

Anthropogenic aerosols offsetting ocean warming less efficiently since the 1980s

Taimoor Sohail^{1,2}, Damien B. Irving³, Jan D. Zika^{1,2,4} and Jonathan M. Gregory⁵

¹School of Mathematics and Statistics, University of New South Wales, Sydney, Australia

²Australian Center for Excellence in Antarctic Science, University of New South Wales, Sydney, Australia

³Commonwealth Scientific and Industrial Research Organisation (CSIRO), Hobart, Australia

⁴UNSW Data Science Hub (uDASH), University of New South Wales, Sydney, Australia

⁵Department of Meteorology, University of Reading, Reading, United Kingdom

Key Points:

- Since 1980, aerosol-driven ocean cooling has decelerated substantially, alongside a drop in ocean heat uptake efficiency.
- The drop in ocean heat uptake efficiency is limited to the tropics, which have equilibrated to sustained aerosol-driven radiative forcing.
- Air-sea fluxes into the coldest fraction of the ocean continue to offset greenhouse gas-driven ocean warming.

Corresponding author: T. Sohail, t.sohail@unsw.edu.au

17 **Abstract**

18 Greenhouse gases and aerosols play a major role in controlling global climate change.
 19 Greenhouse gases drive a radiative imbalance which warms the ocean, while aerosols cool
 20 the ocean. Since 1980, the effective radiation felt by the planet due to anthropogenic aerosols
 21 has levelled off, global ocean cooling due to aerosols has decelerated, and greenhouse gas-
 22 driven ocean warming has accelerated. We explore the deceleration of aerosol-driven ocean
 23 cooling by quantifying a time- and spatially-varying ocean heat uptake efficiency, defined
 24 as the change in the rate of global ocean heat storage per degree of cooling surface tem-
 25 perature. In aerosol-only simulations, ocean heat uptake efficiency has decreased by 69%
 26 since the 1900s. The tropics and sub-tropics have driven this decrease, while the cold-
 27 est fraction of the ocean continues to sustain cooling and high ocean heat uptake effi-
 28 ciency. Our results identify a growing trend towards less efficient ocean cooling due to
 29 aerosols.

30 **Plain Language Summary**

31 The composition of the atmosphere has a major impact on our climate. Greenhouse
 32 gases warm the planet, while aerosols (i.e., suspensions of particles in the atmosphere)
 33 cool the planet, and most of this change is absorbed by the oceans. Since 1980, the rate
 34 of cooling of the planet due to aerosols has plateaued. In the past few decades, the ocean
 35 has begun to equilibrate to this change, and this work explores where and when this equi-
 36 libration has occurred in the ocean based on global climate models. To understand this
 37 change, we use an ‘ocean heat uptake efficiency’ metric which describes how much ad-
 38 ditional heat builds up in the ocean for a given degree of surface temperature gain (or
 39 loss). We find that the ocean is cooling more slowly given a degree of surface cooling due
 40 to aerosols compared to the pre-1980s. This change is largely driven by the tropics and
 41 sub-tropics, where the ocean has stopped cooling in response to aerosol-driven negative
 42 surface temperatures. Polar and sub-polar regions, however, continue to cool due to aerosols.
 43 These changes are occurring alongside accelerating greenhouse gas-driven warming, sug-
 44 gesting that the relative role of aerosols in cooling our climate is weakening.

45 **1 Introduction**

46 Variations in the composition of the earth’s atmosphere, namely, changing concen-
 47 trations of greenhouse gases, aerosols and ozone, cause anthropogenic climate change.
 48 Greenhouse gases (GHGs) primarily inhibit loss of heat to space, and thus warm the cli-
 49 mate system. The ocean absorbs a large proportion of the heat that consequently ac-
 50 cumulates. A recent estimate suggests that approximately 89% of the excess heat in the
 51 climate system has been absorbed by the global oceans since 1970 (Schuckmann et al.,
 52 2020). Anthropogenic aerosols (AAs) cool the climate system (Ramanathan et al., 2001;
 53 Gleckler et al., 2006), and thus offset some of the GHG-driven ocean warming. They also
 54 introduce spatial heterogeneity into the ocean warming pattern (Delworth et al., 2005;
 55 D. B. Irving et al., 2019).

56 In order to understand the effect of GHGs and AAs on the ocean and climate, single-
 57 forcing simulations have been conducted by the Detection and Attribution Model Inter-
 58 comparison Project (DAMIP) (Gillett et al., 2021), which is part of the Coupled Model
 59 Intercomparison Project (phase 6; CMIP6; Eyring et al. (2016)). In DAMIP, coupled atmosphere-
 60 ocean three-dimensional climate models are integrated from 1850 to 2014 in separate ex-
 61 periments forced by historical atmospheric concentrations of anthropogenic GHGs (*GHG-*
 62 *only*), aerosols (*AA-only*) and natural forcings (volcanic aerosol and variability in solar
 63 irradiance) individually, to compare their impact on the climate system (Gillett et al.,
 64 2016). The single-forcing simulations are compared with the standard CMIP6 histori-
 65 cal experiment which includes all forcings together. Past research with DAMIP simu-

66 lations has shown that GHG-driven ocean warming has been offset by aerosols across
 67 a range of depths in an ensemble of climate models (Bilbao et al., 2019). This offsetting
 68 effect is stronger in the Northern Hemisphere than the Southern Hemisphere due to un-
 69 even aerosol concentrations in the atmosphere (D. B. Irving et al., 2019).

70 Since 1980, however, the aerosol-driven change in top-of-atmosphere net downward
 71 radiative flux, or effective radiative forcing (ERF), has levelled off at approximately -1W/m^2
 72 relative to the pre-industrial climate (allowing for atmospheric adjustments) (C. J. Smith
 73 et al., 2021). In contrast, the GHG-driven effective radiative forcing has continued to grow,
 74 reaching over 2W/m^2 (relative to the pre-industrial climate) by 2014 (C. J. Smith & Forster,
 75 2021). As a result of this ERF stabilisation, ocean cooling due to aerosols has plateaued
 76 in the top 2000m of the ocean, while GHG-driven warming has continued to accelerate
 77 (Lyu et al., 2021). Assuming this trend continues in the future, Lyu et al. (2021) pro-
 78 duced future projections of top-2000m ocean heat content that are purely GHG-driven.
 79 However, the deep ocean adjusts to atmospheric changes on relatively long timescales,
 80 so the impact of the levelling off of aerosol-driven ERF on the *global* ocean heat uptake
 81 (that is, including the portion of the ocean deeper than 2000m) remains unclear. In gen-
 82 eral, it is unclear which regions of the ocean have equilibrated to the approximately con-
 83 stant aerosol ERF since 1980.

84 Changes in ocean heat content due to GHGs and aerosols may be expressed in terms
 85 of the ‘ocean heat uptake efficiency’ (in $\text{W m}^{-2} \text{K}^{-1}$), defined as the change in the rate
 86 of storage of heat by the global ocean per degree of change in the global-mean surface
 87 temperature (e.g., Raper et al. (2002); Kuhlbrodt and Gregory (2012)). For a given model,
 88 this diagnostic is a single number which measures the response of the physical mecha-
 89 nisms driving ocean heat uptake in that model. Typically, ocean heat uptake efficiency
 90 has been quantified in strongly forced (i.e., 1% CO_2 or abrupt- $4\times\text{CO}_2$) coupled climate
 91 models (Kuhlbrodt & Gregory, 2012; Exarchou et al., 2015). Despite its utility in un-
 92 derstanding past and future climate change, a similar estimate of ocean heat uptake ef-
 93 ficiency has not been produced for simulated historical climate change.

94 In this work, we explore aerosol-driven changes to the heat content tendency bud-
 95 get and their constituent processes and heat uptake efficiency in a suite of CMIP6 sim-
 96 ulations. In section 2, we lay out the diagnostics used in the analysis, and in section 3,
 97 we detail the models used in the study. Finally, in section 4 we explore the temporal and
 98 spatial change in heat content and heat uptake efficiency, followed by a discussion and
 99 summary of the results in section 5.

100 2 Theory

101 2.1 Global diagnostics

102 We begin by exploring the response of the global ocean to changing surface air tem-
 103 peratures and ERF since 1850. The globally-integrated ocean heat content is calculated
 104 from the three-dimensional temperature field $\Theta(x, y, z, t)$:

$$105 \mathcal{H}(t) = \rho_0 C_p \int \Theta(x, y, z, t) d\mathcal{V}, \quad (1)$$

106 where \mathcal{V} is the volume, ρ_0 is the density and C_p is the specific heat capacity of seawater.

107 The effective radiative forcing is sourced directly from C. J. Smith and Forster (2021)
 108 (see their figure 1d and h for equivalent plots for GHG-only and AA-only runs, respec-
 109 tively). C. J. Smith and Forster (2021) calculate ERF by applying the linearised energy
 110 balance equation for the Earth to a broad suite of CMIP5 and CMIP6 models: $\Delta F =$
 111 $\Delta N_{TOA} - \lambda \Delta T_a$, where F is the effective radiative forcing, N_{TOA} is the top-of-atmosphere

112 energy imbalance, T_a is the mean surface air temperature (relative to pre-industrial), and
 113 λ is the climate feedback parameter. We opt to source the values of ERF directly from
 114 C. J. Smith and Forster (2021) because their methodology enables us to track changes
 115 to ERF over the entire historical period.

116 2.2 Ocean heat uptake efficiency

117 For timescales greater than about a year, we may assume that that all of the heat
 118 storage in the earth system occurs in the ocean (Palmer & McNeall, 2014). In this case,
 119 ocean heat content tendency can be related to global-mean surface air temperature change
 120 T_a through the ocean heat uptake efficiency, as:

$$N/A_T = \kappa T_a, \quad (2)$$

121 where $N = \partial\mathcal{H}(t)/\partial t$ is the rate of ocean heat content change (in W), A_T is the total
 122 surface area of earth (in m^2) and κ is the ocean heat uptake efficiency (in $\text{Wm}^{-2}\text{K}^{-1}$).
 123 Typically, κ is conceived as a single number for a given model which captures all of the
 124 ocean heat uptake processes of the model, and does not depend on time or the nature
 125 of the forcing. However, here we do not assume that κ is a single, fixed value character-
 126 istic of a specific model, and instead, we use κ to characterise the response of model heat
 127 content to changing surface air temperature over different time periods and over differ-
 128 ent layers of the ocean. These diagnostics enable us to understand when, how and in which
 129 layers aerosols have modified the underlying temperature structure of the ocean in re-
 130 cent decades.

131 2.3 Layer-wise heat content tendency

132 In this work, we track changes to the ocean heat content in depth layers, as done
 133 by Bilbao et al. (2019). Ocean heat content as a function of depth (in Jm^{-3}) is calcu-
 134 lated by horizontally integrating the global temperature field:

$$\mathcal{H}(z, t) = \rho_0 C_p \int \int \Theta(x, y, z, t) dx dy, \quad (3)$$

135 We also investigate heat content in temperature-percentiles, following Sohail et al.
 136 (2021) and Holmes et al. (2022), to complement the depth-based analysis. Temperature-
 137 percentiles are defined in terms of the cumulatively-integrated ocean volume, from hot
 138 to cold, and represent the warmest $x\%$ of the ocean by volume. The temperature $\Theta^*(p, t)$
 139 of temperature-percentile p (between 0 and 100) is defined such that $p\%$ of the ocean
 140 volume is warmer than $\Theta^*(p, t)$, i.e.,

$$\int_{V_T} W(\Theta(x, y, z, t) - \Theta^*(p, t)) dx dy dz = 0.01pV_T, \quad (4)$$

141 where V_T is the total volume of the ocean, and the function $W(\delta) = 1$ for $\delta > 0$, and
 142 $W(\delta) = 0$ for $\delta \leq 0$.

143 Temperature-percentiles provide a complementary perspective to existing water mass-
 144 based methods of tracking property changes (e.g., at constant absolute temperature),
 145 as detailed in Sohail et al. (2021); Holmes et al. (2022); Sohail et al. (2022). By track-
 146 ing changes to the warmest $x\%$ of the ocean by volume, we:

- 147 1. Directly compare different models or observational products which may be uni-
 148 formly warmer or cooler, but have the same cumulative volume distribution in per-
 149 centile space.

- 150 2. Remove the effect of uniform warming or cooling over time, which would change
 151 the outcrop location of isotherms, making tracing property changes to a surface
 152 region more challenging.
- 153 3. Derive a simple heat content tendency budget where heat content tendency is a
 154 consequence of surface fluxes and mixing only (see Sohail et al. (2021) for further
 155 information).

156 Ocean heat content (in Jm^{-2}) of the part of the ocean colder than percentile p is
 157 calculated as:

$$\mathcal{H}(p, t) = 0.01V_T\rho_0C_p \int_0^p \Theta^*(p', t)dp', \quad (5)$$

158 where p is a given temperature-percentile, V_T is the total volume of the ocean, and $\Theta^*(p, t)$
 159 is the temperature at a given percentile. From equation (4), the volume bounded by a
 160 temperature-percentile surface and the domain boundary is $0.01pV_T$. To ease in the vi-
 161 sualisation of the results, we plot our results in terms of the time-mean temperature of
 162 a given percentile, Θ^p .

163 In temperature-percentiles, the ocean heat content tendency, $\partial\mathcal{H}(p, t)/\partial t$, is the sum
 164 of surface fluxes and mixing, $\partial\mathcal{H}(p, t)/\partial t = \mathbf{Q}(p, t) + \mathcal{M}(p, t)$. There is no term for ad-
 165 vection (that is, large-scale adiabatic transport in physical space but not in Θ^* space),
 166 which does not change the distribution of volume as a function of temperature. Thus,
 167 we can not only diagnose the temperature-percentile layers where ocean heat content changes
 168 the most, but also the surface fluxes and mixing that contribute to this change.

169 3 Models

170 We analyse temperature and surface fluxes in nine climate models from DAMIP
 171 single-forcing experiments and their CMIP6 historical simulations. In order to limit the
 172 influence of model internal variability on ocean heat content tendency results, we take
 173 an ensemble mean across all ensemble members available for each model. For the nine
 174 models analysed, between two and 15 ensemble members are archived, resulting in a to-
 175 tal of 48 ensemble members. Further details of the models analysed and their associated
 176 ensemble members are provided in the Supplementary Materials (Table S1).

177 We analyse the monthly-averaged ocean potential temperature (*thetao*), surface air
 178 temperature (*tas*) and net downward surface heat flux (*hfds*) from January 1850 to De-
 179 cember 2014 in the DAMIP and CMIP6 simulations, also using surface grid cell area (*area-*
 180 *cella*) and ocean grid cell volume (*volcello*). Ocean heat content in temperature percentiles
 181 $\mathcal{H}(p, t)$ is calculated by volume-integrating the three-dimensional temperature field Θ
 182 as a function of temperature percentile, p (Eq 5).

183 Heat content tendency N is calculated for a given temperature-percentile or depth
 184 layer or for the global ocean by taking the regression slope of ocean heat content against
 185 time in decadal windows, to obtain a quantity in J m^{-2} . The ocean heat uptake efficiency
 186 is calculated as N divided by T_a and by global surface area. The global surface heat flux
 187 is calculated by integrating the global area-mean of the two-dimensional surface heat flux
 188 field cumulatively in temperature-percentile. The result is cumulatively time-integrated
 189 and $\mathbf{Q}(p, t)$ is recovered as the regression slope of the time-integral against time in decadal
 190 windows. In this analysis, we assume the density of sea water is $\rho_0 = 1035 \text{ kgm}^{-3}$, and
 191 its specific heat capacity $C_p = 4000 \text{ J kg}^{-1} \text{ K}^{-1}$, unless directed otherwise by modelling
 192 groups.

193 Following D. Irving et al. (2020), Sohail et al. (2021) and Sohail et al. (2022), model
 194 drift is removed from all variables of interest by subtracting a fit as a function of time

195 during some reference simulation, in order to exclude any effect of drift on forced trends
 196 and the global ocean heat budget. For GHG-only and AA-only DAMIP simulations, which
 197 do not include natural forcing variations, we use a cubic fit to the 500 years of the pre-
 198 industrial control (*piControl*), which likewise has no natural forcing variation. A cubic
 199 fit is selected to ensure that centennial-scale drift is captured without overfitting onto
 200 modes of natural variability. For the CMIP6 historical simulations, a linear fit of the his-
 201 torical natural (*hist-nat*) simulations from the DAMIP single-forcing experiments is used,
 202 because the historical simulation *does* include historical natural forcings, and we wish
 203 to remove the long-term drift they cause, following Silvy et al. (2020). A linear fit is suf-
 204 ficient for this shorter (165 year) time series.

205 In the decades from 1995 to 2014, where the slowdown of aerosol-driven cooling is
 206 most evident, we focus on a single model (ACCESS-CM2) as an example, for which more
 207 detailed ocean heat budget tendency diagnostics are available (Bi et al., 2020). These
 208 diagnostics enable us to distinguish between the mixing mechanisms, shortwave redis-
 209 tribution and surface heat and volume fluxes that are driving ocean heat content change.
 210 Specifically, we diagnose the role of vertical mixing, numerical mixing, neutral diffusion,
 211 surface heat fluxes from surface volume fluxes, penetrative shortwave heating, total sur-
 212 face heat fluxes and other miscellaneous mixing on the ocean heat content in the AA-
 213 only and GHG-only runs of the ACCESS-CM2 model. As above, all heat content diag-
 214 nostics are surface- or volume-integrated onto temperature percentiles and de-drifted from
 215 the piControl run. Tendencies are calculated from regression slopes in each percentile
 216 layer against time over the 1994 to 2014 time period. The discussion and results of this
 217 single-model analysis are presented in Figure S1 of the Supplementary Materials.

218 4 Results

219 We begin by exploring the global effective radiative forcing, surface air tempera-
 220 ture and ocean heat content in the AA-only, GHG-only and historical (i.e. all forcings)
 221 simulations. Across all models, the GHG-only simulations exhibit a rapid acceleration
 222 in ERF, T_a and N since 1850 (figure 1a - c). However, the AA-only simulations have a
 223 different ocean and ERF response. As mentioned earlier, the ERF due to aerosols has
 224 plateaued since the 1980s at sustained levels below pre-industrial, as shown in figure 1a.
 225 This is largely mimicked by the surface air temperature field which has plateaued (fig-
 226 ure 1b). By contrast, although the rate of ocean cooling has slowed, ocean heat content
 227 has not stabilised (see figure 1c). This could be due to the fact that the ocean equilibrates
 228 to changing surface forcing over long (decadal or greater) timescales, which have not yet
 229 been reached (Yang & Zhu, 2011). There may also be compensation in ocean heat con-
 230 tent occurring between different oceanic regions, such that ocean warming in one region
 231 is offset by cooling elsewhere.

232 Figure 1d - f shows the quantities N , T_a and κ , which are related by equation (2),
 233 in the GHG- and AA-only and historical simulations in decadal time windows since 1855.
 234 Global ocean heat content tendency, N , has continued to grow since the 1980s in the GHG-
 235 only simulations (figure 1d), reflecting the acceleration of ocean warming due to green-
 236 house gases. Aerosol-driven ocean cooling, on the other hand, has begun to decelerate
 237 since the 1980s, matching the stabilisation of ERF and mean surface air temperature dur-
 238 ing this time (figure 1e), and N has tended to become less negative since the 1970s. The
 239 historical OHC, N , T_a , and ERF (purple lines in figure 1a - e) are a linear combination
 240 of the AA-only, GHG-only and natural forcings (not shown).

241 Ocean heat uptake efficiency κ has qualitatively different GHG-only, AA-only and
 242 historical responses. Prior to 1900, inter-model variability in κ is extremely large in all
 243 simulations due to the small signal in N and T_a . This is especially true in the histori-
 244 cal simulations, where a combination of a weak signal in N and T_a and higher inter-model
 245 variability leads to extremely uncertain estimates of κ until the 2000s and no discernible

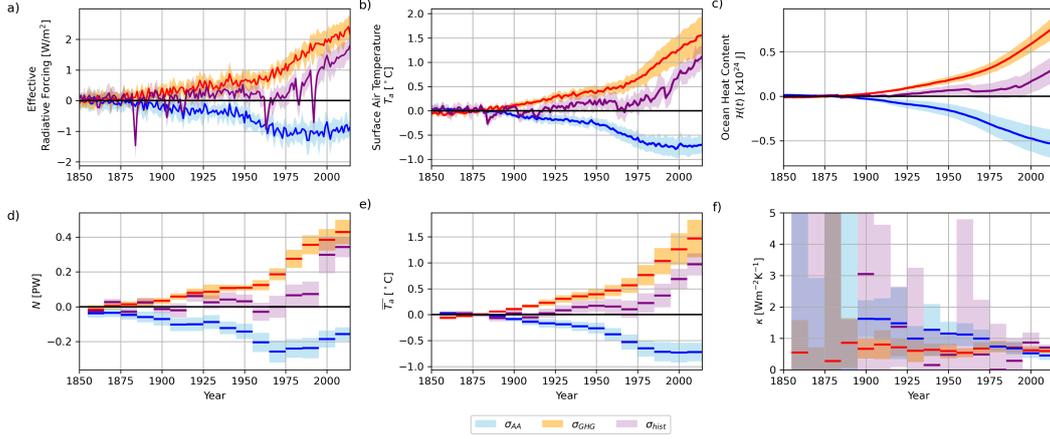


Figure 1. The multi-model-mean a) effective radiative forcing ERF (sourced from figure 1h of C. J. Smith and Forster (2021)), b and e) global-mean surface air temperature T_a , c) globally-integrated ocean heat content \mathcal{H} , d) globally-integrated ocean heat content tendency N , f) ocean heat uptake efficiency κ . All quantities are shown separately for GHG-only (orange), AA-only (blue) and historical (all-forcings, purple) simulations, and plotted relative to their 1850-1900 time-mean. (a-c) are annual means, (d-f) decadal means. The standard deviation (shaded regions) illustrates the inter-model variance of the nine model ensemble-means analysed.

246 trend (see purple lines and shading in figure 1f). In addition, the historical κ is not a
 247 linear combination of the two single forcing simulations. This non-linearity is inevitable,
 248 given that historical N and \overline{T}_a are linearly additive combinations of GHG-only and AA-
 249 only, but with different κ .

250 In the GHG-only simulations, κ has remained largely stable since the 1900s, at \sim
 251 $0.66\text{W}/\text{m}^2\text{K}$, as shown in figure 1f. This value of κ is broadly consistent with previous
 252 estimates in strongly-forced climate models and observations (Winton et al., 2014; Kuhlbrodt
 253 & Gregory, 2012; Cael, 2022). In AA-only simulations, κ has exhibited a sustained de-
 254 cline since the 1900s, from $\sim 1.6\text{W}/\text{m}^2\text{K}$ to $\sim 0.5\text{W}/\text{m}^2\text{K}$. The inter-model variabil-
 255 ity in this AA-only estimate has sharply reduced since the 1980s, signalling that the ocean
 256 in most model members is exhibiting this drop in ocean heat uptake efficiency. Prior to
 257 1970, AA-only ocean heat uptake efficiency is greater than that in the GHG-only simu-
 258 lations before becoming indistinguishable in more recent decades. This difference may
 259 be due to the fact that a large proportion of ocean cooling occurs in the rapidly venti-
 260 lating sub-polar and polar regions in the AA-only simulations, as indicated in the cold-
 261 est temperature-percentiles in figure 2.

262 Our results show that, globally, both the rates of ocean cooling and the ocean heat
 263 uptake efficiency have dropped in response to aerosols (blue lines in figure 1d and f). In
 264 order to distinguish *in which layers* this drop occurs, we calculate the layer-wise ocean
 265 heat uptake, following equations (5) for temperature-percentiles and (3) for depth lay-
 266 ers in decadal time windows (see figure 2a, b, e and f). The layer-wise κ is then $N(p, t)/\overline{T}_a$
 267 in temperature-percentiles and $N(z, t)/\overline{T}_a$ in depth (figure 2c, d, g and h). Due to the
 268 different layer thickness in temperature-percentiles and depth, we cumulatively integrate
 269 from hot to cold or from top to bottom to enable a direct comparison.

270 The profiles of the multi-model-mean ocean heat content tendency in temperature-
 271 percentiles (or depth) are shown in figure 2a and b for GHG-only and figure 2e and f for
 272 AA-only simulations, from 1945 to 2014. The ocean heat content tendency in GHG-only

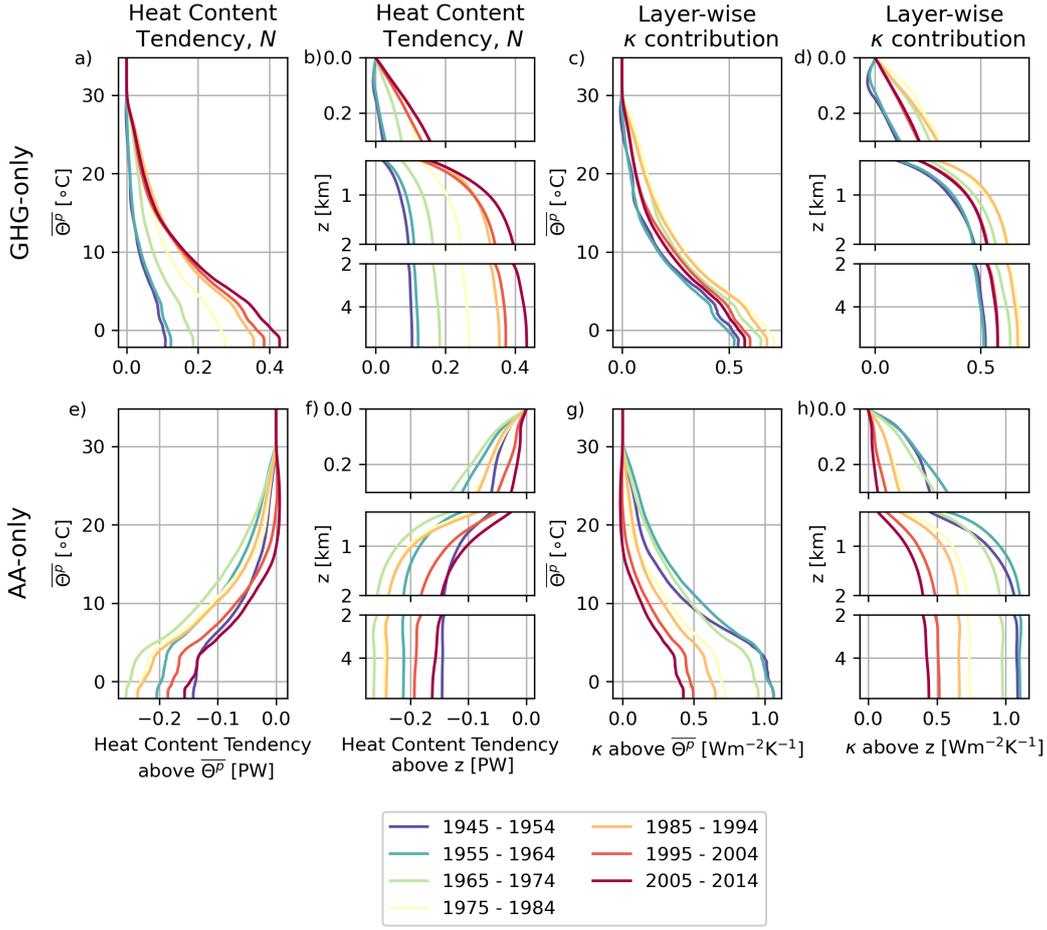


Figure 2. The multi-model-mean layer-wise ocean heat content tendency, N (a, b, e and f), and ocean heat uptake efficiency contribution, κ (c, d, g and h), in temperature-percentile (a, c, e and g) and depth (b, d, f and h) layers, across decadal windows from 1945-1954 (blue) to 2005-2014 (red). The top row shows the GHG-only simulation results (a - d), while the bottom row shows the AA-only simulation results (e - h). To ease in visualisation, we plot our temperature-percentile results in terms of the time-mean temperature of a given percentile surface, $\overline{\Theta^p}$.

273 simulations shows a warming across all temperature-percentile and depth classes (fig-
 274 ure 2a and b).

275 In an aerosol-forced ocean, prior to the 1980s, all water masses are cooling, and the
 276 $\overline{\Theta^p} > 5^\circ \text{C}$ ($z < 2 \text{ km}$) layer emerges as a key zone of ocean cooling (see figure 2e and
 277 f). After the 1980s, the rate of cooling declines, and there is warming after the 1990s above
 278 20°C . Since warming cannot be seen at any depth, even at the surface, we infer that, soon
 279 after the plateauing of aerosol ERF, cooling of the upper ocean ceases in the warm tropics
 280 and subtropics, where equilibration to forcing is rapid, while continuing at colder polar
 281 and sub-polar regions and in the global mean. Cooling continues also in the deep ocean
 282 (i.e. $N < 0$) at all latitudes, as the effect of previously growing negative ERF continues
 283 to spread downwards.

284 The layer-wise κ (shown in figure 2c, d, g and h) shows that under continuously
 285 increasing GHG forcing (figures 2c-d) all layers make a fairly constant contribution to

286 ocean heat uptake efficiency, which initially increases and later decreases slightly. In con-
 287 trast, for aerosol forcing (Figures 2g - h) the contribution is fairly constant until the ERF
 288 plateaus but subsequently declines at all levels. Since 1994, the tropics and sub-tropics
 289 (above 20°C) have stopped cooling in response to surface cooling (see orange and red
 290 lines in figure 2g). This is not seen in depth layers (figure 2h) due to the combination
 291 of surface tropics and sub-tropics with the surface polar and sub-polar regions.

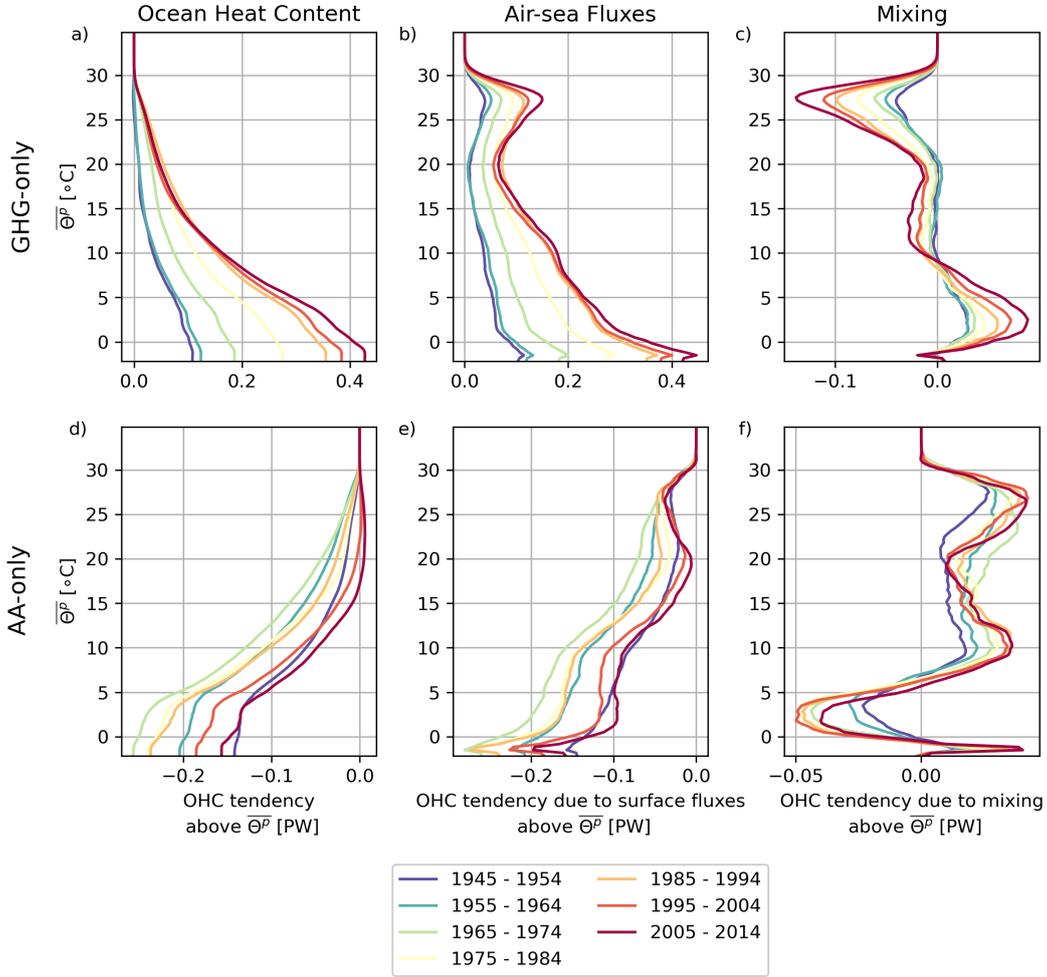


Figure 3. The multi-model-mean ocean heat content tendency (a and d), change in surface heat flux (b and e), and change in diathermal heat flux (c and f) in GHG-only (top row) and AA-only (bottom row) CMIP6 simulations, plotted in temperature-percentile space for decadal time windows from 1945 to 2014. Note that panels a) and d) are identical to panels a) and e) in figure 2. To ease in visualisation, we plot our results in terms of the time-mean temperature of a given percentile surface, $\bar{\Theta}^p$.

292 Air-sea heat fluxes can be diagnosed in temperature-percentiles from the DAMIP
 293 model outputs, and the diathermal mixing is calculated as the difference between the
 294 ocean heat content tendency and the change in surface heat flux, given that there are
 295 no other terms in the heat budget. Thus, in figure 3b and e, we present the ocean heat
 296 content tendency due to change in surface heat flux in the GHG-only and AA-only sim-
 297 ulations, respectively, and in figure 3c and f we show the tendency due to change in diather-

mal mixing in the GHG-only and AA-only simulations. The GHG-only simulations show a growing heat flux into the warmest fraction of the ocean ($\overline{\Theta^p} > 27^\circ \text{C}$) and parts of the sub-tropics and sub-polar ocean. Notably, there is a patch of surface cooling in the water masses aligning with $20 < \overline{\Theta^p} < 27^\circ \text{C}$, corresponding with the location of the sub-tropical gyres in the ocean (not shown). The mixing tends to oppose the surface flux tendency in most water masses, except in the temperature-percentiles corresponding to the polar and sub-polar regions and deep ocean ($4 < \overline{\Theta^p} < 10^\circ \text{C}$).

The surface heat flux in the AA-only simulations becomes increasingly negative for all temperature-percentile classes until the 1980s. Following the stabilisation of the AA-only ERF, surface flux-driven cooling comes to equilibrium with the stabilised forcing above $\overline{\Theta^p} = 20^\circ \text{C}$. Similarly to the GHG-only runs, a warm-cold dipole forms over the temperature-percentiles corresponding to the sub-tropical gyres in the AA-only simulations. This warm-cold dipole is consistent with a shift of surface warming (or cooling) towards the warmest temperature-percentiles at the ocean surface. Air-sea cooling below $\overline{\Theta^p} = 20^\circ \text{C}$, on the other hand, continues to drive a cooling at a constant rate despite the ERF stabilisation. This implies that AA-driven ocean cooling is sustained outside of the tropics and sub-tropics via air-sea fluxes.

Apart from the air-sea heat redistribution signal in the warmest temperature-percentiles (i.e., the bumps in figures 3b and e in the $20 < \overline{\Theta^p} < 27^\circ \text{C}$ layers), there is a strong qualitative similarity between the total OHC tendency and the OHC tendency due to the change in air-sea heat flux. This similarity suggests that surface heat uptake is transported mostly isothermally in the interior, consistent with the isopycnal analysis of (Saenko et al., 2021) and the isothermal analysis of (Sohail et al., 2021) in historical CMIP6 simulations.

Prior work from Exarchou et al. (2015); Saenko et al. (2021) has shown that high-latitude heat uptake is driven by vertical mixing and neutral diffusion in strongly warming climate models under idealised future forcing scenarios. In order to explore whether this is true in the CMIP6 experiments assessed here, we analyse the 1995-2014 time period in a single climate model - ACCESS-CM2 - for which detailed tendency budget terms are available. Our results corroborate this prior work, confirming that, in ACCESS-CM2, neutral diffusion (due to lateral eddies) and vertical mixing (likely due to convection) drive a significant amount of the ocean heat uptake in both the GHG-only and the AA-only simulations (see Figure S1 in Supplementary Materials). In addition, the mixing profiles in figure 3 (which are inferred as the residual of ocean heat content tendency and surface fluxes) match the explicitly diagnosed mixing in ACCESS-CM2. Prior work has established that the representation of high latitude ocean processes in CMIP6 models is highly uncertain (Heuzé, 2020), and our results emphasise the importance of better representing polar and sub-polar processes in future climate projections of ocean heat content.

5 Discussion

The temperature-percentile framework used in this work shows promise for use with other strongly-forced model simulations. The temperature-percentile framework enables more direct comparisons between models and observations which may have different mean states but have a similar volumetric temperature distribution. It also allows for large multi-model analyses by simplifying the heat budget and reducing the impact of drifting isotherms on the analysis of strongly-forced simulations, such as in future scenarios. On the other hand, the temperature-percentile framework collapses oceanic changes to a single dimension. We are thus unable to distinguish between hemispheric changes (likely important for aerosols, see D. B. Irving et al. (2019)), or across basins. New two-dimensional percentile methods, such as the partitioning method presented in Sohail et al. (2023), show

348 some promise in preserving spatial details in percentile space, and may be used to ex-
 349 tend this analysis into the future.

350 In this work, we have explored the slowdown in ocean cooling due to aerosols since
 351 1980. We use a set of single-forcing DAMIP simulations (comprising nine climate mod-
 352 els and 48 ensemble members in total) forced by greenhouse gases (GHGs) or anthro-
 353 pogenic aerosols (AA) to investigate the time period, layers and processes driving this
 354 deceleration in aerosol-driven ocean cooling. We find that, despite aerosols' historic role
 355 in suppressing warming due to GHGs, since 1980 the effective radiative forcing due to
 356 aerosols has stabilised and the ocean has begun to equilibrate to this new climate state.
 357 Globally, the AA-only ocean has begun to cool less rapidly, and the global ocean heat
 358 uptake efficiency, which represents the response of ocean heat content to a degree of sur-
 359 face warming or cooling, has dropped by 69%. These decelerating rates of cooling and
 360 reduced heat uptake efficiency have been confined to the tropics and sub-tropics, driven
 361 by warmer-than-usual air-sea fluxes. In contrast, in the coldest layers (corresponding to
 362 the polar and sub-polar regions), air-sea heat fluxes have continued to sustain ocean cool-
 363 ing and drive up ocean heat uptake efficiency. Thus, the polar and sub-polar regions can
 364 still be cooled efficiently in response to aerosol forcing, offsetting the relatively efficient
 365 ocean warming due to greenhouse gasses.

366 6 Open Research

367 The CMIP6 data used in this work is publicly available from ESGF: <https://esgf->
 368 node.llnl.gov/search/cmip6/. The full list of CMIP6 models used as well as their cor-
 369 responding ensemble members is provided in Table S1. The ACCESS-CM2 tendency di-
 370 agnostics, binned into temperature-percentile space (see Supplementary Materials), are
 371 available via *Zenodo*: 10.5281/zenodo.8123005. Effective Radiative Forcing data and code
 372 for calculations are sourced directly from (C. Smith, 2021).

373 Acknowledgments

374 We acknowledge the World Climate Research Program, the CMIP6 climate modeling groups,
 375 the Earth System Grid Federation (ESGF), and the funding agencies supporting CMIP6
 376 and ESGF. Modeling and analysis were undertaken with National Computational Infra-
 377 structure (NCI) facilities, supported by the Australian Government. This work is sup-
 378 ported by the Australian Research Council (ARC) Centre of Excellence for Climate Ex-
 379 tremes, the Australian Center for Excellence in Antarctic Science, the Centre for South-
 380 ern Hemisphere Oceans Research and the ARC Discovery Project scheme (DP190101173).

381 References

- 382 Bi, D., Dix, M., Marsland, S., O'Farrell, S., Sullivan, A., Bodman, R., . . . Heerde-
 383 gen, A. (2020). Configuration and spin-up of ACCESS-CM2, the new gen-
 384 eration Australian Community Climate and Earth System Simulator Cou-
 385 pled Model. *Journal of Southern Hemisphere Earth Systems Science*. doi:
 386 10.1071/es19040
- 387 Bilbao, R. A. F., Gregory, J. M., Bouttes, N., Palmer, M. D., & Stott, P. (2019).
 388 Attribution of ocean temperature change to anthropogenic and natural forcings
 389 using the temporal, vertical and geographical structure. *Climate Dynamics*,
 390 53(9-10), 5389–5413. doi: 10.1007/s00382-019-04910-1
- 391 Cael, B. B. (2022). Ocean Heat Uptake Efficiency Increase Since 1970. *Geophysical*
 392 *Research Letters*, 49(19). doi: 10.1029/2022gl100215
- 393 Delworth, T. L., Ramaswamy, V., & Stenchikov, G. L. (2005). The impact of
 394 aerosols on simulated ocean temperature and heat content in the 20th century.
 395 *Geophysical Research Letters*, 32(24). doi: 10.1029/2005gl024457
- 396 Exarchou, E., Kuhlbrodt, T., Gregory, J. M., & Smith, R. S. (2015). Ocean Heat

- 397 Uptake Processes: A Model Intercomparison. *Journal of Climate*, 28(2), 887–
398 908. doi: 10.1175/jcli-d-14-00235.1
- 399 Eyring, V., Bony, S., Meehl, G. A., Senior, C. A., Stevens, B., Stouffer, R. J., &
400 Taylor, K. E. (2016). Overview of the Coupled Model Intercomparison Project
401 Phase 6 (CMIP6) experimental design and organization. *Geoscientific Model
402 Development*, 9(5), 1937–1958. doi: 10.5194/gmd-9-1937-2016
- 403 Gillett, N. P., Kirchmeier-Young, M., Ribes, A., Shiogama, H., Hegerl, G. C.,
404 Knutti, R., . . . Ziehn, T. (2021). Constraining human contributions to ob-
405 served warming since the pre-industrial period. *Nature Climate Change*, 11(3),
406 207–212. doi: 10.1038/s41558-020-00965-9
- 407 Gillett, N. P., Shiogama, H., Funke, B., Hegerl, G., Knutti, R., Matthes, K., . . .
408 Tebaldi, C. (2016). The Detection and Attribution Model Intercomparison
409 Project (DAMIP v1.0) contribution to CMIP6. *Geoscientific Model Develop-
410 ment*, 9(10), 3685–3697. doi: 10.5194/gmd-9-3685-2016
- 411 Gleckler, P. J., Wigley, T. M. L., Santer, B. D., Gregory, J. M., AchutaRao, K., &
412 Taylor, K. E. (2006). Krakatoa’s signature persists in the ocean. *Nature*,
413 439(7077), 675–675. doi: 10.1038/439675a
- 414 Heuzé, C. (2020). Antarctic Bottom Water and North Atlantic Deep Water in
415 CMIP6 models. *Ocean Science*, 17(1), 59–90. doi: 10.5194/os-17-59-2021
- 416 Holmes, R. M., Sohail, T., & Zika, J. D. (2022). Adiabatic and Diabatic Signatures
417 of Ocean Temperature Variability. *Journal of Climate*, 35(5), 1459–1477. doi:
418 10.1175/jcli-d-21-0695.1
- 419 Irving, D., Hobbs, W., Church, J., & Zika, J. (2020). A Mass and Energy Conser-
420 vation Analysis of Drift in the CMIP6 Ensemble. *Journal of Climate*, 34(8),
421 3157–3170. doi: 10.1175/jcli-d-20-0281.1
- 422 Irving, D. B., Wjiffels, S., & Church, J. A. (2019). Anthropogenic Aerosols, Green-
423 house Gases, and the Uptake, Transport, and Storage of Excess Heat in the
424 Climate System. *Geophysical Research Letters*, 46(9), 4894–4903. doi:
425 10.1029/2019gl082015
- 426 Kuhlbrodt, T., & Gregory, J. M. (2012). Ocean heat uptake and its consequences for
427 the magnitude of sea level rise and climate change. *Geophysical Research Let-
428 ters*, 39(18). doi: 10.1029/2012gl052952
- 429 Lyu, K., Zhang, X., & Church, J. A. (2021). Projected ocean warming constrained
430 by the ocean observational record. *Nature Climate Change*, 1–6. doi: 10.1038/
431 s41558-021-01151-1
- 432 Palmer, M. D., & McNeall, D. J. (2014). Internal variability of Earth’s energy
433 budget simulated by CMIP5 climate models. *Environmental Research Letters*,
434 9(3), 034016. doi: 10.1088/1748-9326/9/3/034016
- 435 Ramanathan, V., Crutzen, P. J., Kiehl, J. T., & Rosenfeld, D. (2001). Aerosols,
436 Climate, and the Hydrological Cycle. *Science*, 294(5549), 2119–2124. doi: 10
437 .1126/science.1064034
- 438 Raper, S. C. B., Gregory, J. M., & Stouffer, R. J. (2002). The Role of Climate Sen-
439 sitivity and Ocean Heat Uptake on AOGCM Transient Temperature Response.
440 *Journal of Climate*, 15(1), 124–130. doi: 10.1175/1520-0442(2002)015<0124:
441 trocsa>2.0.co;2
- 442 Saenko, O. A., Gregory, J. M., Griffies, S. M., Couldrey, M. P., & Dias, F. B. (2021).
443 Contribution of Ocean Physics and Dynamics at Different Scales to Heat Up-
444 take in Low-Resolution AOGCMs. *Journal of Climate*, 34(6), 2017–2035. doi:
445 10.1175/jcli-d-20-0652.1
- 446 Schuckmann, K. v., Cheng, L., Palmer, M. D., Hansen, J., Tassone, C., Aich,
447 V., . . . Wjiffels, S. E. (2020). Heat stored in the Earth system: where
448 does the energy go? *Earth System Science Data*, 12(3), 2013–2041. doi:
449 10.5194/essd-12-2013-2020
- 450 Silvy, Y., Guilyardi, E., Sallée, J.-B., & Durack, P. J. (2020). Human-induced
451 changes to the global ocean water masses and their time of emergence. *Nature*

- 452 *Climate Change*, 1–7. doi: 10.1038/s41558-020-0878-x
- 453 Smith, C. (2021). chrisroadmap/cmip5-cmip6-forcing: Suppressed late-20th Century
454 warming in CMIP6 models explained by forcing and feedbacks. Retrieved from
455 <https://doi.org/10.5281/zenodo.5421581> doi: 10.5281/zenodo.5421581
- 456 Smith, C. J., & Forster, P. M. (2021). Suppressed Late-20th Century Warming
457 in CMIP6 Models Explained by Forcing and Feedbacks. *Geophysical Research*
458 *Letters*, 48(19). doi: 10.1029/2021gl094948
- 459 Smith, C. J., Harris, G. R., Palmer, M. D., Bellouin, N., Collins, W., Myhre, G.,
460 ... Forster, P. M. (2021). Energy Budget Constraints on the Time History
461 of Aerosol Forcing and Climate Sensitivity. *Journal of Geophysical Research:*
462 *Atmospheres*, 126(13). doi: 10.1029/2020jd033622
- 463 Sohail, T., Holmes, R. M., & Zika, J. D. (2023). Watermass Co-Ordinates Isolate the
464 Historical Ocean Warming Signal. *Journal of Climate*, 1–40. doi: 10.1175/jcli
465 -d-22-0363.1
- 466 Sohail, T., Irving, D. B., Zika, J. D., Holmes, R. M., & Church, J. A. (2021,
467 4). Fifty Year Trends in Global Ocean Heat Content Traced to Sur-
468 face Heat Fluxes in the Sub-Polar Ocean. *Geophysical Research Letters*,
469 48(e2020GL091439), 1 – 13. doi: 10.1029/2020gl091439
- 470 Sohail, T., Zika, J. D., Irving, D. B., & Church, J. A. (2022). Observed poleward
471 freshwater transport since 1970. *Nature*, 602(7898), 617–622. doi: 10.1038/
472 s41586-021-04370-w
- 473 Winton, M., Anderson, W. G., Delworth, T. L., Griffies, S. M., Hurlin, W. J., &
474 Rosati, A. (2014). Has coarse ocean resolution biased simulations of transient
475 climate sensitivity? *Geophysical Research Letters*, 41(23), 8522–8529. doi:
476 10.1002/2014gl061523
- 477 Yang, H., & Zhu, J. (2011). Equilibrium thermal response timescale of global
478 oceans. *Geophysical Research Letters*, 38(14), n/a–n/a. doi: 10.1029/
479 2011gl048076