Southern Ocean heat and momentum uptake are sensitive to the vertical resolution at the ocean surface

K.D. Stewart^{a,b,*}, A.McC. Hogg^{a,c}

^a*Research School of Earth Sciences, Australian National University, Australia*

^b*Climate Change Research Centre, University of New South Wales, Australia*

^c*Australian Research Council Centre of Excellence for Climate Extremes*

Abstract

The Southern Ocean plays a leading oceanic role in Earth's changing climate, accommodating over two thirds of the excess anthropogenic heat to date. In spite of this, climate and ocean models exhibit substantial variability in Southern Ocean heat uptake and storage, reflecting the modelling challenges of representing the responsible mechanisms. Here, Southern Ocean heat uptake is investigated in a suite of global ocean-sea-ice simulations run over a range of vertical and horizontal resolutions and forced with an idealized thermal perturbation. It is found that the ocean surface speed is sensitive to the vertical resolution at the ocean surface. Subsequently, the wind stress and its curl, which are important factors for Southern Ocean heat uptake, are also sensitive to the vertical resolution at the ocean surface, tending to decrease with refined surface resolution. It follows that the Southern Ocean heat uptake itself is sensitive to the vertical resolution at the surface; indeed, the simulations here with relatively coarse surface resolution (10m) exhibit heat content changes at rates nearly double that of simulations with finer surface resolution (1m). There is relatively little sensitivity of Southern Ocean heat uptake to the horizontal resolutions examined here (1° and 0.25°), although the higher resolution simulations exhibit increased wind-driven upwelling. Importantly, all simulations demonstrate that the location of the zero wind stress curl determines the location of the maximum heat uptake. These findings offer guidance for examining Southern Ocean heat uptake and storage in future modelling studies and observations.

Keywords:

Southern Ocean heat uptake; wind stress curl; Ekman/wind-driven upwelling/downwelling, vertical/horizontal resolution

1. Introduction

Global warming is predominantly ocean warming, with the oceans storing over 90% of the excess heat recently trapped in Earth’s climate system (Levitus *et al.*, 2012). Moreover, ocean warming is dominated by Southern Ocean warming, with the Southern Ocean accounting for between 67–98% of the total ocean warming since 2005 (Roemmich *et al.*, 2015; Llovel & Terray, 2016). This Southern Ocean warming is strongest in the upper 2km, primarily a result of the wind-driven advection, ventilation and subduction of Mode and Intermediate Waters and their subsequent circulations, as well as the reduction in high-latitude deep convection. The resulting pattern of warming is complex, time-dependent and inhomogeneous, challenging the limits of observations and representations in global climate models (e.g., Armour *et al.*, 2016; Sallée, 2018). Understanding the governing dynamics of Southern Ocean heat uptake is therefore of leading order importance for understanding and predicting the ocean’s response in Earth’s changing climate.

Southern Ocean heat uptake in the Coupled Model Intercomparison Project 5th phase (CMIP5) suite of climate models exhibits substantial variability between models; the inter-model standard deviation of total Southern Ocean heat uptake is 40% of the mean (Frölicher *et al.*, 2015). The partitioning of heat between the major Southern Ocean water masses in CMIP5 models is also variable, and is biased warm (Sallée *et al.*, 2013). Furthermore, Southern Ocean heat uptake is particularly sensitive to model biases in their parameterizations of air-sea fluxes and unresolved dynamics (e.g. Huber & Zanna, 2017). Considering the dominant role of the Southern Ocean in accommodating excess heat from the climate, it is a matter of urgency to improve the representation of Southern Ocean heat uptake in climate models.

Southern Ocean heat uptake is a result of the intricate interplay between surface fluxes of heat and momentum, ocean circulation, and the associated redistribution of the existing oceanic heat reservoir (e.g., Xie & Vallis, 2011; Frölicher *et al.*, 2015; Marshall *et al.*, 2015; Armour *et al.*, 2016). As such, investigations of Southern Ocean heat uptake requires detailed knowledge of all three contributors, all of which have non-trivial dynamics or representations in climate and ocean models (e.g., Morrison *et al.*, 2016). By perturbing the balance of these contributors and examining the ocean response, we can gain insight into their relative influences and potential feedbacks upon

one another and, ultimately, ocean heat uptake.

Global ocean–sea-ice models, as opposed to fully-coupled climate models, offer a practical means of isolating and attributing oceanic responses to the imposed perturbation in atmospheric forcings. In terms of Southern Ocean heat uptake, global ocean–sea-ice models with imposed, or partially-coupled, atmospheric thermal conditions are able to reproduce the principle responses of both observations and their fully-coupled climate model counterparts. These responses include a characteristic distribution of the heat uptake and ocean warming: the majority of the additional heat tends to enter the ocean poleward of $\sim 50^\circ\text{S}$ and be subsequently advected northward, converging around 45°S (e.g. *Marshall et al.*, 2015; *Armour et al.*, 2016; *Huber & Zanna*, 2017; *Liu et al.*, 2018). Importantly, the distribution of additional Southern Ocean heat uptake and storage is largely insensitive to the distribution of the atmospheric perturbation, suggesting that the thermal responses are primarily shaped by existing ocean currents rather than circulation changes. Therefore, for modelling the thermal response of the Southern Ocean to changes in atmospheric forcing, it is important to have an accurate representation of the existing ocean circulation, as well as an understanding of its sensitivity to atmospheric forcing and associated perturbations.

One dynamical consequence of the increase in greenhouse gases and stratospheric ozone depletion is the strengthening and poleward shifting of atmospheric winds over the Southern Ocean, which is evident in both observations and global climate models (e.g., *Thompson et al.*, 2000; *Kushner et al.*, 2001; *Marshall*, 2003; *Fyfe & Saenko*, 2006; *Swart & Fyfe*, 2012). These changes to the wind field influence the momentum uptake by the Southern Ocean, ultimately affecting the heat uptake (e.g., *Oke & England*, 2004; *Fyfe et al.*, 2007). Recent work detailed by *Llovel & Terray* (2016) uses three gridded Argo products to link the rapid upper ocean warming of the Southern Ocean to enhanced Ekman pumping associated with the poleward shift of the mean wind stress curl field. Their analysis identifies the South-East Indian and South Pacific Oceans as regions with the largest warming signals, where enhanced Ekman pumping acts to deepen isopycnal surfaces and allow for greater penetration of surface heat. That is, the adjustment of the surface momentum fluxes has modified the ocean surface density structure to facilitate additional heat uptake. A corollary to this result is that an adjustment of the surface heat fluxes could also be able to modify the ocean surface density structure (relative to the wind field) such that the ocean is able to uptake additional heat.

106 The 6th phase of Coupled Model Intercomparison Project (CMIP6) will
 107 feature some ocean models with nominal horizontal resolutions of 0.25° , a
 108 refinement from the nominally 1° models used for CMIP5 (*Griffies et al.*,
 109 2016). It is anticipated that this increased horizontal resolution will improve
 110 the representation of mesoscale dynamics and associated variability (e.g.,
 111 *Griffies et al.*, 2015). These efforts to increase the horizontal resolution can
 112 be undermined if unaccompanied by adequate improvements to the vertical
 113 resolution (e.g., *Stewart et al.*, 2017). Considering the important role of
 114 surface dynamics in Southern Ocean heat uptake, the representation of these
 115 surface dynamics in ocean models will likely govern the ability of the model
 116 to uptake heat. Thus, vertical resolution at the ocean surface will likely
 117 influence the modelled Southern Ocean heat uptake; as it is indeed the case
 118 for horizontal resolution (e.g., *Spence et al.*, 2010; *Morrison et al.*, 2013;
 119 *Newsom et al.*, 2016). Understanding how resolution (both vertical and
 120 horizontal) affects Southern Ocean heat uptake will aid in the comparison
 121 and interpretation of the CMIP5 and CMIP6 suite of models.

122 Here we investigate the effect of vertical and horizontal resolution on
 123 the uptake of heat and momentum throughout the Southern Ocean using a
 124 global ocean–sea-ice model with an imposed idealized warming scenario. We
 125 begin with a description of the model, forcing and methodology (§2), present
 126 analysis of how the surface forcing changes with model resolution (§3.1), and
 127 then examine the response of the Southern Ocean to an imposed thermal
 128 atmospheric perturbation (§3.2). Our perturbation simulations reproduce
 129 the characteristic distribution of additional heat uptake and storage (e.g.
 130 *Armour et al.*, 2016), which, following *Llovel & Terray* (2016), we examine
 131 in terms of changes in the wind stress curl field, with particular attention on
 132 the location of the zero wind stress curl in isopycnal space. These findings
 133 are discussed (§4), as well as their implications for Earth’s climate and future
 134 modelling efforts.

135 **2. Model & Methodology**

136 *2.1. Model Configurations & Forcing*

137 We use two configurations of the Australian Community Climate and
 138 Earth System Simulator (ACCESS) suite of global ocean–sea-ice models:
 139 ACCESS-OM2 and ACCESS-OM2-025, which are nominally 1° and 0.25°
 140 horizontal resolutions, respectively. These configurations have been devel-
 141 oped by the Consortium for Ocean–Sea-Ice Modelling in Australia (COSIMA)

so as to be as consistent as possible to permit direct comparison across resolutions (*Kiss et al.*, 2019). The ACCESS-OM2 suite of global ocean–sea-ice models use the Modular Ocean Model version 5.1 (MOM5.1; *Griffies*, 2012) with a z^* generalized vertical coordinate, and the Community Ice CodE version 5 (CICE5; *Hunke et al.*, 2015). For the ACCESS-OM2 configuration, four different vertical grids of varying resolutions are employed here (see Table 1 and *Stewart et al.*, 2017), ranging from a relatively-coarse 50-level grid with 10m grid spacing at the surface (GFDL50), to a higher-resolution 135-level grid with 1m grid spacing at the surface (KDS135). For the ACCESS-OM2-025 configuration, two of these vertical grids are used; the relatively-coarse GFDL50 grid, and a higher-resolution 75-level grid with 1.1m grid spacing at the surface (KDS75).

Both the GFDL50 and KDS75 vertical grids were developed specifically for use with the generalized z^* vertical coordinate of MOM5. For the GFDL50 vertical grid; when the ocean is at rest, the vertical grid spacing of the uppermost cell is 10m thick. The vertical grid spacing gradually and smoothly increases with depth to ~ 20 m grid spacing by ~ 200 m depth, and rapidly increasing to ~ 200 m grid spacing by ~ 2 km deep. In the case of the KDS75, the vertical grid spacing follows an hyperbolic tangent to increase smoothly with depth from 1.1m thick at the surface through ~ 150 m at ~ 2 km, to ~ 200 m at ~ 6 km deep; the KDS75 is objectively-constructed so as maximise the regions where the baroclinic mode-2 dynamics are permitted (see *Stewart et al.*, 2017). Note that for the purposes of this study the key difference between these two vertical grids is the resolution at the ocean surface; 10m for GFDL50 and 1.1m for KSD75.

The sub-grid parameterisations for under-represented variability of *Gent & McWilliams* (1990) (GM) and *Redi* (1982) are used for the ACCESS-OM2 simulations. The GM diffusivities are locally scaled in accordance with the ratio of local horizontal grid spacing to the first baroclinic Rossby radius, as suggested by *Hallberg* (2013), through the ranges of $50\text{--}600\text{m}^2\text{s}^{-1}$, with the Redi coefficient globally constant at $600\text{m}^2\text{s}^{-1}$. The K-profile parameterisation (KPP; *Large et al.*, 1994), used for the vertical diffusion of tracer and momentum in the surface boundary layer and unstably-stratified occurrences in the interior (i.e., convection is parameterised as an enhanced vertical diffusivity), operates in the simulations at both horizontal resolutions.

The models are initialized with temperature and salinity from the World Ocean Atlas 2013; these consist of the January ocean state above 1500m, and the boreal winter average state below 1500m (WOA13; *Locarnini et al.*,

2013; *Zweng et al.*, 2013). Surface forcing is a repeating year taken from the
high-resolution, self-consistent Japanese 55-year atmospheric reanalysis data-
set for driving ocean-sea-ice models (JRA55-do *Tsujino et al.*, 2018). The
JRA55-do offers atmospheric state, radiation and precipitation forcing fields
at 3-hourly, ~ 55 km resolution, as well as daily runoff fields that include solid
and liquid runoff from Greenland (from *Bamber et al.*, 2012) and Antarctica
(from *Depoorter et al.*, 2013). The repeated year forcing period is 1st May
1984 – 30th April 1985, referred to as RYF8485. In practice, the RYF8485
period runs from 1st January 1985 to 31st December 1984 with a sudden
transition from 30th April 1985 back to 1st May 1984. The date for this
transition is chosen so as to be at a time of year with low synoptic variability
at high latitudes and minimal deep water formation and convection, thereby
minimising any disturbances arising from the step change in forcing. The
justification for this period and forcing methodology is given in *Stewart et al.*
(in prep.); in essence, the RYF8485 is a quasi-climatological 12-month period,
with neutral values for all major climate indices. This repeated year forcing
allows for an extended period of model adjustment from the initial condition
so as to gauge model drift and long-term equilibration, and to serve as control
simulations for comparison with perturbation simulations. Surface salinity is
restored to a monthly climatology developed from WOA13, and the restoring
timescale is adjusted according to the vertical resolution at the ocean surface
such that the piston velocity of all simulations is 50m per 300 days.

The perturbation to the RYF8485 forcing is the globally-uniform increases
of both surface air temperature and downward longwave radiation, with the
remaining RYF8485 forcing fields unaltered. These increases are $+1.5^{\circ}\text{C}$
for surface air temperature and $+8\text{W/m}^2$ for downward longwave radiation,
selected to be representative of the CMIP5 multimodel mean responses in
these fields under the Representative Concentration Pathway 4.5 (RCP4.5)
scenario of moderate greenhouse gas emissions (see Fig. 2 of *Snow et al.*,
2016); here we refer to these perturbed forcings as RYF8485-RCP45. The
idealized nature of this globally-uniform thermal perturbation allows for a
direct comparison with the control simulations to assess excess heat uptake
by the ocean.

2.2. Methodology & Analysis

All control simulations are initialized from rest with the WOA13 hydrog-
raphy. The ACCESS-OM2 and ACCESS-OM2-025 control simulations are
run for 590 and 200 years, respectively, with the RYF8485 forcing. The

Configuration Reference	Horizontal Grid	Vertical Levels	Thickness at Surface	Start Year	Control Duration	Perturbation Duration
<i>GFDL50</i>	1°	50	10m	-300	590	290
KDS50	1°	50	2.3m	-300	590	290
<i>KDS75</i>	1°	75	1.1m	-300	590	290
KDS135	1°	135	1.0m	-300	590	290
<i>GFDL50-025</i>	0.25°	50	10m	-100	200	50
<i>KDS75-025</i>	0.25°	75	1.1m	-100	200	50

Table 1: ACCESS-OM2 configurations. The start years are relative to the addition of the perturbation. The four configurations in bold form the basis of the comparison from §3.1.2 onwards.

model states of the ACCESS-OM2 control simulations after 300 years, and after 100 years for ACCESS-OM2-025, are used to spin off perturbation simulations with the RYF8485-RCP45 forcing. The perturbation simulations for ACCESS-OM2 and ACCESS-OM2-025 are run for 290 and 50 years, respectively, and in parallel with their respective control simulations, allowing for the direct comparison of the perturbation-control differences between the various configurations. For ease of this comparison, we define model time relative to the time of perturbation, so that the ACCESS-OM2 and ACCESS-OM2-025 control simulations begin at model year -300 and -100, respectively, with all perturbation simulations beginning at model year 0. Considering the idealized nature of the imposed perturbation, particularly the lack of climate feedbacks on the atmospheric forcing, we focus our analysis on the initial response of the model at a time that is both long enough since perturbation to have an attributable signal, yet early enough that we can be sure the models are still in the same climate regime. So whilst the perturbation simulations are extended for nearly 3 centuries in the case of ACCESS-OM2, and 50 years for ACCESS-OM2-025, we focus on the 4th decade after the addition of the perturbation (model years 30-40).

The analysis is performed and presented in two parts. We first compare the surface forcings (of both heat flux and momentum transfer) of the simulations to assess the sensitivity to horizontal and vertical resolutions. This comparison involves globally- and zonally-averaged diagnostics, focusing on the differences between each configuration in the control state, and inferring the thermal responses we can expect from the perturbation simulations. We then compare these thermal responses, concentrating on the upper 2km of the

Figure 1: (a) Timeseries of the global mean ocean temperature for the control (solid) and RCP45 perturbation (dashed) simulations. The vertical dashed line indicates model year 0 when the perturbation is imposed. (b) Timeseries of the global mean temperature difference between the RCP45 perturbation simulations and their respective control simulation. The vertical dotted lines indicate the model years 30–40, the period of focus for much of the subsequent analysis.

242 Southern Ocean south of 35°S , and the model configurations that have com-
 243 mon vertical grids at both horizontal resolutions (GFDL50 and KDS75). This
 244 analysis extends from traditional spatial coordinates into isopycnal space,
 245 with a particular emphasis on identifying the isopycnals that outcrop near
 246 the latitude of zero wind stress curl.

247 3. Results

248 *Global Mean Ocean Temperature*

249 The global mean ocean temperature for the ACCESS-OM2 control sim-
 250 ulations exhibit a general cooling trend throughout the duration of the runs
 251 (Fig. 1a). This cooling rate is initially $\mathcal{O}(0.01)^{\circ}\text{C}/\text{decade}$ for the first cen-
 252 tury, and decreases to roughly half that for the remainder of the simulations.
 253 The cooling drift reflects an ongoing, gradual adjustment of the deep and
 254 abyssal waters from the initial hydrography. For these ACCESS-OM2 simu-
 255 lations, the global mean ocean temperature shows no obvious sensitivity to
 256 the vertical resolutions tested here.

257 In contrast to the 1° cases, the ACCESS-OM2-025 control simulations
 258 exhibit steady warming trends of $\sim 0.15^{\circ}\text{C}/\text{decade}$ for the duration of the
 259 simulations. Again, the drift reflects a continued adjustment of the deep and

abyssal waters from the initial hydrography. The warming drift is reduced by $\sim 30\%$ when vertical resolution is enhanced, but this reduced warming for ACCESS-OM2-025 isn't conclusive with only the two vertical grids examined here.

The addition of the RCP45 perturbation acts to increase the global mean ocean temperature (dashed lines in Fig. 1a). The difference in global mean ocean temperatures between the perturbation and control simulations shows the warming to initially be on the order of 0.05°C per decade before adjusting to 0.01°C per decade, for an increase of 0.1°C for the first 50 years (Fig. 1b). By way of comparison, this global ocean warming is approximately half the rate of that for the ACCESS Coupled Model versions (ACCESS1.0 and ACCESS1.3) contributed to the CMIP5 under the RCP4.5 scenario (Marshall *et al.*, 2013). This bulk thermal response over the first 5 decades to the perturbation exhibits no strong sensitivity to vertical or horizontal resolutions, despite the model drift in the 1° and 0.25° control simulations being of different signs. This result indicates that the initial response is surface-dominant, such that the convection-dominated deep ocean adjustment (which is responsible for the longterm drift in the control simulation) continues in the perturbation simulations. In addition, these consistent and predictable responses suggest the global ocean state of the perturbation and control simulations remain comparable for at least 50 years.

3.1. Surface Forcing

3.1.1. Control Simulations

A comparison of the zonally-averaged Southern Ocean surface heat fluxes (diagnosed by the ocean model) shows that the latitudinal structure is sensitive to the horizontal resolution, but largely insensitive to the vertical resolution tested here (Fig. 2a). Likewise, the upper 10m average ocean speed depends on horizontal resolution, with both ACCESS-OM2-025 cases exhibiting higher speeds between $45\text{--}55^\circ\text{S}$ (Fig. 2b), but is insensitive to vertical resolution. There is, however, a sensitivity to vertical resolution in the ocean surface speed ($\mathbf{u}_o$; Fig. 2c), with the surface speed increasing as the vertical grid spacing of the uppermost cell decreases; here from 10m thick (GFDL50) through 2.3m (KDS50) to $\sim 1\text{m}$ thick (KDS75 and KDS135). This result can be understood by considering the momentum transfer between the atmosphere and ocean; given the same variable wind field, the decreased ocean mass of thinner surface cells are able to accelerate and respond to the wind variability more readily than thicker surface cells. For the $45\text{--}55^\circ\text{S}$ latitude

Figure 2: The model year 30–40 means of the zonal averages of the (a) surface heat flux, (b) upper 10m ocean speed, (c) ocean surface speed, (d) the wind stress norm, (e) the difference in wind stress norm from the respective GFDL50 simulation (normalized by a nominal value of wind stress norm, 0.1N/m^2), and (f) the wind stress curl. The horizontal dotted line in (f) indicates the latitude where the zonally-averaged wind stress curl passes through zero.

band there is a $>30\%$ increase in ocean surface speed for simulations using vertical grids with surface resolutions of $\sim 1\text{m}$ (KDS75 and KDS135) relative to their 10m counterparts (GFDL50).

Ocean models typically use the ocean surface speed to dynamically calculate the sensible heat and evaporative fluxes, as well as the wind stresses exerted onto the ocean surface, by way of the relative difference between the wind and ocean surface velocities. In the cases of sensible heating and evaporation, the fluxes have linear dependences on both the relative difference between the wind and ocean surface velocities, and on the ocean–air temperature difference and specific humidity of the atmosphere, respectively. For wind stress, however, the dynamic forcing exerted onto the ocean has a quadratic dependence on the relative difference between the wind and ocean

309 surface velocities. Given the sensitivity of ocean surface speed to the vertical
 310 resolution at the ocean surface, we should expect the wind stress ($\boldsymbol{\tau}$; N/m²)
 311 to also exhibit a sensitivity to the vertical resolution at the ocean surface.

312 From *Pacanowski* (1987), the wind stress exerted on the ocean surface is
 313 computed as,

$$\boldsymbol{\tau} = C_d \rho_a |\mathbf{u}_a - \mathbf{u}_o| (\mathbf{u}_a - \mathbf{u}_o), \quad (1)$$

314 where C_d is the drag coefficient, ρ_a is the density of air, and $\mathbf{u}_a$ is the velocity
 315 of the atmospheric winds. Figure 2d shows the zonal average of the wind
 316 stress norm (N/m²), calculated as,

$$\|\boldsymbol{\tau}\| = \sqrt{\tau_x^2 + \tau_y^2}, \quad (2)$$

317 where τ_x and τ_y are the zonal and meridional components of the model di-
 318 agnosed wind stresses, respectively. The difference between the wind stress
 319 norm of a given simulation and that of the corresponding GFDL50 simula-
 320 tion is shown in Figure 2e, normalized by a nominal value of wind stress,
 321 0.1N/m². Simulations with thinner surface cells experience a reduced wind
 322 stress, as the increased ocean surface speed is closer the imposed wind speed.
 323 For the simulations with vertical resolutions of ~ 1 m at the surface (KDS75
 324 and KDS135), the reduction in wind stress is on the order of 2.5% between
 325 45–55°S. These reductions in wind stress are substantial, especially consid-
 326 ering that each simulation is forced with the same atmospheric wind fields,
 327 and that the differences in the wind stress norm vary systematically with the
 328 vertical resolution at the ocean surface.

329 Wind stress drives vertical velocities (Ekman pumping, written as w_{Ek})
 330 through the wind stress curl (N/m³),

$$w_{Ek} = \frac{\nabla \times \boldsymbol{\tau}}{\rho_0 f}, \quad (3)$$

331 which is of greater dynamical importance to vertical flow than the wind
 332 stress itself. In the Southern Hemisphere, regions of negative wind stress
 333 curl are upwelled (positive w_{Ek}), while positive wind stress curl results in
 334 downwelling (negative w_{Ek}). For all the JRA55-RYF8485 forced simulations
 335 examined here, the zero crossing of the zonally-averaged wind stress curl is
 336 at 52°S (Fig. 2f).

337 The distributions of Southern Ocean wind stress curl for the GFDL50
 338 simulations are shown in Figure 3a,c. The meridional structure indicated by
 339 Figure 2f is evident; the wind stress curl tends to be positive north of $\sim 45^\circ$ S

340 and negative south of $\sim 60^\circ\text{S}$, with the exception of Antarctic shelf regions
 341 (especially in ACCESS-OM2-025). However, unlike Figure 2f suggests, the
 342 transition from positive to negative wind stress curl is gradual and varies
 343 with longitude; this variation is especially strong for ACCESS-OM2-025,
 344 where the transition region is populated with relatively intense and localised
 345 occurrences of positive and negative wind stress curl. Values of the mean
 346 wind stress curl for regions representative of the Indian, Pacific and Atlantic
 347 basins are listed in Table 2.

348 By way of comparison, the differences in wind stress curl between the
 349 GFDL50 simulations and their respective KDS75 simulations are shown Fig-
 350 ure 3b,d. For ACCESS-OM2 (Fig. 3b), the wind stress curl difference is
 351 generally negative to the north of $\sim 55^\circ\text{S}$ and generally positive to the south;
 352 the differences in wind stress curl for the three basins are listed in Table 2.
 353 This pattern suggests that for regions where the wind stress curl are gener-
 354 ally positive (north of $\sim 52^\circ\text{S}$), the wind-driven downwelling in the KDS75
 355 simulation will be relatively weaker than that of the GFDL50 simulation.

356 Patterns in the wind stress curl difference for ACCESS-OM2-025 are less
 357 obvious, as the distribution is dominated by mesoscale features, particularly
 358 at higher latitudes. The strength of these mesoscale features are weaker
 359 in regions where the wind stress curl is generally positive, and tending to
 360 intensify to the south as the wind stress curl transitions to negative, especially
 361 in the Indian and Pacific basins. Overall, the wind stress curl difference
 362 between the ACCESS-OM2-025 simulations is generally negative where the
 363 wind stress curl is positive, and generally positive (with intense and localised
 364 regions of negative) where the wind stress curl transitions through zero to
 365 negative; again, this is reflected in the calculated differences of wind stress
 366 curl for the three regions (Tab. 2). These differences in wind stress curl
 367 suggest the simulations with coarse surface vertical resolution will tend to
 368 have enhanced wind-driven vertical motions.

369 The upwelling and downwelling of wind-driven Ekman flows are typically
 370 considered in terms of vertical motion, but their effect on interior circulation
 371 is to drive residual motion along isopycnals (e.g., *Doddridge et al.*, 2016). As
 372 such, it makes sense to examine the wind stress curl in isopycnal space; here
 373 we use σ_2 , the potential density referenced to 2000dbar. For this analysis, we
 374 bin the surface ocean south of 35°S in σ_2 -space and calculate the average wind
 375 stress curl for each σ_2 bin. Figure 4a,c shows distributions of the average wind
 376 stress curl by surface σ_2 for model years 30–40 for the GFDL50 simulations.
 377 These are the same data depicted in Figure 3a,c only now shown in σ_2 -

Region Geography/ σ_2	GFDL50	KDS75	1° diff. (%)	GFDL50 -025	KDS75 -025	0.25° diff. (%)
Nth. Indian	1.92	1.89	-1.44	1.91	1.89	-1.48
Sth. Indian	-2.10	-2.07	1.49	-1.98	-1.95	1.81
Nth. Pacific	1.44	1.41	-1.53	1.45	1.43	-1.64
Sth. Pacific	-0.98	-0.97	1.16	-0.94	-0.94	0.72
Nth. Atlantic	1.31	1.28	-2.17	1.30	1.29	-0.82
Sth. Atlantic	-1.28	-1.26	1.61	-1.23	-1.21	1.19
Light Indian	1.02	1.42	16.57	1.52	1.69	0.03
Dense Indian	-1.52	-1.41	9.75	-1.16	-1.31	-16.46
Light Pacific	0.62	0.91	24.63	1.31	1.38	4.00
Dense Pacific	-0.88	-0.83	9.00	-0.86	-0.86	0.72
Light Atlantic	0.87	1.00	11.22	1.12	1.18	5.02
Dense Atlantic	-0.64	-0.62	5.40	-0.54	-0.78	-30.72

Table 2: Values of the mean wind stress curl (N/m^3) in simulations with GFDL50 and KDS75 for the regions depicted in Figures 3 (upper 3 rows) and 4 (lower 3 rows). The KDS75-GFDL50 difference is given as a percentage of their average. For northern/light regions where the wind stress curl is positive, a negative difference (indicated in blue) suggests the KDS75 will experience less wind-driven downwelling. For southern/dense regions where the wind stress curl is negative, a positive difference (indicate in red) suggests the KDS75 will experience less wind-driven upwelling.

378 longitude space. Ekman downwelling occurs on isopycnals where the wind
 379 stress curl is positive, and Ekman upwelling where it is negative; these regions
 380 are separated by the isopycnal at the location of the zero wind stress curl,
 381 which is denoted here as σ_2^0 . For the GFDL50 control simulation, the zonal-
 382 average of $\sigma_2^0 = 1035.85 \text{ kg/m}^3$, and $\sigma_2^0 = 1035.75 \text{ kg/m}^3$ for GFDL50-025.
 383 Values for the mean wind stress curl in the “light” and “dense” regions of
 384 each major basin (indicated by the dashed regions) are listed in Table 2;
 385 these are positive and negative for the light and dense regions, respectively.
 386 While there appears much in common between the wind stress curl analysis
 387 in Eulerian and σ_2 -space, one important difference is the increased zonal
 388 coherency of the *isopycnal* of the zero wind stress curl σ_2^0 relative to that
 389 of the *latitude* of zero wind stress curl. This is particularly the case in the
 390 Indian and Pacific basins, where the increased zonal coherency is evidenced
 391 by the sharper transition between negative and positive wind stress curl and
 392 greater consistency in the isopycnal at which this transition occurs.

393 The difference in wind stress curl on surface σ_2 between the GFDL50
 394 simulations and the KDS75 simulations with relatively thinner surface cells
 395 are shown in Figure 4b,d. Again, these are the same data depicted in Figure
 396 3b,d mapped onto σ_2 -longitude space. It is apparent that the intensity of the
 397 wind stress curl differences have increased relative to the spatial distribution
 398 comparisons in Figure 3. This enhanced variability is to be expected; in Eule-
 399 rian space with a spatially-fixed wind field, the only way the wind stress curl
 400 can change is by a change in ocean surface speed (e.g., see Eqns. 1 and 3).
 401 In σ_2 -space, however, the isopycnals are able to move relative to the wind
 402 field, granting the wind stress curl on σ_2 an additional degree of freedom.
 403 The strongest features in Figure 3b,d remain evident in σ_2 -longitude space,
 404 as well as the relatively reduced differences in regions of predominantly pos-
 405 itive wind stress curl, especially for ACCESS-OM2-025. Also apparent in
 406 ACCESS-OM2-025 is the variability in the Atlantic extending to isopycnals
 407 of $\sigma_2 = 1034.5 \text{ kg/m}^3$, which is less-dense than the zonal average zero wind
 408 stress curl σ_2^0 of 1035.75 kg/m^3 . Importantly, the calculated differences for
 409 the light and dense regions of the major basins differ from those calculated
 410 for the corresponding north and south regions. Of note are the 16% and 30%
 411 differences in wind stress curl in the dense Indian and Atlantic regions of
 412 ACCESS-OM2-025, respectively, indicating that the KDS75-025 simulation
 413 will have enhanced upwelling of denser waters in these regions relative to the
 414 GFDL50-025 simulation. However, considering the relatively noisy distri-
 415 butions of the wind stress curl fields of the ACCESS-OM2-025 simulations,

Figure 4: As for Figure 3, except transformed into σ_2 -longitude space. The grey regions are isopycnals that do not outcrop (and hence have no value of wind stress curl) at that longitude. The zonally-averaged isopycnal of zero wind stress curl σ_2^0 is indicated by the solid line, and mean values of wind stress curl and ΔWSC in the dashed regions are listed in Table 2.

416 especially in the southern/dense Indian and Atlantic Ocean (e.g., Figs. 3c,d,
417 4c,d), these differences may be sensitive to the selection of the region extent.

418 3.1.2. RCP45 Perturbation Simulations

419 The RCP45 perturbation imposed here is purely thermal, consisting of
420 idealized, globally-uniform increases in both the atmosphere temperature
421 (+1.5°C) and downwelling longwave radiation (+8W/m²) fields. It is impor-
422 tant to note that this perturbation is not necessarily a direct increase of the
423 surface heat flux (SHF; W/m²) experienced by ocean model; for example,
424 dynamics in the ocean circulation and associated heat transports and/or re-
425 distributions can initiate feedbacks that mitigate SHF changes locally, even
426 resulting in a cooling of certain regions. Figure 5a shows the SHF anomaly
427 (Δ SHF; RCP45 perturbation minus control) for the KDS75-025 simulations
428 averaged for the model years 30–40; all cases explored here exhibit similar
429 Δ SHF distributions and magnitudes, especially for the Southern Ocean, and
430 are consistent with the characteristic distribution reported previously (e.g.
431 *Marshall et al.*, 2015; *Armour et al.*, 2016). The zonal averages of Δ SHF
432 for the simulations with the GFDL50 and KDS75 vertical grids are shown
433 in Figure 5b; these show the increased SHF is strongest poleward of $\sim 55^\circ$ S,
434 consistent with the findings of previous studies (e.g. *Frölicher et al.*, 2015;
435 *Liu et al.*, 2018). Importantly, there is no evidence of resolution dependence
436 to the SHF for the simulations examined here. The largest increases in SHF
437 occur poleward of the latitude of zonally-averaged zero wind stress curl, how-
438 ever even here the increase in SHF is approximately half that of longwave
439 radiation perturbation. Examining these Southern Ocean SHF anomalies in
440 σ_2 -space reveals a sensitivity to horizontal resolution, however the magni-
441 tude of the changes in SHF are again smaller than the imposed longwave
442 radiation perturbation (Fig. 5c). These relatively weak Δ SHF responses
443 suggest that substantial thermal anomalies in the RCP45 perturbation simu-
444 lations will likely be due to heat redistributions and/or changes in the ocean
445 circulations.

446 The thermal perturbation of the RCP45 simulations both act to lighten
447 the surface waters of the model ocean, thereby shifting the outcrops of isopy-
448 cnals poleward. In this way, the imposed wind field, which is not perturbed,
449 operates on relatively lighter surface waters, shifting the distribution of wind
450 stress curl on σ_2 towards less-dense waters (compare Figs. 5d and 5e). Im-
451 portantly, the isopycnal located at the zero wind stress curl σ_2^0 lightens; by
452 year 30–40 of the perturbed simulation, the density of σ_2^0 of all simulations

453 has decreased by 0.1–0.15 kg/m³.

454 The difference in wind stress curl between the perturbation and control
455 simulations on surface σ_2 is shown in Figure 5f. All simulations exhibit a
456 decrease in wind stress curl for isopycnals $\sigma_2 > 1035.2 \text{ kg/m}^3$, suggesting
457 that the RCP45 perturbation simulations will experience,

- 458 1. more wind-driven upwelling on isopycnals denser than the zero wind
459 stress curl σ_2^0 of their respective controls ($> \sigma_2^{0,c}$, where the superscript
460 c denotes the σ_2^0 of the control simulation);
- 461 2. less wind-driven downwelling on isopycnals lighter than their new σ_2^0
462 under the perturbation ($< \sigma_2^{0,p}$, where the superscript p denotes the σ_2^0
463 of the perturbation simulation); and
- 464 3. a range of isopycnals between $\sigma_2^{0,p} < \sigma_2 < \sigma_2^{0,c}$ that were downwelled in
465 the control simulations, now upwelled in the perturbation simulations.

466 These effects are schematically depicted in Figure 6. Taking these effects
467 into consideration, we anticipate the enhanced upwelling of isopycnals denser
468 than the isopycnal at the location of the zero wind stress curl in the RCP45
469 perturbation simulations will actually result in a subsurface cooling in regions
470 throughout the Southern Ocean. Additionally, the reduced downwelling of
471 isopycnals lighter than the isopycnal at the location of the zero wind stress
472 curl in the RCP45 perturbation simulations will increase the time that these
473 waters are near-surface and exposed to the warmer atmospheric conditions.
474 This concurrent enhanced upwelling and reduced downwelling either side of
475 the isopycnal location of the zero wind stress curl represents a watermass
476 transformation from dense-to-light waters. In σ_2 -space, these effects will
477 appear as a light-dense (north-south) dipole of heat increase and decrease
478 centered about the isopycnal of zero wind stress curl.

479 3.2. Thermal Response to the RCP45 Perturbation

480 Figure 7 shows the Southern Hemisphere zonal averages of temperature
481 difference in years 30–40 between the RCP45 perturbation and control simu-
482 lations for cases with the GFDL50 and KDS75 vertical grids. The enhanced
483 surface temperature increase is evident in all cases, as is the apparent absence
484 of deep ($> 2500 \text{ m}$) temperature changes. The upper ocean temperature in-
485 crease is strongest between 40–50°S and extending to depths nearing 700m;
486 this response is consistent with that observed and in fully-coupled climate
487 models (e.g., *Marshall et al.*, 2015; *Llovel & Terray*, 2016; *Armour et al.*,
488 2016), in spite of the idealized perturbation employed here. All simulations

Figure 5: (a) Distribution of surface heat flux anomaly (ΔSHF ; RCP45 perturbation minus control) of the KDS75-025 case averaged for model years 30–40. The latitude dashed is that of the zero wind stress curl. The latitudes dotted indicate the meridional extent of panel (b), which shows the zonally-averaged ΔSHF of years 30–40 for the cases with the GFDL50 and KDS75 vertical grids. For reference, the imposed downward longwave radiation perturbation is indicated by the vertical dotted line at 8W/m^2 , as with the latitudinal location of the zero wind stress curl (52°S). (c) The same ΔSHF values as (b), only transformed into surface σ_2 coordinates and zonally averaged. (d,e) The zonally averaged wind stress curl by surface σ_2 for the region south of 35°S for the control and RCP45 perturbation simulations, respectively, with their difference shown in (f). For (c-f), the dotted horizontal lines indicate the isopycnal locations of the zero wind stress curls for the relevant simulations.

Figure 6: A schematic representation of the expected changes in wind-driven motions along isopycnals between the control and perturbation simulations.

489 exhibit a deepening and southward shift of the σ_2 isopycnals, particularly
 490 for the $\sigma_2=1035.5$ and 1036 kg/m^3 isopycnals, and $\sigma_2=1037 \text{ kg/m}^3$ isopy-
 491 cnal for ACCESS-OM2-025 south of 60°S . The overall distribution of the
 492 temperature increase through the upper 1000m north of 50°S appears rel-
 493 atively insensitive to horizontal and vertical resolution, including isolated
 494 patches of temperature decrease in the upper tropics. South of 40°S , how-
 495 ever, the ACCESS-OM2-025 simulations exhibit a temperature decrease for
 496 $1036.5 < \sigma_2 < 1037 \text{ kg/m}^3$ that extends towards 250m depth that is not ap-
 497 parent in ACCESS-OM2. We propose that this decrease in temperature
 498 likely results from enhanced upwelling along these isopycnals in these simu-
 499 lations where eddies are partially resolved and augmented with a weak eddy
 500 parameterisation. Both simulations with KDS75 display increased thermal
 501 anomaly deeper than $\sim 1\text{km}$ and south of 60°S , likely resulting from the de-
 502 crease in convection due to surface warming; this response is stronger in
 503 ACCESS-OM2-025, reflecting the increased role of convection in the 0.25°
 504 model relative to the 1° model.

505 The sensitivity of the warming response to the RCP45 perturbation can
 506 be better characterised by partitioning the zonally-averaged temperature dif-
 507 ferences into the major ocean basins (Fig. 8). In all simulations, the response
 508 of the Atlantic sector is most unlike the global response, reflecting the leading
 509 role of the southern Pacific and Indian Oceans in global ocean heat uptake.
 510 The largest impact of horizontal resolution relates to the temperature de-
 511 crease signal associated with upwelling between $1036.5 < \sigma_2 < 1037 \text{ kg/m}^3$ in
 512 all basins. This signal is apparent in both ACCESS-OM2-025 simulations,
 513 reaching depths shallower than 250m in the Atlantic and Indian sectors,
 514 and is stronger in the KDS75-025 simulation, as suggested by the relative
 515 strength of the wind stress curl in these regions (Tab. 2). In all simula-
 516 tions, the temperature increase of the upper 700m in the Atlantic is centered
 517 around 40°S and $\sigma_2 = 1035.5 \text{ kg/m}^3$, which is relatively northward and on
 518 lighter isopycnals than the warming in the Indo-Pacific basins. There is,
 519 however, the strongest evidence of a temperature decrease by upwelling in
 520 the Atlantic in all simulations, especially KDS75-025. Interestingly, isopy-
 521 cnal $\sigma_2 = 1036.5 \text{ kg/m}^3$ that bounds the strong cold signal in this sector
 522 is the only isopycnal to exhibit a northward shift under the perturbation,
 523 although the outcrop of this isopycnal has shifted southward. In all simu-
 524 lations, the Indian sector exhibits a temperature increase on isopycnals
 525 spanning $1035 < \sigma_2 < 1036.5 \text{ kg/m}^3$.

526 Temperature differences in Eulerian space cannot provide a complete pic-

Figure 7: Zonally-averaged temperature difference between the RCP45 perturbation and control simulations for model years 30–40, for (a) the 1° simulation with the GFDL50 vertical grid; (b) 1° simulation with the KDS75 vertical grid; (c) 0.25° simulation with the GFDL50 vertical grid and (d) 0.25° simulation with the KDS75 vertical grid. The $\sigma_2=1034\text{--}1037\text{ kg/m}^3$ contours are included, at 0.5 kg/m^3 intervals, for the control (solid) and RCP45 perturbation (dashed) simulations, with the $\sigma_2=1036\text{ kg/m}^3$ coloured magenta for reference, and the σ_2^0 contoured in green. The dotted lines bound the 2000m, 35°S region of interest for subsequent analysis.

Figure 8: As for Figure 7, focusing on the region bound by 2000m and 35°S, showing the zonal average temperature difference by major ocean basin. Again, the $\sigma_2=1034$ – 1037 kg/m³ contours are included, at 0.5 kg/m³ intervals, for the control (solid) and RCP45 perturbation (dashed) simulations, with the $\sigma_2=1036$ kg/m³ coloured magenta for reference, and the σ_2^0 contoured in green.

527 ture of ocean heat uptake. A more comprehensive view can be gained by
 528 employing frameworks that delineate the movement of isopycnal surfaces
 529 and the transformation of watermasses. One such approach is to compare
 530 the responses of the meridional overturning streamfunctions in σ_2 -latitude
 531 space, denoted here as Ψ (Sv). Figure 9a-d shows the differences in South-
 532 ern Hemisphere overturning streamfunctions, $\Delta\Psi = \Psi^p - \Psi^c$, between the
 533 RCP45 perturbation (Ψ^p) and control simulations (Ψ^c) for cases with the
 534 GFDL50 and KDS75 vertical grids. All simulations exhibit an increased
 535 overturning of the upper cell of over 5Sv in locations between 60–30°S and
 536 $\sigma_2=1035.5\text{--}1036.0\text{ kg/m}^3$ (i.e., spanning the isopycnal locations of zero wind
 537 stress curl). These increases in upper cell overturning correspond with a shift
 538 in the streamfunction distributions to lighter isopycnals; in Figure 9a-d, these
 539 shifts appear as an isopycnal displacement of the upper cell Ψ contours by
 540 approximately $0.1\text{--}0.15\text{ kg/m}^3$. This isopycnal shift of the upper overturning
 541 cell, which is often referred to as the adiabatic “wind-driven” overturning
 542 cell (e.g., *Wolfe & Cessi*, 2011), is consistent with the isopycnal shift of the
 543 wind stress curl between the RCP45 perturbation and control simulations
 544 (e.g., Figs. 5,6). That is, in the RCP45 perturbation simulations, the im-
 545 posed wind field operates on relatively less-dense surface waters, resulting
 546 in a wind-driven overturning circulation of similar magnitude and structure,
 547 only shifted towards lighter isopycnals. There is relatively little change in the
 548 structure and magnitude of the lower cell overturning northward of $\sim 60^\circ\text{S}$.

549 Examining the overturning streamfunctions in σ_2 -latitude space provides
 550 a common framework in which we can compare simulations with differ-
 551 ent vertical grids. Figure 9e,f shows the difference in the overturning re-
 552 sponses between the simulations with the KDS75 and GFDL50 vertical grids,
 553 $\Delta\Delta\Psi = \Delta\Psi_{KDS75} - \Delta\Psi_{GFDL50}$. At both horizontal resolutions, the upper
 554 cell overturning of the simulations with the GFDL50 vertical grids is more
 555 affected by the RCP45 thermal perturbation than those of the KDS75 ver-
 556 tical grids. This result is consistent with the relatively weaker wind stress
 557 and wind stress curl experienced by the KDS75 simulations due to the finer
 558 vertical resolution at the ocean surface.

559 To further delineate the effects of isopycnal movements and watermass
 560 transformations, the model output for the region bound by 2000m and 35°S
 561 is binned in σ_2 -space, and its total heat content, Q (J), is calculated as,

$$Q(\sigma_2) = \sum \sigma_2 c_p T_{\sigma_2} V_{\sigma_2}, \quad (4)$$

562 where c_p is the specific heat capacity of seawater, and T_{σ_2} and V_{σ_2} are the

Figure 9: (a-d) Southern Hemisphere differences in overturning streamfunctions ($\Delta\Psi$) in σ_2 -latitude space for the simulations with the GFDL50 and KDS75 vertical grids. The streamfunctions for the control and RCP45 perturbation cases are contoured in black and green, respectively; the contour levels are at $\Psi = \pm 1, 6, 12, 18$, with dashed contours as negative. (e,f) show the differences between the overturning anomalies ($\Delta\Delta\Psi$) of the KDS75 and GFDL50 cases for the ACCESS-OM2 and ACCESS-OM2-025 simulations, respectively. Note these anomaly differences are scaled by a factor of 2 so as to be visible with the same colorscale.

563 temperatures and volumes on σ_2 , respectively. From this formulation, and
564 with a sufficiently narrow σ_2 bin, it is clear that changes in Q in a given σ_2
565 bin can arise from changes in T_{σ_2} or V_{σ_2} . Considering the equation of state
566 of seawater, substantial changes in T_{σ_2} will require density-compensating
567 changes in salinity in order for a water parcel to remain in a given σ_2 bin.
568 Given that we impose no such perturbations to the hydrological forcing,
569 we therefore expect the largest changes in $Q(\sigma_2)$ to arise from changes in
570 V_{σ_2} . That is, heat increases or decreases in σ_2 -space are characterised by the
571 transformation of volume between σ_2 bins.

572 In order to compare between the model resolutions, the perturbation-
573 to-control change in total heat content for a given σ_2 bin is divided by the
574 product of the thickness-weighted area average of that σ_2 bin and the cumu-
575 lative model time under perturbation, returning a value for the rate of change
576 of heat content for a given σ_2 bin, in dimensions of W/m^2 . Figures 10a-d
577 show this rate of heat content change in σ_2 -longitude space for the simulations
578 with GFDL50 and KDS75 vertical grids. Also indicated is $\sigma_2^{0,c}$, the northern-
579 most σ_2 contour of the control simulation with zero wind stress curl at each
580 longitude. The expected light-dense (north-south) dipole of heat increases
581 and decreases about the σ_2^0 is obvious in all simulations. This representation
582 of heat uptake, which is specifically the movement of volume between den-
583 sity classes, highlights the close coupling between the heat uptake and the
584 location of the zero wind stress curl in σ_2 -space. The dominant roles of the
585 Pacific and Indian basins in ocean heat uptake are evident, particularly in
586 regions with a zonally-coherent zero wind stress curl location in σ_2 -longitude
587 space. Furthermore, the lack of zonal coherence of the zero wind stress curl
588 σ_2^0 in the Atlantic sector may account for the relatively weaker rates of heat
589 uptake estimates there.

590 From Figures 10a-d it is apparent that the changes in heat content in
591 σ_2 -space for the simulations with GFDL50 are relatively stronger than those
592 with KDS75. This is more evident in Figures 10e,f, which shows the dif-
593 ference between the rates of heat content change for KDS75 and GFDL50
594 simulations in σ_2 -longitude space at 1° and 0.25° degree resolutions, respec-
595 tively. Here the negative values for light $\sigma_2 < 1035.5 \text{ kg}/\text{m}^3$ indicate a rela-
596 tively weaker heat increase (downwelling), and the positive values for denser
597 $\sigma_2 > 1035.5 \text{ kg}/\text{m}^3$ indicate a relatively weaker heat decrease (upwelling), in
598 the KDS75 simulations. The difference in the rate of heat content change on
599 a given σ_2 is more pronounced in the ACCESS-OM2-025 simulations. This
600 result that the GFDL50 simulations exhibit stronger rates of heat content

601 change on σ_2 is consistent with the relative differences in surface momen-
 602 tum fluxes and wind stress curl on σ_2 between the coarse-surface-resolution
 603 GFDL50 and finer-surface-resolution KDS75 (e.g., Fig. 2-4).

604 The enhanced rates of heat content change exhibited by the simulations
 605 with GFDL50 are more apparent in their zonal averages either globally or by
 606 major ocean basin (Fig. 11a-d). The heat uptake rates for $\sigma_2 < 1035.5 \text{ kg/m}^3$
 607 appear to be more sensitive to the vertical grid than the horizontal resolu-
 608 tion; here the two simulations with the GFDL50 grid experience heat uptake
 609 rates at nearly twice that of the two simulations with KDS75. The isopycnal
 610 located at the zero wind stress curl represents a breakpoint in the heat up-
 611 take globally and in each basin; isopycnals immediately denser than the zero
 612 wind stress curl σ_2 exhibit the largest rates of heat decrease. This effect is
 613 least obvious in the Atlantic basin where the heat content change on isopy-
 614 cnals appears more complicated, presumably reflecting the zonal variability
 615 of the wind stress curl in both Eulerian and σ_2 -space (Figs. 3 and 4). Addi-
 616 tionally, the effect of horizontal resolution is most obvious in the strong heat
 617 decrease of the σ_2 immediately denser of the zero wind stress curl σ_2 , a signal
 618 associated with the enhanced wind-driven upwelling apparent in ACCESS-
 619 OM2-025.

620 We can take this analysis further by calculating the cumulative aver-
 621 age rates of heat content change on isopycnals between $\sigma_2=1034.0 \text{ kg/m}^3$
 622 (approximate density of the lightest surface waters at 35°S) and σ_2 , both
 623 globally and by major ocean basin (Fig. 11e-h). These curves indicate the
 624 average Southern Ocean heat uptake on σ_2 is positive and increasing up to
 625 the zero wind stress curl σ_2^0 , beyond which it decreases rapidly, highlighting
 626 the importance of this isopycnal outcrop for locating heat uptake. Again,
 627 this diagnostic suggests the heat uptake is more sensitive to the vertical grid
 628 than the horizontal resolution, with the simulations that use the relatively
 629 coarse GFDL50 grid exhibiting heat increases at nearly twice the rate of the
 630 KDS75 simulations.

631 4. Discussion

632 The simulations and analysis shown in this manuscript demonstrate a
 633 number of important points which we list here and develop further in the
 634 following paragraphs:

- 635 1. that the surface speed, wind stress and wind stress curl are sensitive to
 636 the vertical resolution at the ocean surface;

Figure 10: (a-d) The rates of heat content change (W/m²) in σ_2 -space by longitude for the region bound by 2000m and 35°S in the simulations with GFDL50 and KDS75 vertical grids at 1° and 0.25° horizontal resolutions. The northern-most location of zero wind stress curl for densities >1035 kg/m³ are indicated by the black dots. The KDS75-GFDL50 rate differences for (e) 1° and (f) 0.25° horizontal resolutions. Note the difference are scaled by a factor of 2 so as to be visible with the same colorscale.

Figure 11: Zonal averages of the rates of heat content (W/m^2) in σ_2 -space for the simulations with GFDL50 and KDS75 vertical grids for the (a) Southern Ocean region of interest, (b) Indian, (c) Pacific and (d) Atlantic basins. For reference, the vertical dotted line at 8W/m^2 indicates the imposed downward longwave radiation of the RCP45 perturbation, and the horizontal dotted lines indicate the range of the locations of zonally-averaged zero wind stress curl isopycnal, σ_2^0 for each of the regions. (e-h) The cumulative average rates of heat content change for σ_2 bins denser than $\sigma_2=1034.0\text{kg/m}^3$.

- 637 2. that the surface momentum fluxes and associated upwelling can change
638 from a purely thermal forcing perturbation;
- 639 3. that the zero wind stress curl isopycnal is critical to determine which
640 regions experience an increase or decrease of heat when the system is
641 perturbed;
- 642 4. that increased rates of heat uptake can occur with relatively coarse
643 vertical resolution at the ocean surface; and
- 644 5. that the upwelling branch of the Southern Ocean circulation is en-
645 hanced when going from 1° to 0.25° resolution.

646 The sensitivity of the surface speed to the vertical resolution at the ocean
647 surface (Fig. 2c) leads to a similar sensitivity in the wind stress field (e.g.,
648 Fig. 3 and Tab. 2). The differences in the spatial distribution of wind
649 stress curl for simulations with coarse and fine surface vertical resolution sug-
650 gest the simulations with coarse resolution will exhibit relatively enhanced
651 wind-driven upwelling and downwelling in the Southern Ocean. The same
652 analysis performed in σ_2 -space (Fig. 4) suggests a more complicated reso-
653 lution dependence as the isopycnals are free to move relative to the wind
654 field. Nevertheless, for the ACCESS-OM2 simulations we expect enhanced
655 wind-driven upwelling on isopycnals denser than the zero wind stress curl σ_2
656 in simulations with coarse resolution, which differs from the expectation for
657 ACCESS-OM2-025, especially in the Indian and Atlantic basins. The simu-
658 lations themselves show that the models with coarse vertical resolution at the
659 ocean surface do indeed exhibit relatively stronger upwelling and downwelling
660 about the zero wind stress curl isopycnal. An interesting point to consider is
661 that as models continue to refine surface resolutions, and the ocean surface
662 speed approaches the wind speed, does the *Pacanowski* (1987) method for
663 calculating wind stress remain appropriate (Eqn. 1)? If not, should it begin
664 to employ a bulk upper ocean speed that is resolution-independent, such as
665 the ocean Ekman layer speed?

666 The purely thermal RCP45 perturbation imposed here acts to disrupt the
667 three-way balance between the surface momentum fluxes, surface heat fluxes
668 and ocean circulation. It initially warms the ocean surface, shifting the isopy-
669 cnal outcrops poleward beneath the fixed winds, which subsequently operate
670 on relatively less-dense waters. It follows that a purely hydrological pertur-
671 bation could have a similar effect: if the imposed hydrological perturbation
672 acted to freshen ocean surface and shift isopycnals to higher latitudes (as
673 per the high latitude surface freshening trend of *Durack & Wijffels*, 2010),

674 we should expect similar thermal responses as those presented here. This
675 expectation differs from the results of previous modelling efforts that impose
676 freshwater perturbations (*Armour et al.*, 2016; *Pauling et al.*, 2016), calling
677 for further investigations in this space.

678 Shifting the location of the zero wind stress curl to lighter waters has the
679 effect of weakening the wind-driven downwelling of adjacent lighter waters,
680 strengthening the wind-driven upwelling of adjacent denser waters, and re-
681 versing the wind-driven motions to now upwell waters between the old and
682 new zero wind stress curl isopycnals. Our findings here indicate a leading
683 order response to a purely thermal surface perturbation is the wind-driven
684 upwelling of deep waters immediately denser than the new zero wind stress
685 curl isopycnal. Considering that the deep ocean is a dominant carbon reser-
686 voir, this upwelling of deep waters could lead to a substantial outgassing of
687 CO₂ from the ocean to the atmosphere, potentially having dynamical conse-
688 quences for Earth’s climate.

689 The ocean–sea-ice simulations here employ the repeating wind field of
690 JRA55-RYF8485. This strategy allows us to clearly separate cause and ef-
691 fect in these simulations, but in reality, the winds will also shift as Earth’s
692 climate changes. Observations and climate models agree that winds over
693 the Southern Ocean are strengthening and shifting poleward. Alone, this
694 poleward movement of the wind field would shift the location of the zero
695 wind stress curl to a denser isopycnal. However, the atmosphere itself is
696 warming and increasing the longwave radiation to the ocean, and changes
697 in the hydrological cycle are freshening the ocean surface at high latitudes,
698 all of which tend to shift the zero wind stress curl to a less-dense isopycnal.
699 Presumably, additional degrees of freedom that enable the surface isopycnals
700 and wind field to shift relative to one another will likely broaden to pattern
701 of warming compared to that found here (e.g. *Marshall et al.*, 2015). So the
702 question now becomes one of competition between these opposing effects: in
703 isopycnal space, are the winds shifting to denser isopycnals faster than the
704 warming and freshening is reducing their densities?

705 The simulations here highlight the importance of the zero wind stress
706 curl for locating the regions of heat increase and decrease in the Indian and
707 Pacific Oceans. Considering the variability in ocean heat uptake and storage
708 already present in CMIP5 models, this finding provides a framework for fu-
709 ture investigations of heat uptake in climate and ocean models. Furthermore,
710 the sensitivity of the wind-driven response to the vertical resolution at the
711 ocean surface offers an additional means of delineating and understanding

712 heat biases amongst models; from our findings here we expect the models
 713 with relatively coarse resolution at the ocean surface to have enhanced wind
 714 stress and therefore enhanced wind-driven processes. This sensitivity to the
 715 vertical resolution at the ocean surface is novel and of particular interest since
 716 previous resolution sensitivity studies typically focus on the effects of hori-
 717 zontal resolution (e.g., *Spence et al.*, 2010; *Newsom et al.*, 2016). Given the
 718 expected refined horizontal resolution of the CMIP6 class of climate models,
 719 our results call for attention to the vertical resolution as well, especially at
 720 the ocean surface.

721 Our findings also provide guidance for investigating heat uptake in obser-
 722 vations. Following on from *Llovel & Terray* (2016), who identify shifts in the
 723 wind stress curl field associated with regions of heat uptake in observations,
 724 we link the changes of the wind stress curl to changes of the heat content in
 725 isopycnal space. *In situ* and gridded observational products of these fields
 726 are available, albeit from a range of observing platforms, making it possible
 727 to investigate this mechanism in observational data. This presents an avenue
 728 for future work on Southern Ocean heat uptake.

729 5. Conclusion

730 The simulations presented here were designed in part to examine the effect
 731 of model resolution on Southern Ocean heat uptake, predicated on hypothesis
 732 that spurious mixing may be altered at higher resolution. However, we found
 733 that the net rate of heat uptake depended only marginally on horizontal
 734 resolution. Instead, the biggest effect that resolution has involves the vertical
 735 resolution at the ocean surface, which influences the ocean surface speed and
 736 subsequently the wind stress exerted onto the ocean model. It follows that
 737 the Southern Ocean heat uptake itself is sensitive to the vertical resolution at
 738 the ocean surface; simulations with relatively coarse resolution at the surface
 739 (10m) exhibited Southern Ocean heat content changes at rates nearly double
 740 that of simulations with finer resolution at the surface (1m). The biggest
 741 effect of increasing the horizontal resolution is in the wind-driven upwelling.
 742 A consistent finding in all simulations is that peak upwelling and downwelling
 743 straddle the isopycnal of zero wind stress curl. These findings are of particular
 744 interest for understanding Southern Ocean heat uptake in climate model
 745 intercomparison projects with varying ocean resolutions, and for identifying
 746 ocean warming signals in observations.

747 **Acknowledgments**

748 Many thanks to M. England, R. Holmes and V. Tamsitt for very help-
749 ful discussions, and to the COSIMA consortium (www.cosima.org.au) for
750 making available the ACCESS-OM2 suite of models. COSIMA is supported
751 by an Australian Research Council Linkage Project (LP160100073). KDS
752 was supported by the Australian Government Department of the Environ-
753 ment through the National Environmental Science Program (NESP). The
754 numerical simulations and analysis were performed with the resources of the
755 National Computational Infrastructure (Canberra, Australia), which is sup-
756 ported by the Australian Commonwealth Government.

757 **Bibliography**

- 758 Armour, K.C., J. Marshall, J.R. Scott, A. Donohoe & E.R. Newsom (2016),
759 Southern Ocean warming delayed by circumpolar upwelling and equator-
760 ward transport, *Nature Geoscience*, *9*:549–554. DOI: 10.1038/NGEO2731
- 761 Bamber, J.L., M.R. van den Broeke, J. Ettema, J.T.M. Lenaerts & E. Rig-
762 not (2012), Recent large increases in freshwater fluxes from Greenland
763 into the North Atlantic, *Geophysical Research Letters*, *39*, L19501. DOI:
764 10.1029/2012GL052552
- 765 Depoorter, M.A., J.L. Bamber, J.A. Griggs, J.T.M. Lenaerts, S.R.M. Ligt-
766 berg, M.R. van den Broeke & G. Moholdt (2013), Calving fluxes and basal
767 melt rates of Antarctic ice shelves, *Nature*, *502*, 89–92. DOI: 10.1038/na-
768 ture12567
- 769 Doddridge, E.W., D.P. Marshall & A.McC. Hogg (2016), Eddy Cancellation
770 of the Ekman Cell in Subtropical Gyres, *Journal of Physical Oceanography*,
771 *46*, 2995–3010. DOI: 10.1175/JPO-D-16-0097.1
- 772 Durack, P.J. & S.E. Wijffels (2010), Fifty-Year Trends in Global Ocean Salin-
773 ities and Their Relationship to Broad-Scale Warming, *Journal of Climate*,
774 *23*:4342–4362. DOI: 10.1175/2010JCLI3377.1
- 775 Frölicher, T.L., J.L. Sarmiento, D.J. Paynter, J.P. Dunne, J.P. Krasting &
776 M. Winton (2015), Dominance of the Southern Ocean in anthropogenic
777 carbon and heat uptake in CMIP5 models, *Journal of Climate*, *28*:862–
778 886. DOI: 10.1175/JCLI-D-14-00117.1

- 779 Fyfe, J.C. & O.A. Saenko (2006), Simulated changes in Southern Hemi-
780 sphere winds and currents, *Geophysical Research Letters*, *33*, L01604. DOI:
781 10.1029/2005GL024464
- 782 Fyfe, J.C., O.A. Saenko, K. Zickfeld, M. Eby & A.J. Weaver (2007), The Role
783 of the Poleward-Intensifying Winds on Southern Ocean Warming, *Journal*
784 *of Climate*, *20*:5391–5400. DOI: 10.1175/2007JCLI1764.1
- 785 Gent, P.R. & J.C. McWilliams (1990), Isopycnal mixing in ocean circulation
786 models, *Journal of Physical Oceanography*, *20*:150–155.
- 787 Griffies, S.M. (2012), Elements of the Modular Ocean Model (MOM) 5 (2012
788 release with updates), Technical Report 7, *NOAA/Geophysical Fluid Dy-*
789 *namics Laboratory Ocean Group*.
- 790 Griffies, S.M., M. Winton, W.G. Anderson, R. Benson, T.L. Delworth, C.O.
791 Dufour, J.P. Dunne, P. Goddard, A.K. Morrison, A. Rosati, A.T. Witten-
792 berg, J. Yin, and R. Zhang (2015), Impacts on ocean heat from transient
793 mesoscale eddies in a hierarchy of climate models, *Journal of Climate*, *28*,
794 952–977. DOI: 10.1175/JCLI-D-14-00353.1
- 795 Griffies, S.M., G. Danabasoglu, P.J. Durack, A.J. Adcroft, V. Balaji, C.W.
796 Böning, E.P. Chassignet, E. Curchitser, J. Deshayes, H. Drange, B. Fox-
797 Kemper, P.J. Gleckler, J.M. Gregory, H. Haak, R.W. Hallberg, P. Heim-
798 back, H.T. Hewitt, D.M. Holland, T. Ilyina, J.H. Jungclaus, Y. Komuro,
799 J.P. Krasting, W.G. Large, S.J. Marsland, S. Masina, T.J. McDougall,
800 A.J.G. Nurser, J.C. Orr, A. Pirani, F. Qiao, R.J. Stouffer, K.E. Taylor,
801 A.M. Treguier, H. Tsujino, P. Uotila, M. Valdivieso, Q. Wang, M. Win-
802 ton & S.G. Yeager (2016), OMIP contribution to CMIP6: experimental
803 and diagnostic protocol for the physical component of the Ocean Model
804 Intercomparison Project, *Geoscientific Model Development*, *9*, 3231–3296.
805 DOI: 10.5194/gmd-9-3231-2016
- 806 Hallberg, R. (2013), Using a resolution function to regulate parameteriza-
807 tions of oceanic mesoscale eddy effects, *Ocean Modelling* *72*, 92–103. DOI:
808 10.1016/J.OCEMOD.2013.08.007
- 809 Huber, M.B. & L. Zanna (2017), Drivers of uncertainty in simulated ocean
810 circulation and heat uptake, *Geophysical Research Letters*, *44*. DOI:
811 10.1002/2017GL071587

- 812 Hunke, E., W. Lipscomb, P. Jones, A. Turner, N. Jeffery, &
813 S. Elliott (2015), CICE, The Los Alamos Sea Ice Model,
814 <https://www.osti.gov/servlets/purl/1364126>
- 815 Kiss, A.E., A.McC. Hogg, N. Hannah, F.B. Dias, G.B. Brassington, M.A.
816 Chamberlain, C. Chapman, P. Dobrohotoff, C. Domingues, E.R. Duran,
817 M.H. England, R. Fiedler, S.M. Griffies, A. Heerdegen, P. Heil, R.M.
818 Holmes, A. Klocker, S.J. Marsland, A.K. Morrison, J. Munroe, P.R. Oke,
819 M. Nikurashin, G.S. Pilo, O. Richet, A. Savita, P. Spence, K.D. Stewart,
820 M.L. Ward, F. Wu & X. Zhang (2019), ACCESS-OM2: A Global
821 Ocean-Sea Ice Model at Three Resolutions, *Geoscientific Model Development*
822 *Discussions*. DOI: 10.5194/gmd-2019-106
- 823 Kushner, P.J., I.M. Held & T.L. Delworth (2001), Southern Hemisphere
824 atmospheric circulation response to global warming, *Journal of Climate*,
825 *14*:2238–2249.
- 826 Large, W.G., J.C. McWilliams & S.C. Doney (1994), Oceanic vertical mixing:
827 A review and a model with a nonlocal boundary layer parameterization,
828 *Reviews of Geophysics*, *32*:363–403.
- 829 Levitus, S., J.I. Antonov, T.P. Boyer, O.K. Baranova, H.E. Garcia, R.A. Locarnini,
830 A.V. Mishonov, J.R. Reagan, D. Seidov, E.S. Yarosh & M.M.
831 Zweng (2012), World ocean heat content and thermosteric sea level
832 change (0–2000/m), 1955–2010, *Geophysical Research Letters*, *39*(10).
833 DOI: 10.1029/2012GL051106
- 834 Lui, W., J. Lu, S.-P. Xie & A. Fedorov (2018), Southern Ocean Heat Uptake,
835 Redistribution, and Storage in a Warming Climate: The Role of Meridional
836 Overturning Circulation, *Journal of Climate*, *31*:4727–4743. DOI:
837 10.1175/JCLI-D-17-0761.1
- 838 Llovel, W. & L. Terray (2016), Observed southern upper-ocean warming over
839 2005–2014 and associated mechanisms, *Environmental Research Letters*,
840 *11*:124023. DOI: 10.1088/1748-9326/11/12/124023
- 841 Locarnini, R.A., A.V. Mishonov, J.I. Antonov, T.P. Boyer, H.E. Garcia,
842 O.K. Baranova, M.M. Zweng, C.R. Paver, J.R. Reagan, D.R. Johnson,
843 M. Hamilton & D. Seidov (2013), World Ocean Atlas 2013, Volume 1:

- 844 Temperature. S. Levitus (Ed.), A. Mishonov (Technical Ed.); *NOAA Atlas*
845 *NESDIS*, 73, 40pp.
- 846 Marshall, G.J. (2003), Trends in the Southern Annular Mode from observa-
847 tions and reanalysis, *Journal of Climate*, 16:4234–4143.
- 848 Marshall, J., J.R. Scott, K.C. Armour, J.-M. Campin, M. Kelley & A. Ro-
849 manou (2015), The ocean’s role in the transient response of climate to
850 abrupt greenhouse gas forcing, *Climate Dynamics*, 44:2287–2299. DOI:
851 10.1007/s00382-014-2308-0
- 852 Marsland, S.J., D. Bi, P. Uotila, R. Fiedler, S.M. Griffies, K. Lorabacher,
853 S. O’Farrell, A. Sullivan, P. Uhe, X. Zhou & A.C. Hirst (2013), Evalua-
854 tion of ACCESS climate model ocean diagnostics in CMIP5 simulations,
855 *Australian Meteorological and Oceanographic Journal*, 63, 101–119.
- 856 Morrison, A.K., O.A. Saenko, A.McC. Hogg & P. Spence (2013), The role of
857 vertical eddy flux in Southern Ocean heat uptake, *Geophysical Research*
858 *Letters*, 40:5445–5450. DOI: 10.1002/2013GL057706
- 859 Morrison, A.K., S.M. Griffies, M. Winton, W.G. Anderson & J.L. Sarmiento
860 (2016), Mechanisms of Southern Ocean heat uptake and transport in a
861 global eddying climate model, *Journal of Climate*, 29:2059–2075. DOI:
862 10.1175/JCLI-D-15-0579.1
- 863 Newsom, E.R., C.M. Bitz, F.O. Bryan, R. Abernathey & P.R. Gent (2016),
864 Southern Ocean Deep Circulation and Heat Uptake in a High-Resolution
865 Climate Model, *Journal of Climate*, 29, 2597–2619. DOI: 10.1175/JCLI-
866 D-15-0513.1
- 867 Oke, P.R. & M.H. England (2004), Oceanic response to changes in the lat-
868 itude of the Southern Hemisphere subpolar westerly winds, *Journal of*
869 *Climate*, 17:1040–1054.
- 870 Pacanowski, R.C. (1987), Effect of Equatorial Currents on Surface Stress,
871 *Journal of Physical Oceanography*, 17:833–838.
- 872 Pauling, A.G., C.M. Bitz, I.J. Smith & P.J. Langhorne (2016), The Re-
873 sponse of the Southern Ocean and Antarctic Sea Ice to Freshwater from
874 Ice Shelves in an Earth System Model, *Journal of Climate*, 29:1655–1672.
875 DOI: 10.1175/JCLI-D-15-0501.1

- 876 Redi, M.H. (1982), Oceanic isopycnal mixing by coordinate rotation, *Journal*
877 *of Physical Oceanography*, *12*:1154–1158.
- 878 Roemmich, D., J. Church, J. Gilson, D. Monselesan, P. Sutton & S. Wijffels
879 (2015), Unabated planetary warming and its ocean structure since 2006,
880 *Nature Climate Change*, *5*(3):240–245, 50–65. DOI: 10.1038/nclimate2513
- 881 Sallée, J.-B., E. Shuckburgh, N. Bruneau, A.J.S. Meijers, T.J. Bracegirdle,
882 Z. Wang & T. Roy (2013), Assessment of Southern Ocean water mass
883 circulation and characteristics in CMIP5 models: Historical bias and forc-
884 ing response, *Journal of Geophysical Research*, *118*(4):1830–1844. DOI:
885 10.1002/jgrc.20135
- 886 Sallée, J.-B. (2018), Southern Ocean warming, *Oceanography*, *32*(2):52–62.
887 DOI: 10.5670/oceanog.2018.215
- 888 Snow, K., A.McC. Hogg, B.M. Sloyan, & S.M. Downes (2016), Sensitivity of
889 Antarctic Bottom Water to Changes in Surface Buoyancy Fluxes, *Journal*
890 *of Climate*, *29*, 313–330. DOI: 10.1175/JCLI-D-15-0467.1
- 891 Spence, P., J.C. Fyfe, A. Monrenegro & A.J. Weaver (2010), Southern Ocean
892 Response to Strengthening Winds in an Eddy-Permitting Global Climate
893 Model, *Journal of Climate*, *23*, 5332–5343. DOI: 10.1175/2010JCLI3098.1
- 894 Stewart, K.D., A.McC. Hogg, S.M. Griffies, A.P. Heerdegen, M.L. Ward,
895 P. Spence & M.H. England (2017), Vertical resolution of baroclinic
896 modes in global ocean models, *Ocean Modelling*, *113*, 50–65. DOI:
897 10.1016/j.ocemod.2017.03.012
- 898 Stewart, K.D., W. Kim, S. Urakawa, A.McC. Hogg, S. Yeager, H. Tsujino, H.
899 Nakano, A.E. Kiss & G. Danabasoglu (in prep.), JRA55-do Repeat Year
900 Forcing, *Ocean Modelling*.
- 901 Swart, N.C. & J.C. Fyfe (2012), Observed and simulated changes in the
902 Southern Hemisphere surface westerly wind-stress, *Geophysical Research*
903 *Letters*, *39*, L16711. DOI: 10.1029/2012GL052810
- 904 Thompson, D.W.J., J.M. Wallace & G.C. Hegerl (2000), Annular modes in
905 the extratropical circulation. Part II: Trends, *Journal of Climate*, *13*:1018–
906 1036.

- 907 Tsujino, H., S. Urakawa, H. Nakano, R.J. Small, W.M. Kim, S.G. Yeager,
908 G. Danabasoglu, T. Suzuki, J.L. Bamber, M. Bentsen, C.W. Böning,
909 A. Bozec, E.P. Chassignet, E. Curchitser, F.B. Dias, P.J. Durack, S.M.
910 Griffies, Y. Harada, M. Ilıcak, S.A. Josey, C. Kobayashi, S. Kobayashi,
911 Y. Komuro, W.G. Large, J. Le Sommer, S.J. Marsland, S. Masina, M.
912 Scheinert, H. Tomita, M. Valdivieso & D. Yamazaki (2018), JRA-55 based
913 surface dataset for driving ocean–sea-ice models (JRA55-do), *Ocean Mod-*
914 *elling*, *130*, 79–139. DOI: 10.1016/j.ocemod.2018.07.002
- 915 Wolfe, C.L. & P. Cessi (2011), The Adiabatic Pole-to-Pole Overturn-
916 ing Circulation, *Journal of Physical Oceanography*, *41*:1795–1810. DOI:
917 10.1175/2011JPO4570.1
- 918 Xie, P. & G.K. Vallis (2011), The passive and active nature of ocean heat up-
919 take in idealized climate change experiments, *Climate Dynamics*, *38*:667–
920 684. DOI: 10.1007/s00382-011-1063-8
- 921 Zweng, M.M., J.R. Reagan, J.I. Antonov, R.A. Locarnini, A.V. Mishonov,
922 T.P. Boyer, H.E. Garcia, O.K. Baranova, D.R. Johnson, D. Seidov & M.M.
923 Biddle (2013), World Ocean Atlas 2013, Volume 2: Salinity. S. Levitus
924 (Ed.), A. Mishonov (Technical Ed.); *NOAA Atlas NESDIS*, *74*, 39pp.