratification, MLD, sea ice, and overturning circulation interact. Temperature–salinity mean-state errors in the subpolar North Atlantic affect the simulated response and variability of the AMOC to atmospheric forcing (Reintges et al., 2024; Ferster et al., 2025), and CMIP model spread in Atlantic freshwater budgets and overturning strength has long been recognised (e.g., Deshayes et al., 2015; Weijer et al., 2020; Gong et al., 2022). More broadly, CMIP6 spread in ocean heat uptake has been linked to base-state stratification, especially in the Southern Ocean and North Atlantic (Liu et al., 2023; Newsom et al., 2023; Vogt et al., 2025).

This motivates the focus on stratification control not only for the observed ocean state, but also for climate models. The MLD is commonly used to evaluate climate models, yet MLD alone does not indicate whether a model mixed layer is well represented because the permanent pycnocline could be controlled by the wrong balance of temperature and salinity. A model may therefore produce a plausible MLD for the wrong thermohaline reason, or a biased MLD because alpha, beta, and transition regimes are not located at a realistic place. We therefore use the SCI to make this distinction explicit. Based on observational temperature–salinity profiles, we first produce seasonal EN.4.2.2-based climatologies of MLD and SCI. We then apply the same diagnostics to CMIP6 historical simulations. The model comparison is designed to test whether biases in the thermohaline control of stratification are associated with misplaced regions of deep winter mixing and with differences in overturning strength.

The paper is organised as follows. Section 2 describes the EN.4.2.2 database, the selected CMIP6 simulations, and the methods used to compute MLD, SCI, and the model AMOC index. Section 3 presents the observational climatology, the SCI–MLD relationship, and the corresponding CMIP6 model biases. Section 4 discusses their implications for dense-water formation and overturning, and Section 5 gives the conclusions and priorities for further work.

2 Material and Methods

This section describes how MLD and SCI maps are computed from EN.4.2.2 profiles and CMIP6 model output. The diagnostic definitions are kept identical for observations and models.



2.1 Definition of the SCI

90 The SCI provides information on the relative effect of temperature and salinity on stratification (Caneill et al., 2022; Roquet et al., 2022),

$$\text{SCI} = \frac{N_{\Theta}^2 - N_S^2}{N_{\Theta}^2 + N_S^2} \quad (1)$$

where temperature and salinity contributions to the squared buoyancy frequency are respectively defined as

$$N_{\Theta}^2 = g\alpha \frac{\partial \Theta}{\partial z} \quad (2)$$

$$95 \quad N_S^2 = -g\beta \frac{\partial S_A}{\partial z}, \quad (3)$$

with α and β the thermal expansion and haline contraction coefficients, respectively. The SCI quantifies the relative effect of temperature and salinity on stratification, and is defined only for positive squared buoyancy frequencies $N^2 = N_{\Theta}^2 + N_S^2$.

Based on the SCI, 3 regimes are defined:

1. $1 < \text{SCI}$: alpha ocean where temperature increases and salinity decreases stratification;
- 100 2. $-1 < \text{SCI} < 1$: transition zone where both temperature and salinity increase stratification;
3. $\text{SCI} < -1$: beta ocean where temperature decreases and salinity increases stratification.

The SCI is a monotonic reparameterisation of the vertical Turner angle (Tu), introduced as a measure of the relative thermal and haline contributions to a water column's stability (Ruddick, 1983; You, 2002). With the sign conventions above and on the statically stable branch, $\text{Tu} = \arctan(\text{SCI})$. The regime boundaries $\text{SCI} = \pm 1$ therefore correspond to $\text{Tu} = \pm 45^\circ$.

105 2.2 EN4-based climatology

2.2.1 EN4 database product

The observational profiles are taken from the EN.4.2.2 database (hereafter EN4) (Good et al., 2013), using the Gouretski and Cheng (2020) MBT correction and the Gouretski and Reseghetti (2010) XBT correction. We use the period 2004–2024. The year 2004 marks the beginning of the Argo era in the Southern Ocean and provides a practical lower bound for constructing a
 110 more spatially homogeneous global product.

Data from EN4 are filtered to remove profiles flagged as containing errors. Each profile is individually smoothed and resampled vertically following an algorithm similar to Johnson et al. (2002):

1. interpolate linearly along an evenly spaced vertical grid with a resolution of 1 m, from 0 to 2000 m;
2. smooth using a Hanning window with a window length of 20 m;
- 115 3. interpolate linearly to a lower-resolution vertical grid: every 5 m from 0 to 100 m, every 10 m from 100 to 350 m, every 25 m from 350 to 500 m, and every 50 m from 500 to 2000 m.



2.2.2 MLD and SCI

For each profile, MLD is computed using a density threshold criterion of 0.03 kg m^{-3} relative to the near-surface reference level (de Boyer Montégut, 2004; Sallée et al., 2021). The same threshold is used for the CMIP6 models, so that observed and simulated SCI–MLD relationships are compared with a common mixed-layer definition.

The SCI is computed over the layer extending from 10 to 30 m below the MLD. This interval samples the stratification immediately beneath the mixed layer while avoiding the sharpest gradients at the mixed-layer base. The SCI definition and the alpha, transition, and beta regime thresholds are given in Section 2.1. For the summary metrics, the transition regime is split into a beta-transition side ($-1 < \text{SCI} < 0$) and an alpha-transition side ($0 < \text{SCI} < 1$).

Throughout the paper, “winter” and “summer” refer to the months of deepest and shallowest mixed layers, respectively, rather than to fixed calendar seasons. At each grid point, the winter SCI is defined as the SCI value for the month in which the climatological MLD reaches its annual maximum, and the summer SCI as the SCI value for the month in which the climatological MLD reaches its annual minimum. This definition accounts for the hemispheric reversal of the seasonal cycle and for regional differences in the timing of mixed-layer deepening and restratification. It also ensures that the winter SCI describes the stratification exposed during maximum mixed-layer ventilation, while the summer SCI describes the stratification below the seasonally shallow mixed layer. In the Arctic, defined here as latitudes north of 66°N , an additional winter-coverage criterion is applied. The annual deepest-MLD result is retained only if MLD and SCI are jointly available in at least one of December, January, or February (DJF). This prevents a non-winter observation from being presented as the winter state where the EN4 climatology has no valid DJF coverage, without changing the annual-extremum definition elsewhere.

2.2.3 Interpolation on a regular grid

MLD and SCI values computed from individual EN4 profiles are interpolated onto a regular 1° longitude–latitude grid using radial basis function interpolation. The interpolation accounts for the meridional change in horizontal length scales through a Mercator transformation of latitude and includes a seasonal time coordinate when constructing monthly climatologies. The decorrelation scales are $L_\lambda = L_{\varphi'} = 3.3$ in longitude and Mercator-projected latitude coordinates, and $L_t = 45 \text{ d}$ in time, following scales similar to those used in the MIMOC climatology (Schmidt et al., 2013).

For each target grid point, the interpolation uses the 300 closest profiles. Grid points with fewer than 50 profiles within the local decorrelation neighbourhood are masked. This criterion limits extrapolation in poorly sampled regions, especially in seasonally ice-covered areas. Further details of the interpolation procedure are given in Appendix A.

Sampling remains uneven in some regions and seasons. The Arctic DJF coverage mask removes cells for which no winter constraint is available, but retained values in sparsely sampled, seasonally ice-covered regions should still be interpreted cautiously. Antarctic continental shelves may likewise be affected by low profile availability during peak winter months. These regions are nevertheless expected to be predominantly beta regimes because salinity controls the polar halocline. Local exceptions may occur where winter convection reaches the seafloor, as can happen in regions of dense shelf water formation.



2.3 CMIP6 model simulations

We compute summer and winter MLD and SCI for the 26 retained CMIP6 historical simulations listed in Table 1. For each model, the MLD is recomputed from monthly temperature and salinity profiles using the same 0.03 kg m^{-3} density threshold as for EN4. The SCI is then computed 10–30 m below the recomputed mixed layer.

The primary model comparison uses the historical period overlapping the EN4 product, ending in 2014 for CMIP6 historical simulations. Model diagnostics are computed on the native model grid where possible and then regridded to a common 1° grid for comparison with EN4.

Paired summer and winter SCI and MLD maps for every retained CMIP6 model are provided in the Supplement (Figs. S1–S26). Seasonal multi-model mean fields are calculated grid point by grid point after regridding. Figure 6 alone uses pointwise ensemble percentiles: at each latitude it shows the median and 20th–80th percentile range of the individual model zonal means.

Table 1 reports area-weighted SCI, MLD, AMOC strength, and the areas of alpha (A_α) and beta (A_β) regimes for each model and for the CMIP6 multi-model mean. Before calculating these metrics, every model field is restricted to the common grid-cell domain where both EN4 winter SCI and MLD are valid.

For the subset of 18 models with the required overturning output, we diagnose the strength of the upper AMOC cell from the Atlantic meridional overturning streamfunction in depth coordinates. This streamfunction is the zonally integrated and vertically accumulated meridional volume transport. At the model grid latitude nearest 26°N , we take its maximum below 500 m for each month and average the monthly index over 2004–2014. Restricting the maximum to depths below 500 m reduces the influence of shallow wind-driven circulation on the basin-scale upper-cell index. Transport is reported in Sverdrups ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$), and the CMIP6 mean is the arithmetic mean of the 18 available model indices.

The RAPID array supplies a closely related observational benchmark at 26.5°N . RAPID reconstructs a full-depth overturning streamfunction by combining the geostrophic interior transport inferred from boundary density profiles with directly observed boundary and Florida Current transports and wind-derived Ekman transport, subject to basin-wide mass balance (Frajka-Williams et al., 2019). Its April 2004–February 2017 mean is 17.0 Sv . The model and RAPID indices therefore represent the same depth-coordinate concept at nearly the same latitude and in the same units. The RAPID value is consequently used as an observational reference in Fig. 8 and Table 1.

Both seasons are used to assess the seasonal structure of model biases. The interpretation of ventilation and overturning focuses on winter SCI, because the deepest seasonal mixed layers most directly expose the permanent pycnocline and ventilate the ocean interior.

3 Results

3.1 Seasonal structure of the EN4 SCI climatology

The EN4 SCI climatology shows a marked seasonal contrast in the thermohaline control of stratification below the mixed layer (Fig. 2). In winter, the SCI separates the global ocean into subtropical alpha regimes, subpolar and polar beta regimes, and



narrow transition zones between them. This large-scale organisation is consistent with the alpha–beta structure described from ECCO by Roquet et al. (2022, Fig. 5a), although the EN4 field reveals a sharper regional structure. The smoother ECCO field is expected because ECCO is a coarse-resolution dynamically consistent state estimate.

The winter SCI shows large and sharp spatial variations that closely follow major ocean-circulation structures. The white contours in Fig. 2 show selected contours of the CNES-CLS22 combined mean dynamic topography (CMDT), an estimate of the 1993–2012 mean sea surface height above the geoid (CNES-CLS, 2023; Jousset et al., 2022). The transition from alpha to beta stratification is often localized near frontal regions, including western boundary-current extensions and the Antarctic Circumpolar Current. This correspondence is not exact, because the SCI diagnoses the thermohaline control below the mixed layer while CMDT contours represent a surface geostrophic circulation structure. The comparison nevertheless shows that the winter SCI is closely tied to the permanent pycnocline and to the large-scale pathways of the ocean interior.

The summer SCI field is smoother and less contrasted than the winter field. Most regions shift toward larger SCI values, and fully beta regimes become much less widespread. Surface warming strengthens the stabilising thermal contribution and builds a shallow seasonal thermocline. Because the mixed layer is shallow in summer, the SCI diagnosed immediately below it samples upper-ocean gradients that are strongly shaped by seasonal surface buoyancy forcing and restratification (Helber et al., 2012; DuVivier et al., 2018; Sallée et al., 2021). It therefore reflects the deeper permanent pycnocline less directly than the winter SCI. The tropics are transition-like in both seasons because temperature and salinity commonly both contribute to upper-ocean stratification there.

The zonal climatology provides a compact view of the same seasonal contrast (Fig. 3). In winter, the subtropical oceans are strongly alpha, while the SCI decreases rapidly toward subpolar and polar latitudes. Both the Atlantic and Pacific exhibit substantial longitudinal variability, reflected in the 10th–90th percentile range around their zonal means. This spread is particularly broad in the North Atlantic north of about 40°N, where the map shows alpha and transition regimes extending poleward in the eastern basin while beta conditions extend farther south in the western subpolar gyre (Fig. 2). The North Pacific also shows a pronounced poleward decrease in SCI, together with regional structure that is not captured by the zonal mean alone. In summer, the meridional gradients weaken and the distribution shifts toward higher SCI values, consistent with the greater influence of the shallow seasonal thermocline.

3.2 Observed relationship between SCI and MLD

The joint distribution of SCI and MLD shows how the seasonal mixed layer samples different stratification regimes (Fig. 4). In summer, mixed layers are shallow over most of the ocean and the SCI distribution is shifted toward more positive values. Most grid cells lie in the transition regime or on its alpha side, consistent with the stabilising role of the seasonal thermocline. A weaker secondary maximum remains near the beta boundary, reflecting high-latitude regions where salinity still contributes strongly to stratification below the shallow summer mixed layer.

The winter distribution is markedly different (Fig. 4b). Mixed layers are systematically deeper, and the SCI distribution is more clearly bimodal, with large area in both alpha and beta regimes. Deep winter mixed layers are mostly found for SCI > 1, with MLD larger than about 200 m concentrated on the alpha side of the distribution. This result is central to the model

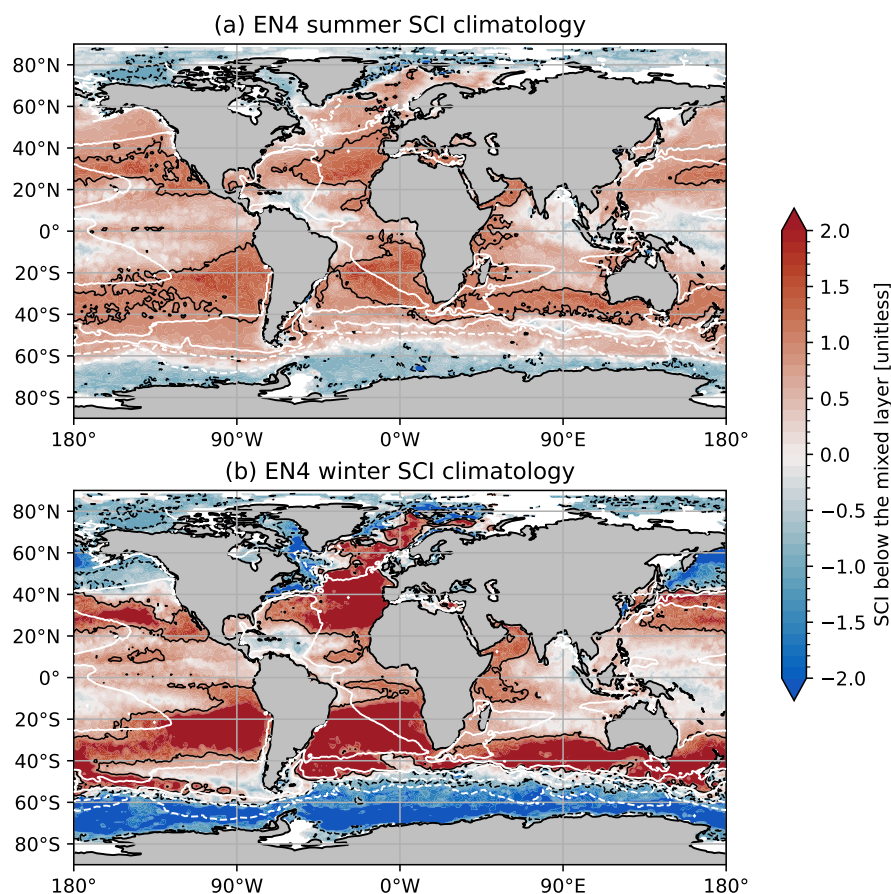


Figure 2. EN4 SCI computed 10–30 m below the mixed layer for 2004–2024, computing for (a) summer SCI, and (b) winter SCI. The black contours mark $SCI = 1$ and $SCI = -1$, separating alpha regimes ($SCI > 1$), transition regimes ($-1 < SCI < 1$), and beta regimes ($SCI < -1$). White contours show selected contours of CNES-CLS22 combined mean dynamic topography (CMDT).

215 comparison: deep winter mixing in EN4 is not simply associated with weak stratification or with polar beta conditions, but preferentially occurs where the permanent pycnocline remains temperature-controlled or only partly compensated by salinity.

3.3 CMIP6 seasonal SCI regimes and biases

We next compare the seasonal EN4 SCI structure with the CMIP6 historical simulations multi-model mean. Figure 5 shows the summer and winter multi-model means and their differences from EN4. The difference panels are defined as the CMIP6 historical mean for 2004–2014 minus the EN4 climatology for 2004–2024 and are restricted to their common valid grid.

220 The summer multi-model mean broadly follows EN4, and its differences are comparatively small and regional. CMIP6 represents the thermohaline control beneath the shallow seasonal mixed layer substantially better than the winter control of the permanent pycnocline. In winter, widespread negative differences emerge poleward of about 40° in both hemispheres (Figs. 5

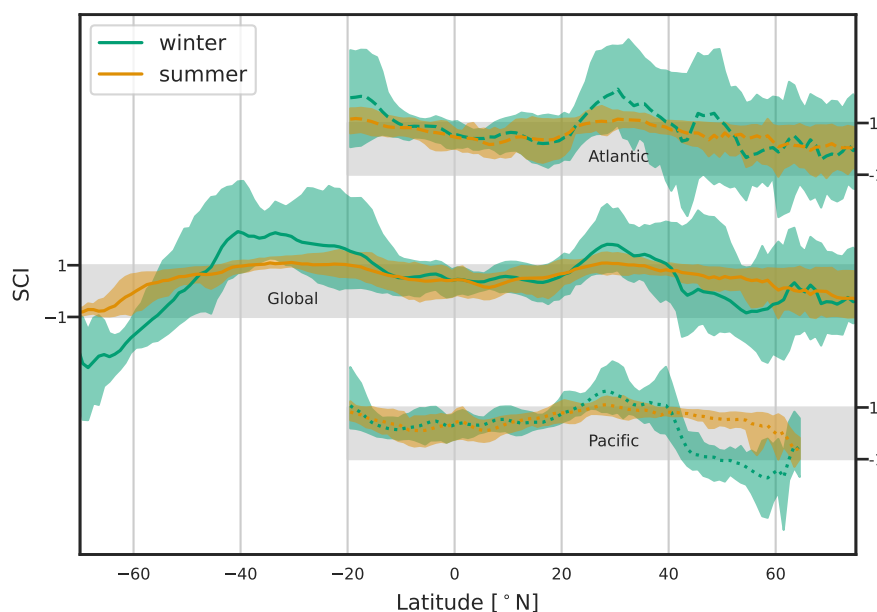


Figure 3. Zonal climatology of SCI below the mixed layer from EN4 for 2004–2024. Green curves show winter SCI and orange curves show summer SCI. Lines are area-weighted zonal means, and shading shows the 10th–90th percentile range across longitude. The middle panel shows the global zonal mean. The upper and lower panels show Atlantic and Pacific basin averages, shifted vertically for clarity. Grey shading marks the transition regime $-1 < \text{SCI} < 1$.

and 6). The CMIP6 mean places $\text{SCI} < -1$ over broader high-latitude regions than EN4, indicating excessive haline control
 225 and/or excessively strong subsurface temperature inversions below the winter mixed layer. Positive regional differences remain,
 including parts of the western boundary-current extensions, but they do not offset the broad high-latitude beta bias.

The zonal comparison (Fig. 6) confirms that the winter beta bias is not restricted to isolated regions. Across both hemispheres,
 many CMIP6 models show winter SCI values that are too low poleward of about 40° , with broad excursions into the beta
 regime in the global, Atlantic, and Pacific averages. By contrast, the summer model median follows EN4 closely through most
 230 latitudes and basins, and the intermodel spread narrows markedly. The seasonal thermocline and its thermohaline control are
 therefore substantially better represented than the winter permanent pycnocline. This contrast focuses attention on processes
 that become decisive during winter ventilation, including high-latitude freshwater distribution, sea-ice formation and melt,
 surface buoyancy loss, and vertical mixing.

3.4 SCI–MLD relationship in CMIP6

235 The SCI–MLD distributions provide a direct comparison of the thermohaline regimes associated with seasonal mixing in EN4
 and CMIP6. Figure 7 is constructed from the grid-point-wise CMIP6 multi-model mean summer and winter MLD and SCI
 fields, using the same limits, binning, latitude classes, and area weighting as for EN4 in Fig. 4.

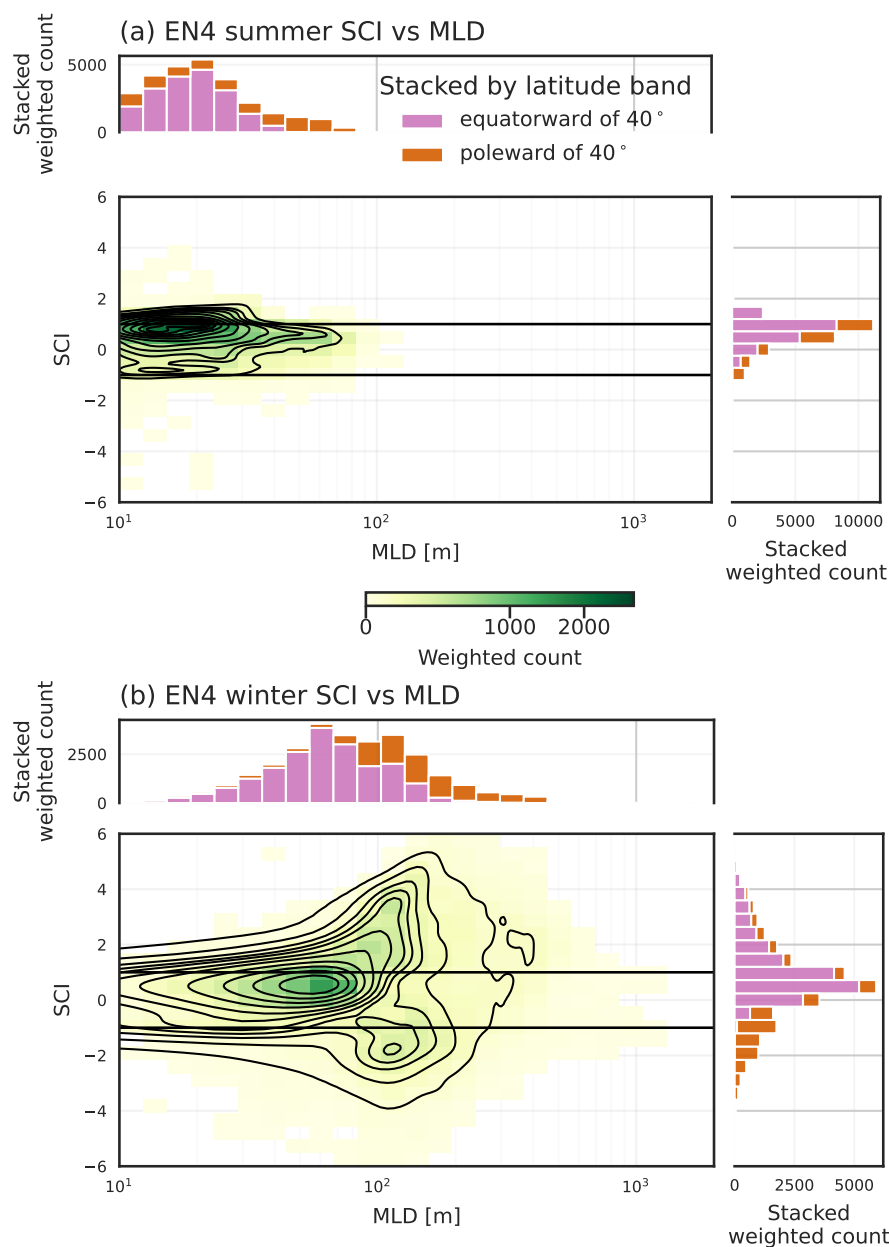


Figure 4. Relationship between SCI and MLD in EN4 for 2004–2024 in (a) summer and (b) winter. The two-dimensional histograms use all ocean grid cells and are area weighted. MLD is shown on a logarithmic axis from 10 to 2000 m. The central two-dimensional histograms combine both latitude classes. The coloured marginal histograms separate grid cells equatorward and poleward of 40° latitude and are stacked: the poleward contribution is added to the equatorward contribution, so the full bar height or width is their combined area-weighted count.

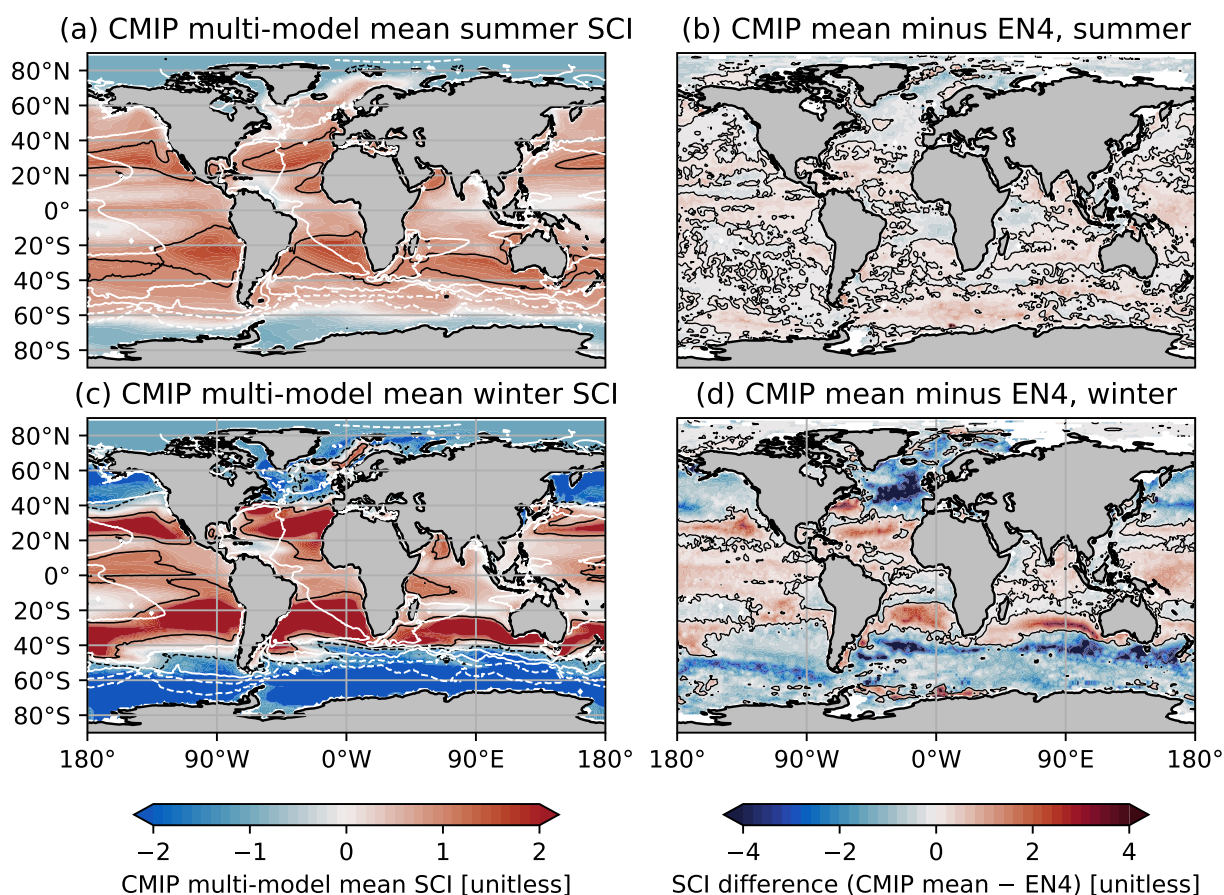


Figure 5. Seasonal SCI in the CMIP6 multi-model mean and its difference from EN4. Panels show (a) the CMIP6 summer mean, (b) the summer difference, (c) the CMIP6 winter mean, and (d) the winter difference. Differences are CMIP6 historical 2004–2014 multi-model mean minus the EN4 2004–2024 climatology on their common valid grid; negative values therefore indicate more haline-controlled stratification in CMIP6. Black contours in panels (a) and (c) mark $SCI = \pm 1$, and black contours in panels (b) and (d) mark zero difference. White contours in the mean panels show selected CNES-CLS22 CMDT contours.

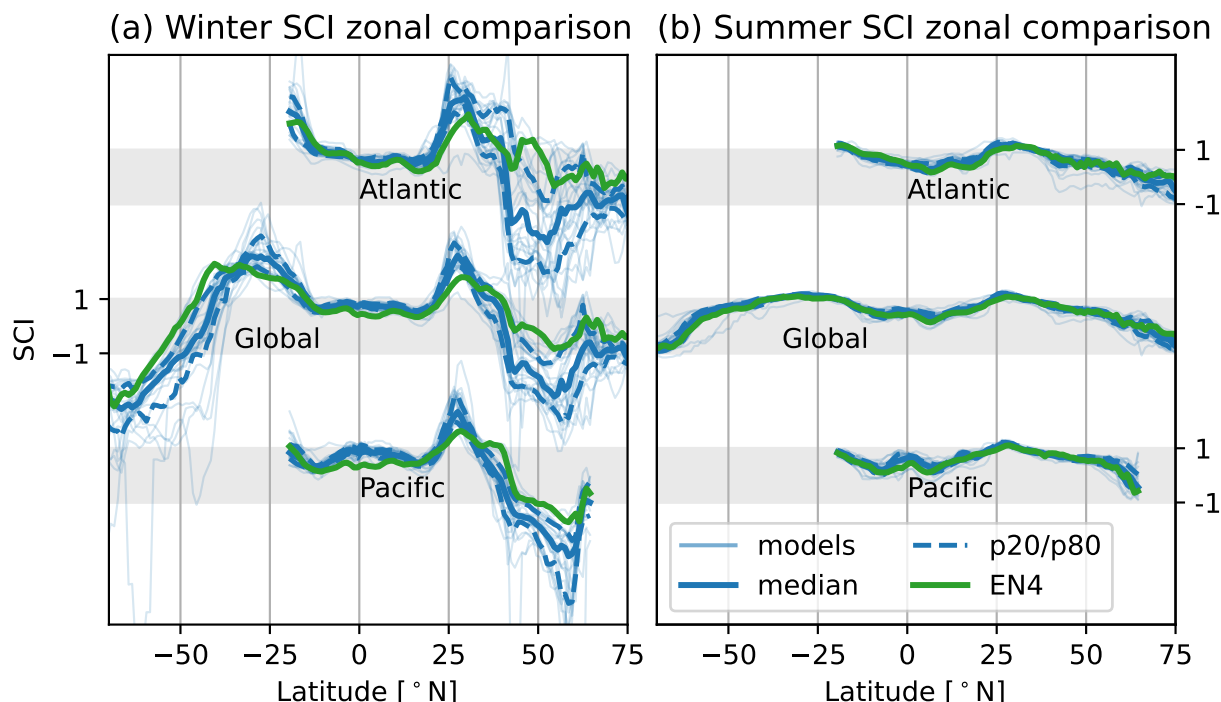


Figure 6. Zonal comparison of SCI below the mixed layer for EN4 and the CMIP6 historical simulations in (a) winter and (b) summer. Each panel follows Fig. 3, with global, Atlantic, and Pacific averages shifted vertically for clarity. Thin blue lines show individual CMIP6 models, solid blue lines the model median, dashed blue lines the 20th–80th percentile range, and green lines the EN4 climatology. Grey shading marks the transition regime $-1 < \text{SCI} < 1$ for each shifted curve.

In summer, the CMIP6 distribution closely resembles EN4: most of the ocean is associated with shallow mixed layers and transition-to-alpha SCI values. This similarity is consistent with the relatively small differences between EN4 and CMIP6 in the summer SCI climatology.

The winter distributions differ more substantially. In EN4, the SCI distribution is bimodal and deep mixed layers occur predominantly on the alpha side, with most MLDs deeper than about 200 m associated with $\text{SCI} > 1$ (Fig. 4). In the CMIP6 multi-model mean, the distribution remains bimodal, but a much larger fraction of the winter mixed-layer distribution lies in the beta regime. Deep mixed layers are also shifted toward lower SCI, with a pronounced concentration close to the $\text{SCI} = -1$ boundary rather than within the alpha regime.

This shift in SCI–MLD space complements the geographical biases shown in Figs. 5 and 6. Those figures show that the excessive beta conditions occur predominantly at high latitudes, whereas Fig. 7 shows that these stratification biases are also associated with a different thermohaline environment for winter mixing. SCI therefore provides information that is not contained in MLD alone, by identifying whether a given mixed-layer depth occurs under realistic thermal and haline control of the underlying stratification.

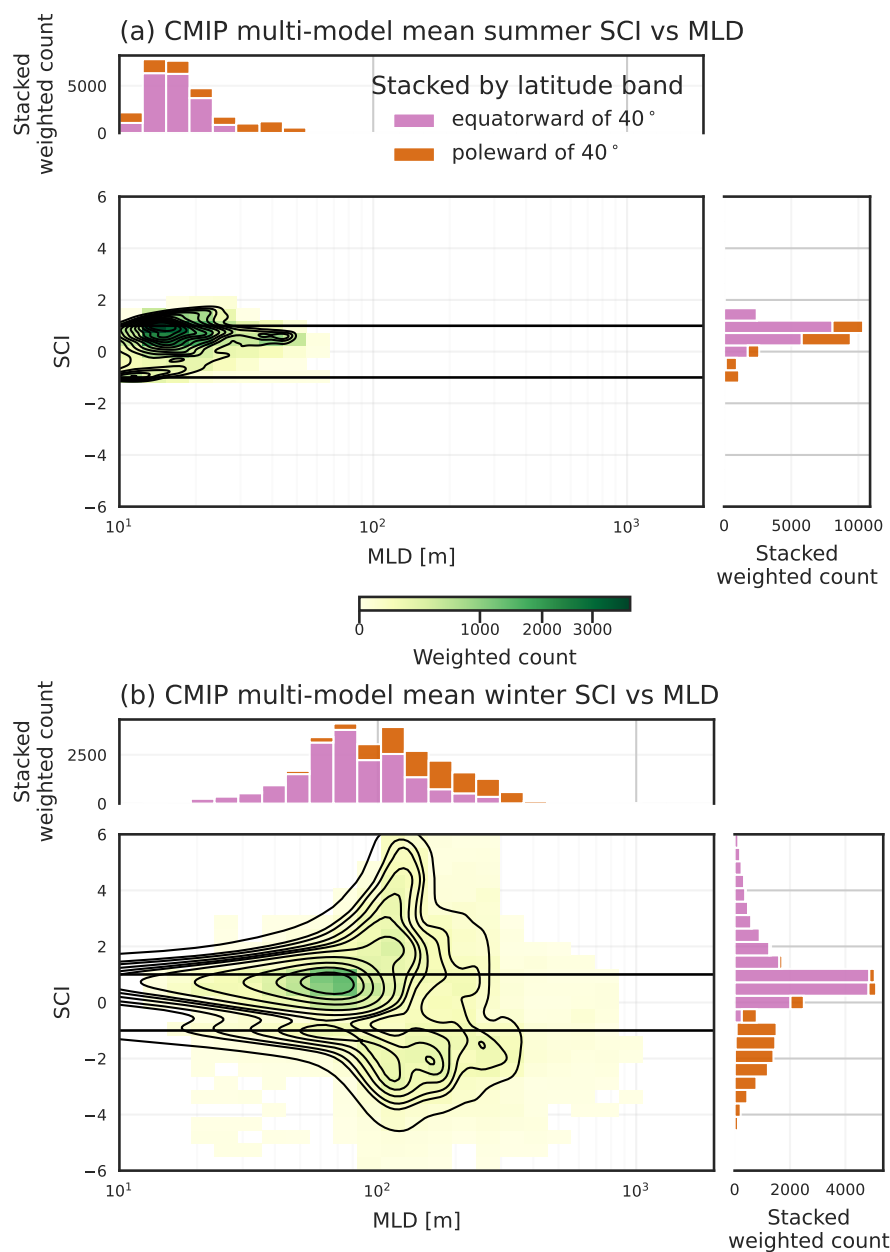


Figure 7. Relationship between SCI and MLD in the CMIP6 multi-model mean for (a) summer and (b) winter. MLD and SCI are first averaged grid point by grid point across the 26 retained models, after which the area-weighted joint distributions are computed using the same limits and binning as in Fig. 4. The central two-dimensional histogram combines all latitudes. The marginal histograms separate regions equatorward and poleward of 40° , with the two contributions stacked.



3.5 Summary metrics for model evaluation

The model-by-model metrics listed in Tab. 1 quantify the SCI biases seen in the maps and histograms. All model metrics are evaluated on the common EN4-valid ocean domain. For the global ocean, EN4 has a mean winter SCI of 0.64, a mean winter MLD of 97 m, an alpha-regime area of $124 \times 10^6 \text{ km}^2$, and a beta-regime area of $46 \times 10^6 \text{ km}^2$. The CMIP6 mean product is shifted toward lower SCI, with a global mean SCI of 0.35, a mean MLD of 119 m, an alpha area of $105 \times 10^6 \text{ km}^2$, and a beta area of $81 \times 10^6 \text{ km}^2$. Thus, the CMIP6 mean product increases the global beta area by about 76 % relative to EN4.

The bias is stronger in the North Atlantic north of 40°N . There, EN4 has a regional mean SCI of 0.40, a mean winter MLD of 186 m, an alpha area of $6.5 \times 10^6 \text{ km}^2$, and a beta area of $4.3 \times 10^6 \text{ km}^2$. The CMIP6 mean product instead has a mean SCI of -1.11 , a mean MLD of 275 m, an alpha area of only $0.68 \times 10^6 \text{ km}^2$, and a beta area of $8.6 \times 10^6 \text{ km}^2$. Individual models span a broad range, but the models closest to the EN4 regime areas lie consistently on the high-SCI side of the ensemble (Table 1).

Figure 8 connects the North Atlantic SCI metrics with MLD and AMOC strength. Across the 26 CMIP6 models, North Atlantic mean SCI is positively correlated with North Atlantic mean winter MLD ($r = 0.81$, $p < 0.001$), indicating that models with stronger thermal control tend to have deeper winter mixed layers in this region. Across the 18 models with overturning output, AMOC strength is positively correlated with North Atlantic SCI ($r = 0.73$, $p < 0.001$) and MLD ($r = 0.80$, $p < 0.001$), and negatively correlated with beta-regime area ($r = -0.80$, $p < 0.001$). The 17.0 Sv RAPID mean reported by Frajka-Williams et al. (2019) is shown with the EN4 stratification metrics as an observational reference, but is not included in the model regressions. The correlations do not establish causality, but they show that SCI captures part of the coupled model spread linking subpolar thermohaline structure, winter mixing, and overturning strength.

These covariations are physically plausible because overturning depends on both buoyancy-driven water-mass transformation and the dynamical export of transformed waters. The SCI does not determine AMOC strength, but it characterises the thermohaline preconditioning of regions where winter mixing ventilates the ocean interior. It is therefore most informative when interpreted together with MLD, density, and circulation metrics rather than as a stand-alone predictor of overturning strength.

4 Discussion

4.1 Seasonal fidelity and physical meaning of the SCI

The EN4 climatology shows that the SCI provides a compact description of the thermohaline structure immediately below the mixed layer. In winter, subtropical oceans are predominantly alpha, subpolar and polar oceans are predominantly beta, and the transition between them is often narrow. The strongest gradients occur near western boundary-current extensions and the ACC. Their broad alignment with CMDT contours is consistent with winter SCI reflecting the structure of the permanent pycnocline and the pathways along which mixed-layer properties enter the ocean interior (Sallée et al., 2010). Exact correspondence is not expected, since SCI describes the vertical thermohaline structure whereas CMDT reflects the surface geostrophic circulation.



Table 1. Quantitative summary of winter SCI and MLD diagnostics for the global ocean and the North Atlantic north of 40°N. Means are area-weighted over the common EN4-valid ocean domain. A_α and A_β denote the area covered by alpha (SCI > 1) and beta (SCI < -1) regimes in units of 10^6 km^2 . Model AMOC values are the 2004–2014 mean maximum of the depth-coordinate Atlantic overturning streamfunction below 500 m near 26°N; the observational row pairs EN4 stratification diagnostics with the 17.0 Sv RAPID mean at 26.5°N (Frajka-Williams et al., 2019).

Model	Global				North Atlantic (> 40°N)				AMOC [Sv]
	SCI	MLD [m]	A_α [10^6 km^2]	A_β [10^6 km^2]	SCI	MLD [m]	A_α [10^6 km^2]	A_β [10^6 km^2]	
Observation (EN4/RAPID)	0.64	97	124	46	0.40	186	6.5	4.3	17.0
(1) GFDL-CM4	0.69	109	133	46	0.65	335	6.8	4.1	17
(2) TaiESM1	0.66	111	115	55	0.63	350	6.8	3.9	–
(3) NorESM2-MM	0.90	147	142	45	0.62	349	5.4	4.8	21
(4) GISS-E2-1-G	0.60	124	116	52	0.52	473	6.8	4.4	–
(5) KIOST-ESM	0.38	142	101	65	0.36	435	6.1	4.9	–
(6) SAM0-UNICON	0.62	113	112	63	0.29	355	5.2	4.8	20
(7) FIO-ESM-2	0.65	114	112	58	0.04	319	4.7	5.6	–
(8) CESM2	0.72	122	119	55	0.00	291	4.6	4.3	18
(9) BCC-ESM1	0.45	154	128	65	-0.17	458	5.2	5.6	–
(10) ACCESS-ESM1-5	0.44	133	124	70	-0.26	309	4.3	5.8	19
(11) MCM-UA-1-0	-0.08	99	128	73	-0.91	361	4.4	5.9	–
(12) CNRM-CM6-1	0.21	105	91	89	-1.33	225	2.0	7.6	17
(13) MRI-ESM2-0	0.60	139	105	73	-1.36	354	2.3	8.4	17
(14) CAMS-CSM1-0	0.11	144	119	83	-1.46	398	2.3	8.7	–
(15) FGOALS-f3-L	0.24	103	91	82	-1.71	216	1.5	9.0	20
(16) NESM3	0.00	103	116	104	-1.75	204	0.78	10	–
(17) EC-Earth3	0.42	111	110	75	-1.85	286	2.2	8.5	18
(18) UKESM1-0-LL	0.38	118	117	79	-2.05	203	1.3	9.7	16
(19) HadGEM3-GC31-LL	0.40	117	115	74	-2.10	197	1.3	10	16
(20) MIROC6	0.34	127	95	83	-2.28	230	1.3	10.0	13
(21) IPSL-CM6A-LR	0.07	110	106	96	-2.28	164	0.90	11	13
(22) CIESM	0.46	107	103	81	-2.31	154	0.61	11	11
(23) CMCC-ESM2	0.47	111	108	69	-2.38	201	1.1	10	14
(24) MPI-ESM1-2-LR	-0.15	119	106	97	-2.47	227	1.6	9.4	17
(25) CanESM5	0.50	116	108	77	-2.51	158	1.1	11	12
(26) E3SM-1-0	-0.58	102	86	100	-2.81	87	0.06	12	10
CMIP6 mean product	0.35	119	105	81	-1.11	275	0.68	8.6	16

SCI is dimensionless. MLD in metres. A_α and A_β in 10^6 km^2 . AMOC in Sverdrups. The model index and its comparison with RAPID are described in Section 2.3. The CMIP6 mean row is the arithmetic multi-model mean; AMOC for that row is the mean across the 18 models with available overturning output.

The comparison with CMIP6 reveals a marked seasonal asymmetry. Models reproduce the smoother, more thermally controlled summer state comparatively well: the model median follows EN4 closely through most latitude bands and the ensemble spread is relatively narrow (Fig. 6b). In winter, the agreement deteriorates strongly at high latitudes in all basins (Fig. 6a). Summer SCI mainly samples the shallow seasonal thermocline produced by surface warming and restratification, whereas winter mixing reaches deeper and is more directly influenced by the thermohaline structure of the permanent pycnocline. The stronger

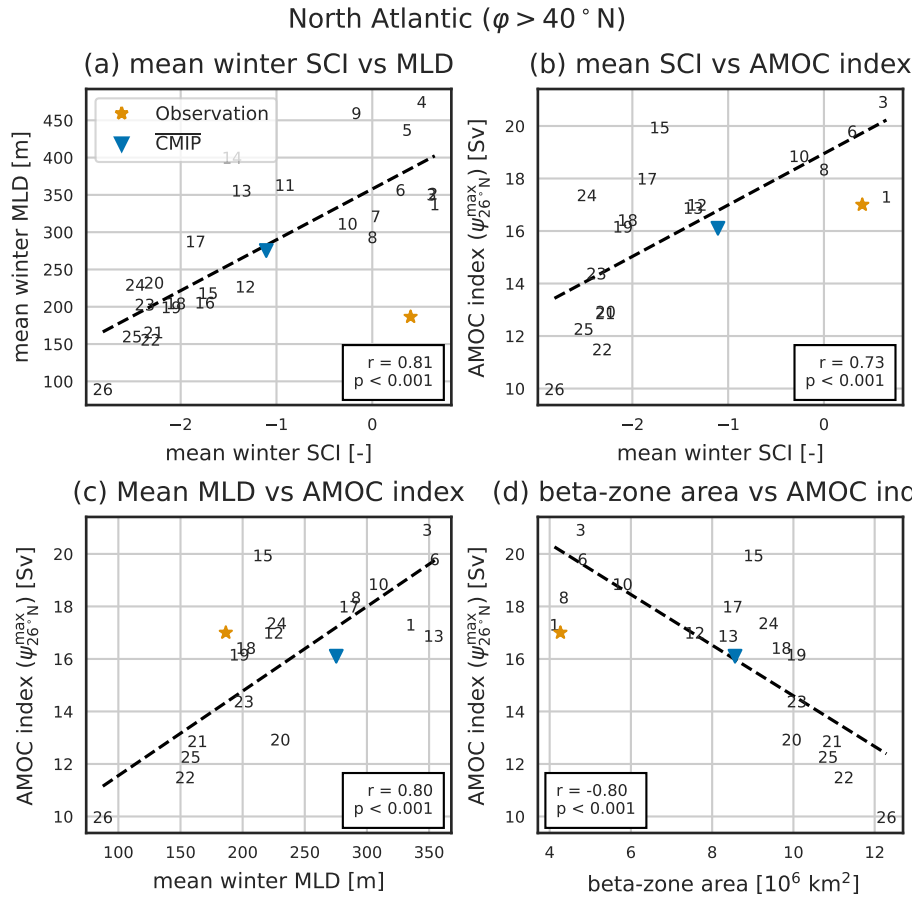


Figure 8. Relationships between winter upper-ocean stratification metrics and Atlantic overturning strength across CMIP6 models. Panel (a) shows the mean winter SCI versus mean winter MLD in the North Atlantic north of 40°N for the EN4 observational estimate, individual CMIP6 models, and the CMIP6 multi-model mean. Panels (b)–(d) show the relationships between $\text{AMOC}_{26^\circ\text{N}}^{\text{max}}$ and the North Atlantic mean winter SCI, mean winter MLD, and beta-regime area A_β north of 40°N , respectively. The model index is the 2004–2014 mean maximum of the depth-coordinate Atlantic overturning streamfunction below 500 m near 26°N . The observational reference combines EN4 stratification diagnostics with the 17.0 Sv RAPID mean at 26.5°N reported by Frajka-Williams et al. (2019). Numbers are the model markers and identify models in the order listed in Tab. 1; the observation and the CMIP6 multi-model mean are labelled explicitly. Dashed lines, Pearson correlation coefficients, and two-sided p values use CMIP6 models only; the observational reference and multi-model mean are excluded from the fits.

winter biases therefore point to errors in the water-mass structure and processes that set high-latitude stratification, rather than simply in the seasonal thermocline.

Some caution is required in interpreting the climatologies. Winter sampling remains sparse in seasonally ice-covered regions, particularly in the central Arctic and on Antarctic continental shelves, and the interpolation cannot fully compensate for this lack of observations. In addition, a single density-threshold definition of MLD is applied globally. Such a criterion is less robust



in weakly stratified or actively convecting regions, where the diagnosed MLD may differ substantially from the depth of active mixing (Våge et al., 2015; Brakstad et al., 2023; Strehl et al., 2024). Finally, because SCI is normalised by total N^2 , values become sensitive to small errors in the thermal and haline contributions when N^2 approaches zero. The diagnostic is therefore most robust where the stratification below the mixed layer is sufficiently strong, while very weakly stratified regions require more cautious interpretation.

4.2 Thermohaline control of mixing and water-mass formation

The EN4 SCI–MLD relationship shows that deep winter mixed layers are not distributed evenly across stratification regimes. Mixed layers deeper than about 200 m occur mainly for $SCI > 1$, whereas fully beta regimes contribute much less to the deepest part of the distribution. A stabilising halocline can therefore inhibit winter mixed-layer deepening even under strong surface heat loss. The SCI adds information that MLD alone lacks by identifying the thermohaline structure into which the mixed layer deepens.

Recent Arctic changes provide a regional analogue. Atlantification in the Barents Sea, Eurasian Basin, and Greenland Sea has been associated with warmer and saltier Atlantic-origin waters, weaker haline stratification, and deeper winter mixing (Årthun et al., 2012; Polyakov et al., 2017; Strehl et al., 2024; Polyakov et al., 2025). Reported shifts toward more thermally controlled stratification in these regions (Hordoir et al., 2022; Gjelstrup and Stedmon, 2024; Ruiz-Angulo et al., 2026) are consistent with the EN4 tendency for deep mixing to occur on the alpha side of SCI space, although total stratification and atmospheric forcing also change.

Mixed-layer depth and density are commonly used to characterise dense-water formation and its contribution to overturning (Heuzé, 2017; Madan et al., 2024). The SCI provides a complementary measure of the thermohaline preconditioning of these formation regions, particularly near transition zones where a modest change in the temperature or salinity contribution can move winter mixing into a different regime. Because the permanent pycnocline can be sampled outside peak winter, stratification control may also provide a useful constraint where direct observations of maximum winter MLD remain sparse.

4.3 Origins of the high-latitude beta bias

The CMIP6 ensemble exhibits a systematic high-latitude shift toward beta stratification. On the common EN4-valid domain, the multi-model mean beta area is $81 \times 10^6 \text{ km}^2$, compared with $46 \times 10^6 \text{ km}^2$ in EN4. North of 40°N in the Atlantic, mean SCI shifts from 0.40 in EN4 to -1.11 in CMIP6, while the beta area nearly doubles. The SCI–MLD distribution also places deep model mixing closer to the $SCI = -1$ boundary than in EN4. These differences indicate excessive haline control, but not necessarily excessive freshwater or stronger total stratification: a low SCI may result from a stronger stabilising salinity gradient, a stronger destabilising temperature inversion, or both.

Several processes may contribute to this bias. High-latitude salinity is controlled by surface freshwater fluxes, river runoff, ice-sheet discharge, sea-ice formation and melt, and ocean freshwater transport. In the North Atlantic, errors in the large-scale circulation provide an additional mechanism. Coarse-resolution climate models commonly separate the Gulf Stream too far north and simulate an overly zonal North Atlantic Current, altering the poleward transport of both heat and salt into the



325 subpolar gyre (Moreno-Chamarro et al., 2022). Such circulation errors can therefore modify both terms entering the SCI, and may contribute to the North Atlantic beta bias.

Biases originating farther north may also be important. CMIP6 models exhibit large errors in Arctic hydrography: the multi-model mean halocline is too fresh (Khosravi et al., 2022), and excessive Arctic liquid freshwater content remains a common model bias (Wang et al., 2022). Errors in Arctic freshwater storage and export could reinforce haline stratification downstream
 330 in the Nordic Seas and subpolar North Atlantic. In the Southern Ocean, excessive surface freshwater input has previously been linked to positive haline-stratification and shallow mixed-layer biases (Sallée et al., 2013), while CMIP6 still misrepresents the location and mechanisms of deep and bottom-water formation (Heuzé, 2021). The present SCI patterns are consistent with contributions from both circulation and freshwater biases, although separating their respective roles requires explicit heat and freshwater budgets.

335 4.4 Implications for Atlantic and Southern Ocean overturning

Our model ensemble establishes a relationship between thermohaline control and the mean AMOC: lower North Atlantic SCI and larger beta area are associated with weaker overturning, whereas higher SCI is associated with deeper winter mixing and stronger overturning. Independent analyses likewise find that subpolar sea-surface salinity covaries with mean and transient AMOC strength, including during abrupt $4 \times \text{CO}_2$ experiments (Dai et al., 2025). A physically plausible sequence is therefore
 340 that excess high-latitude freshwater strengthens haline control, suppresses winter mixing, and contributes to a low AMOC bias. The first step of that sequence, however, is a hypothesis rather than a result of SCI alone, and overturning can also feed back on salinity through heat and freshwater transport.

The consequences for climate-change sensitivity are not unidirectional. Cold–fresh, strongly stratified CMIP6 mean states show a smaller delayed AMOC response to North Atlantic Oscillation forcing than warm–salty states (Reintges et al., 2024).
 345 Under greenhouse warming, models with a less stratified North Atlantic and a stronger, deeper control-state AMOC tend instead to undergo larger absolute weakening because surface buoyancy anomalies penetrate more deeply (Bonan et al., 2025). Hosing experiments further show that recovery cannot be predicted from a single control-state property: it depends on the combined response and on the state reached when freshwater forcing ends (Jackson et al., 2023). Finally, Atlantic freshwater-transport biases can shift the simulated regime of AMOC bistability (Mecking et al., 2017). A beta-biased model may therefore
 350 simulate an AMOC that is too weak and too weakly responsive to some forcings, while its threshold or recovery under sustained freshwater input may be biased in another direction.

The Southern Ocean provides a parallel but dynamically distinct case. Excess freshwater and haline stratification can suppress deep convection, mode-water formation, and the production of dense shelf and bottom waters. Coordinated Antarctic meltwater experiments show reduced open-ocean convection and Antarctic Bottom Water formation, with response magnitude
 355 depending on the model mean state (Chen et al., 2023; Gülk et al., 2026). High-resolution simulations similarly project that Antarctic meltwater strengthens stratification and slows the abyssal overturning cell (Li et al., 2023). At mid-latitudes, CMIP6 stratification biases are also linked to Southern Ocean overturning and ocean heat uptake, and covary with North Atlantic biases (Liu et al., 2023; Vogt et al., 2025). The global winter beta bias found here may thus affect overturning in both hemispheres, but



its consequences differ between the wind- and eddy-influenced upper Southern Ocean cell, Antarctic bottom-water formation,
 360 and the North Atlantic overturning cell.

5 Conclusions

The SCI quantifies the temperature–salinity control of the stratification beneath the mixed-layer depth. In EN4, deep winter mixed layers occur preferentially in alpha stratification, whereas CMIP6 is systematically shifted toward beta stratification at high latitudes and places deep mixing closer to the beta-transition boundary. The contrast is much weaker in summer, when
 365 both observations and models are dominated by the seasonal thermocline. The main discrepancy therefore concerns the winter pycnocline and the thermohaline preconditioning of regions involved in deep-water formation.

In the North Atlantic, models with lower SCI and larger beta area also tend to have a weaker AMOC. This relationship does not identify a unique cause: the beta bias can result from errors in freshwater forcing and transport, sea ice, vertical mixing, subsurface temperature structure, or large-scale circulation pathways. These processes also interact, making a single
 370 explanation for the ensemble-wide bias unlikely. Increasing ocean resolution can improve major circulation features such as the Gulf Stream and North Atlantic Current and reduce associated temperature and salinity biases (Moreno-Chamarro et al., 2022), but higher resolution does not by itself guarantee a realistic high-latitude mean state and greatly increases the computational cost of producing large climate-model ensembles.

The systematic winter SCI bias has direct consequences for projections of overturning change. CMIP6 simulations are
 375 used to quantify future AMOC weakening (Weijer et al., 2020) and, in extended integrations, the risk of a shutdown beyond 2100 (Drijfhout et al., 2025). This is particularly concerning because stratification, convection, mixing and surface buoyancy forcing are integral controls of the global overturning circulation, while several of these processes remain poorly represented in climate models (Roquet et al., 2025). Because overturning sensitivity to warming and freshwater forcing depends on the initial stratification and water-mass structure (Reintges et al., 2024; Gülk et al., 2026), the widespread beta bias documented
 380 here calls into question the quantitative reliability of CMIP projections of overturning weakening or shutdown.

Code and data availability. All analyses and figures are reproducible in few steps.

- The MDT data are available at <https://www.aviso.altimetry.fr/en/data/products/auxiliary-products/mdt/mdt-global-cnes-cls.html> (last access: 1 October 2026) (CNES-CLS, 2023).
- The CMIP6 data are freely available via the Earth Grid System Federation.
- 385 – The EN4.2.2 data are available at <https://www.metoffice.gov.uk/hadobs/en4/download-en4-2-2.html> (last access: 1 october 2026)
- The MLD climatology is available at <https://www.seanoe.org/data/00806/91774/> (last access: 1 october 2026) (de Boyer Montégut, 2023).



- The code used to process all data and produce analyses and figures will be shared on a public git repository and archived on Zenodo upon acceptance for publication. They are for now shared with the reviewers at the following url: <https://cloud.univ-grenoble-alpes.fr/s/WAsPMZMXrKW7XJL>.
- The processed data will be archived on Zenodo upon acceptance for publication. They are for now shared with the reviewers at the following url: <https://cloud.univ-grenoble-alpes.fr/s/WAsPMZMXrKW7XJL>.

Appendix A: Interpolation method

The gridding of profile-derived quantities, such as MLD and SCI, is performed using radial basis function interpolation with the *scipy.interpolate.RBFInterpolator* class.

To account for the meridional change in the physical distance represented by one degree of longitude, latitude is transformed using a Mercator projection,

$$\varphi' = \ln \left[\tan \left(\frac{\varphi}{2} \frac{\pi}{180} + \frac{\pi}{4} \right) \right] \frac{180}{\pi}. \quad (\text{A1})$$

The transformed latitude φ' is then used together with longitude and time to define the distance between a target grid point and an observed profile:

$$d_i^2 = \left(\frac{\Delta\lambda}{L_\lambda} \right)^2 + \left(\frac{\Delta\varphi'}{L_{\varphi'}} \right)^2 + \left(\frac{\Delta t}{L_t} \right)^2. \quad (\text{A2})$$

Here $\Delta\lambda$ is the longitudinal separation, treated periodically, $\Delta\varphi'$ is the separation in transformed latitude, and Δt is the time separation. The decorrelation scales are $L_\lambda = L_{\varphi'} = 3.3^\circ$ and $L_t = 45$ d.

For each target grid point, the 300 closest profiles are selected. If fewer than 50 profiles are found within $d < 1$, the grid point is masked and no interpolation is performed. This mask avoids poorly constrained estimates in sparsely sampled regions.

The interpolation uses a Gaussian kernel,

$$w_i = \exp(-d_i^2 \epsilon^2), \quad (\text{A3})$$

with $\epsilon = 0.5$. The covariance matrix between profiles is

$$E_{ij}^0 = \exp(-d_{ij}^2 \epsilon^2), \quad (\text{A4})$$

and a constant noise-to-signal ratio is added,

$$\mathbf{E} = \mathbf{E}^0 + \lambda \mathbf{I}. \quad (\text{A5})$$

For a profile-derived variable collected in the vector Ψ , the interpolated value at location x is

$$s(x) = \Phi(x) \cdot \mathbf{E}^{-1} \cdot \Psi, \quad (\text{A6})$$

where $\Phi_i(x) = \exp(-d_{xi}^2 \epsilon^2)$.



415 Appendix B: CMIP6 variables

We use historical simulations from 26 coupled models participating in CMIP6 (Eyring et al., 2016), selecting the r1i1p1f1 ensemble member where available. For CNRM-CM6-1, HadGEM3-GC31-LL, and UKESM1-0-LL, the corresponding first realization is available with a different forcing index, and we therefore use r1i1p1f2, r1i1p1f3, and r1i1p1f2, respectively.

To investigate SCI and MLD, we use monthly output fields of sea water temperature ("thetao") and salinity ("so"), together
 420 with Grid Cell Area ("areacello") for area-weighted averaging. Following McDougall et al. (2021), we interpret the model temperature as Conservative Temperature, Θ , and the salinity as Preformed Salinity, S_* . It is recommended to interpret salinity as S_* for TEOS-10-based models and as S/u_{PS} for EOS-80-based models, where $u_{PS} = 35.16504/35$. To apply a consistent calculation across the model ensemble without distinguishing models by their underlying equation of state, we treat the archived
 "so" field directly as S_* for all models and use the TEOS-10 Gibbs SeaWater (GSW) functions throughout. For EOS-80-based
 425 models, this therefore neglects the small rescaling by u_{PS} recommended by McDougall et al. (2021).

To characterize the AMOC, we use monthly fields of either "msftmz" or "msftyz", depending on availability. Both represent the basin-integrated ocean overturning mass streamfunction but differ in their horizontal coordinate: msftmz is evaluated along geographic latitude, whereas msftyz follows the model's native ocean-grid y-coordinate. Where the streamfunction is provided as mass transport in kg s^{-1} , we convert it to volume transport in Sverdrups (Sv) by dividing by a reference seawater density,
 430 $\rho_0 = 1035 \text{ kg m}^{-3}$, and by 10^6 .

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Competing interests. The authors declare that they have no conflict of interest.

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440 EN.4.2.2 data were obtained from <https://www.metoffice.gov.uk/hadobs/en4/> and are ©British Crown Copyright, Met Office, 2013, provided under a Non-Commercial Government Licence <http://www.nationalarchives.gov.uk/doc/non-commercial-government-licence/version/2/>.

LLMs have been used to correct and refine English writing.



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