



**HAL**  
open science

# Combined physical and biogeochemical assessment of mesoscale eddy parameterisations in ocean models: eddy induced advection at non-eddy resolving resolutions

X. Ruan, Damien Couespel, Marina Lévy, J. Li, J. Mak, Y. Wang

## ► To cite this version:

X. Ruan, Damien Couespel, Marina Lévy, J. Li, J. Mak, et al.. Combined physical and biogeochemical assessment of mesoscale eddy parameterisations in ocean models: eddy induced advection at non-eddy resolving resolutions. *Ocean Modelling*, 2023, 183, pp.102204. 10.1016/j.ocemod.2023.102204 . hal-04089164

**HAL Id: hal-04089164**

**<https://hal.science/hal-04089164>**

Submitted on 4 May 2023

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

# Combined physical and biogeochemical assessment of mesoscale eddy parameterisations in ocean models: eddy induced advection at non-eddy resolving resolutions

X. Ruan<sup>a,b,1</sup>, D. Couespel<sup>c,d</sup>, M. Lévy<sup>d</sup>, J. Li<sup>a,b</sup>, J. Mak<sup>a,b,1</sup>, Y. Wang<sup>a,b</sup>

<sup>a</sup>Department of Ocean Science, Hong Kong University of Science and Technology

<sup>b</sup>Southern Marine Science and Engineering Guangdong Laboratory (Guangzhou)

<sup>c</sup>NORCE Norwegian Research Centre, Bjerknes Centre for Climate Research

<sup>d</sup>Sorbonne Université, LOCEAN-IPSL, CNRS/IRD/MNH

---

## Abstract

Ocean components of Earth System Models employed for climate projections do not routinely resolve mesoscale eddies for computational cost reasons, and the associated subgrid processes are still parameterised. While the performance of physics parameterisations in a numerical ocean model is normally assessed by examining the associated physical responses, biogeochemical responses are also important but often treated separately. Given recent advances in mesoscale eddy parameterisations, specifically for the eddy induced advection, this work systematically explores the joint consequences for physical as well as biogeochemical responses brought about by a more updated proposal for the eddy induced velocity coefficient, in the context of an idealised ocean relevant model. Relative to a high resolution mesoscale eddy resolving model, the more updated mesoscale eddy parameterisation is able to capture aspects of the model truth in the physical responses. The biogeochemistry response is however rather more subtle, where a ‘better’ response with the conventional eddy parameterisation with a constant coefficient could arise from a physically inconsistent response, while a parameterisation that improves the bulk physical response may still fall short in its biogeochemical response. The present work highlights a need to assess both physical and biogeochemical aspects when judging the performance of eddy parameterisations, and additionally provides some important baseline model sensitivities that future assessments employing other parameterisations or in more complex settings could compare against.

*Keywords:* mesoscale eddy parameterisation, numerical modelling, climate change, biophysical interactions

---

## 1. Introduction

The ocean circulation plays a crucial role in the Earth system’s heat, carbon and nutrient cycles, and affects the global climate and the marine ecosystem (e.g. Rahmstorf, 2002; Doney et al., 2012). Over the decadal to centennial time-scales, more heat is expected to reside in the upper part of the ocean under climate projection exercises (e.g. IPCC, 2019), strengthening the upper ocean stratification and changing the ocean ventilation pathways (e.g., Bindoff and McDougall, 1994; Li et al., 2020). The ocean meridional overturning circulation is projected to slow down, partly via the shoaling of the pycnocline, though uncertainties still exist (e.g., Bellomo et al., 2021). Changes in the ocean overturning circulation can affect the bulk transport of nutrients, which can then have large-scale impacts on the phytoplankton populations. As primary producers, phytoplankton play an important role in the global carbon cycle and impact issues of food security via their position at the base of most oceanic food webs. While there is large uncertainty in the physiological responses of various marine biomass to the changing marine environment in terms of heat stress, nutrient abundance, acidity and others (e.g., Kwiatkowski et al., 2020; Tagliabue et al., 2021; Martiny

---

*Email addresses:* xruanaa@connect.ust.hk (X. Ruan), julian.c.l.mak@gmail.com (J. Mak)

<sup>1</sup>Corresponding authors.

13 et al., 2022), it is not controversial to say that the physical circulation can impact the broad regional and global  
14 biogeochemical response. One such link is the impact of the circulation on nutrient supply, though such projections  
15 often come with large uncertainties given the nonlinear interactions present in the complex Earth system (e.g., Lotze  
16 et al., 2019).

17 Earth System Models are invaluable tools for probing and constraining the physical and biogeochemical responses  
18 in the marine system to the changing environment. These numerical models simulate the evolution of the Earth system  
19 components and their interactions (e.g., Bonan and Doney, 2018; Séférian et al., 2019; Lee et al., 2022), with the  
20 assumption that the processes implemented into the numerical models are correct. However, even with the increasing  
21 computational power available, present state-of-the-art Earth System Models still mostly utilise ocean components at  
22 approximately 1° horizontal resolution that do not explicitly permit geostrophic mesoscale eddies (e.g., Hewitt et al.,  
23 2020, 2022). Geostrophic mesoscale eddies play a crucial role in regulating the ocean circulation responses, which  
24 not only have local effects, but also impact the larger scale regional and global mean state (e.g. Lévy et al., 2012).  
25 Although there is an increasing push for ocean models to be mesoscale eddy resolving (Fox-Kemper et al., 2019;  
26 Kwiatkowski et al., 2020; Hewitt et al., 2022) or at least eddy rich (at around 1/12° horizontal resolutions, Hallberg,  
27 2013), models at such resolutions remain computationally prohibitive, and global Earth System Models with an ocean  
28 component at the mesoscale eddy permitting regime at around 1/4° horizontal resolution is a more realistic target  
29 (Hewitt et al., 2017, 2020, 2022; Roberts et al., 2020). Given the physical influence on the biogeochemical response  
30 and the anticipated developments of ocean models over the next decade, there is a need to probe, constrain and  
31 understand the sensitivities of the physical and biogeochemical responses in ocean models at the non-eddy resolving,  
32 eddy permitting as well as the eddy rich/resolving resolutions.

33 For those models that do not explicitly permit mesoscale eddies, parameterisations are often employed to mimic  
34 the feedback of geostrophic mesoscale eddies. Often employed are what we would term here as *diffusive* closures,  
35 such as isoneutral diffusion (e.g., Redi, 1982; Griffies, 1998) and the Gent–McWilliams (GM) scheme (Gent and  
36 McWilliams, 1990; Gent et al., 1995). Isonneutral diffusion leads to tracer diffusion along the isoneutral directions,  
37 while the GM scheme leads to an eddy induced advection of tracers that flattens isoneutral slopes, and both are  
38 consistent with the adiabatic nature of mesoscale eddies generated by baroclinic instabilities (e.g., Vallis, 2006). Such  
39 diffusive closures were designed for coarse resolution models with no explicitly resolved eddies, considered more  
40 standard, and variants of such schemes exist in most numerical ocean models (e.g. MITgcm, Marshall et al. 1997a,b;  
41 NEMO, Madec 2008; FESOM, Wang et al. 2014; MOM, Adcroft et al. 2019). On the other hand, mesoscale eddies can  
42 also lead to sharpening of large-scale jets (via inverse cascades, eddy induced momentum convergence, or otherwise,  
43 e.g. Waterman and Jayne, 2012; Waterman and Hoskins, 2013), which is increasingly modelled by *backscatter* based  
44 parameterisations (e.g., Bachman, 2019; Jansen et al., 2019). Recent advances in both classes of parameterisations  
45 have led to lower resolution models that are more in line with the eddy rich/resolving models at least in the physical  
46 response. Advances in diffusive schemes tend to focus more on coarse resolution models, some of which have led to  
47 improvements in sensitivities of the circulation to changing forcing scenarios (e.g., Farneti et al., 2015; Mak et al.,  
48 2018, 2022b). Backscatter schemes have received more attention in eddy permitting models because of their ability  
49 to strengthen the represented eddy energy levels and ocean currents (e.g., Bachman, 2019; Jansen et al., 2019).

50 The biogeochemical response to such recent updates in physics parameterisations have been, on the other hand,  
51 lacking when it is known that model represented physics can have a very substantial impact on the resulting physical  
52 and/or biogeochemical metrics of interest. Modifying the represented eddy-mean feedbacks can have a significant  
53 effect on the ventilation rate and pathways, affecting the represented ocean heat content (e.g., de Boer et al., 2007;  
54 Zhang and Vallis, 2013; Zanna et al., 2019b; Mak et al., 2022b; Newsom et al., 2022), carbon (England and Rahmstorf,  
55 1999; Gnanadesikan et al., 2015; Khwatiwala et al., 2018), oxygen (Matear et al., 2000; Helm et al., 2011; Bopp  
56 et al., 2017; Takano et al., 2018), and nutrient distributions (Lévy et al., 1999; Tschumi et al., 2011; Bopp et al.,  
57 2013; Couespel et al., 2021). With the prevalent use of numerical ocean general circulation models for probing and  
58 predicting biophysical interactions (e.g., Bopp et al., 2013; Berthet et al., 2019; Swearer et al., 2019; Séférian et al.,  
59 2019), it is important to investigate how the physics parameterisations (i) modify the modelled physical states, and (ii)  
60 affect the biogeochemical responses. Such an investigation is required since there is no guarantee that improvements  
61 in physical processes necessarily lead to a ‘better’ biogeochemical response, given the nonlinear interactions inherent  
62 in a complex system.

63 Global and/or realistic models, while useful for making predictions and informing policies (e.g., IPBES, 2019), are  
64 computationally expensive and possess a large number of degrees of freedom, making it difficult to attribute the various

causalities. While ultimately these realistic and complex Earth System Models should be used when quantitatively assessing the impacts of eddy parameterisations, for delineating the causality and interactions between the physical parameterisations and the resulting physical and biogeochemical responses, we consider here a complementary approach by utilising idealised numerical models, focusing on the qualitative differences arising from the choice of eddy parameterisations. We focus on a systematic assessment of mesoscale eddy parameterisations and their qualitative impact on the nutrient stream or relay, and their subsequent impact on Net Primary Production (NPP) (e.g., Williams et al., 2017, 2011; Whitt and Jansen, 2020; Gupta et al., 2022). We employ a double gyre setting with a simple biogeochemistry model, with prescribed atmospheric forcing and an idealised climate change scenario (Couespel et al., 2021). The double gyre setting has the benefit that the model behaviours and limitations are relatively well-known (e.g., Jackson et al., 2006; Lévy et al., 2010, 2012, 2014; Stewart et al., 2021), and the high resolution eddy resolving ‘model truths’ are more computationally accessible because of the limited spatial extent.

Even with the reduced complexity afforded by the choice of numerical model, there are multiple parameterisations for mesoscale turbulence. Here we focus on diffusive eddy closures, specifically on the GM-based parameterisations for the eddy induced advection; an analogous investigation into the effects of isoneutral diffusion, backscatter type eddy parameterisations, and extensions into the eddy permitting models will be reported in subsequent publications. The models to be investigated here are non-eddy resolving, differing by the GM-type closures they employ, and the qualitative performance of these will be judged against the high resolution eddy resolving model truths. The GM parameterisation variants and the numerical model set up are described in §2. In §3 we report the qualitative differences in both the physical and biogeochemical responses arising from the choice of closures. In §4 we subject the models to idealised climate change scenarios to investigate analogous model sensitivities. The article concludes in §5, critically evaluating the advantages and shortfalls provided by the choices of GM-based closures.

## 2. Mesoscale eddy parameterisations and numerical set up

Two canonical types of diffusive closures associated with geostrophic turbulence are those based on isoneutral diffusion (e.g., Redi, 1982) and the Gent–McWilliams scheme (GM, Gent and McWilliams, 1990). The former refers to diffusion of tracers along the isoneutral direction, while the latter is an eddy induced advection (e.g., Gent et al., 1995; Treguier et al., 1997; Griffies, 1998) although it resembles a horizontal buoyancy diffusion (in the quasi-geostrophic limit, e.g. Treguier et al., 1997) or a layer thickness diffusion (e.g., Gent and McWilliams, 1990). The isoneutral diffusion and GM schemes are both known to affect the physical and biogeochemical response. Isoneutral diffusion modifies the rate of tracer ventilation, and the GM schemes affect the structure of the tracer ventilation through its impact on the density stratification. Relatively speaking, there are more studies on assessing GM-based schemes (e.g., Visbeck et al., 1997; Eden and Greatbatch, 2008; Cessi, 2008; Hofman and Morales Maqueda, 2011; Munday et al., 2013; Zhang and Vallis, 2013; Bates et al., 2014; Farneti et al., 2015; Mak et al., 2018, 2022b), although there have also been increasing interest in isoneutral diffusion, assessing its impact as well as improving on the standard implementation with constant diffusivity (e.g., Ferrari and Nikurashin, 2010; Pradal and Gnanadesikan, 2014; Jones and Abernathy, 2019, 2021; Groeskamp et al., 2021; Holmes et al., 2022; Chouksey et al., 2022). While both processes are related to mesoscale turbulence, and there are works that suggest relationships between the two (e.g., Smith and Marshall, 2009; Abernathy et al., 2013), owing to the larger interest in GM-based closures, in this work we focus primarily on the consequences afforded by different GM-based schemes, and consider a prescribed constant isoneutral diffusivity  $\kappa_{\text{iso}}$ . The model sensitivity to  $\kappa_{\text{iso}}$  by itself was found to be rather mild in the present model, although nonlinear feedback loops can be present, suggesting that further investigation is required in this area; see Appendix A for details.

### 2.1. GM-based parameterisations

The GM-scheme introduces an eddy induced velocity  $\mathbf{u}^*$  to the tracer advection equation (e.g., Griffies, 1998; Ferreira et al., 2005):

$$\mathbf{u}^* = -\nabla \times (\kappa_{\text{gm}} \mathbf{s}). \quad (1)$$

Here,  $\mathbf{s} = -\nabla_H \rho / N^2$  the isopycnal slope in the horizontal directions,  $\nabla_H$  the horizontal gradient operator,  $N^2 \sim -\partial \rho / \partial z$  the vertical buoyancy gradients associated with the resolved state,  $\rho$  denotes the dynamically relevant density, and  $\kappa_{\text{gm}}$  will be termed the GM-coefficient in this work. The GM scheme is widely used because of its inherent properties,

112 such as adiabatic advection leading to slumping of isopycnals, positive-definite generation of eddy energy and layer-  
 113 wise conservation of moments (e.g., Gent et al., 1995), as well as numerical advantages (numerical stability due to  
 114 the slumping action, reduction of unrealistic deep convection; e.g., Danabasoglu et al. 1994). A choice often utilised  
 115 in idealised models takes the simple prescription of

$$\kappa_{\text{gm}} = \kappa_0 = \text{constant.} \quad (2)$$

### 116 2.1.1. GEOMETRIC

117 As it is desirable to maintain the properties afforded by the GM scheme even if one does not believe a simple  
 118 prescription of  $\kappa_{\text{gm}} = \text{constant}$  will suffice, a prevalent research focus has been on improving on the functional form  
 119 of  $\kappa_{\text{gm}}$  (e.g., Visbeck et al., 1997; Treguier et al., 1997; Ferreira et al., 2005; Cessi, 2008; Eden and Greatbatch,  
 120 2008; Hofman and Morales Maqueda, 2011; Marshall et al., 2012; Jansen et al., 2015, 2019). We consider here  
 121 on a form of the GM scheme arising from the GEOMETRIC framework for analysing eddy-mean flow interactions  
 122 (see Marshall et al., 2012; Maddison and Marshall, 2013). In the present form, analysis within the GEOMETRIC  
 123 framework suggests scaling  $\kappa_{\text{gm}}$  as (Mak et al., 2018, 2022b)

$$\kappa_{\text{gm}} = \alpha \frac{\int E \, dz}{\int (M^2/N) \, dz}, \quad (3)$$

124 where  $M^2 \sim |\nabla_H \rho|$  denotes the horizontal buoyancy gradients associated with the resolved state,  $\alpha$  is a non-  
 125 dimensional tuning parameter (bounded in magnitude by one),  $E$  is the total (potential and kinetic) eddy energy,  
 126 and the resulting  $\kappa_{\text{gm}}$  varies in time and in the horizontal (but is depth-independent with the present specification).  
 127 Unlike most other existing proposals for the  $\kappa_{\text{gm}}$  that utilise mixing length type arguments with dependence on the  
 128 eddy kinetic energy, the GM-version of GEOMETRIC arises from a mathematically rigorous bound that results from  
 129 analysing the Eliassen–Palm flux tensor that encodes the eddy-mean feedbacks (Marshall et al., 2012; Maddison  
 130 and Marshall, 2013). The bound results in a linear dependence on the total eddy energy  $E$  (compared to mixing  
 131 length based parameterisations with a square root scaling; e.g., Eden and Greatbatch 2008; Jansen et al. 2015, 2019),  
 132 which leads to a more significant state-dependent response. Notably, out of the GM-based parameterisations, the  
 133 GM-version of GEOMETRIC has more evidence in support of its use, from a diagnostic point of view (Bachman  
 134 et al., 2017; Wang and Stewart, 2020; Wei et al., 2022), and prognostic calculations in idealised models (Mak et al.,  
 135 2017, 2018) as well as in realistic models (Mak et al., 2022b). In particular, the GM-version of GEOMETRIC in  
 136 the aforementioned prognostic calculations have been shown to lead to improved sensitivities of the modelled ocean  
 137 circulation to changes in forcing over the standard prescriptions of the  $\kappa_{\text{gm}}$ , notably in the Antarctic Circumpolar  
 138 Current transport and the global Meridional Overturning Circulation strength.

139 In a prognostic calculation with a coarse resolution model,  $E$  is provided using a depth-integrated eddy energy  
 140 budget. Denoting  $(x, y)$  to be the zonal and meridional directions respectively, following Mak et al. (2022b), the eddy  
 141 energy budget used with GEOMETRIC is given by

$$\frac{d}{dt} \int E \, dz + \underbrace{\nabla_H \cdot \left( \widetilde{\mathbf{u}}^z - |c| \mathbf{e}_x \right) \int E \, dz}_{\text{advection}} = \underbrace{\int \kappa_{\text{gm}} \frac{M^4}{N^2} \, dz}_{\text{source}} - \underbrace{\lambda \int (E - E_0) \, dz}_{\text{dissipation}} + \underbrace{\eta_E \nabla_H^2 \int E \, dz}_{\text{diffusion}}, \quad (4)$$

142 where the depth-integrated eddy energy is advected by the depth average flow  $\widetilde{\mathbf{u}}^z$  with westward propagation at the  
 143 long Rossby wave phase speed  $|c|$  (e.g., Chelton et al., 2011; Klocker and Marshall, 2014). The growth of eddy  
 144 energy comes from the slumping of mean density surfaces, and diffused in the horizontal (Grooms, 2015; Ni et al.,  
 145 2020a,b) with  $\eta_E$  denoting the associated eddy energy diffusivity. A linear dissipation of eddy energy at rate  $\lambda$  (but  
 146 maintaining a minimum eddy energy level  $E_0$ ) is utilized, so  $\lambda^{-1}$  is an eddy energy dissipation time-scale, which is a  
 147 bulk parameterisation of energy fluxes out of the mesoscales resulting from numerous dynamical processes (e.g., Mak  
 148 et al., 2022a).

149 In this work we focus on a comparison between calculations employing the GM-version of the GEOMETRIC  
 150 parameterisation (denoted GEOM), to be compared with the calculations employing a prescribed constant  $\kappa_{\text{gm}}$   
 151 (denoted CONST); we have also performed calculations with simpler proposals of  $\kappa_{\text{gm}}$  that are state dependent



152 (Treguier et al. 1997; cf. Visbeck et al. 1997), and will comment on the results from those calculations where  
153 appropriate. Although we have not performed calculations employing other existing energetically constrained  
154 proposals (e.g.,  $\kappa_{gm} \sim \sqrt{K}$  of Jansen et al., 2019, where  $K$  is the eddy kinetic energy), we speculate in the discussion  
155 section the expected responses given the results from in the present work.

## 156 2.2. Model set up

157 Our main focus is to systematically assess the qualitative differences arising from different eddy parameterisation  
158 variants, and for this purpose an idealised numerical ocean model is employed. A double gyre model based on the set  
159 up of Couespel et al. (2021) using the Nucleus for European Modelling of the Ocean (NEMO; Madec 2008) was used.  
160 The gyre model here is a “straightened” version of the standard gyre configuration test case that comes with NEMO.  
161 The model is already coupled to an idealised biogeochemistry model within NEMO (cf. Lévy et al., 2010, 2012)  
162 and has been used to study both physical and biogeochemical responses in the aforementioned work. To recap, the  
163 domain is square with sides of length 3180 km and depth 4 km, formulated on a  $\beta$ -plane with the Southern boundary  
164 at  $20^\circ$  N, extending to the Northern boundary at  $50^\circ$  N. The domain has no bathymetry, and is bounded by vertical  
165 walls that are aligned with longitudes and latitudes on all sides. While the presence of bathymetry is known to have  
166 impacts on the large-scale circulation (e.g., Jackson et al., 2006; Gula et al., 2015; Stewart et al., 2021), there is the  
167 added subtlety on how one should parameterise eddy feedbacks over slope regions (e.g., Wang and Stewart, 2020; Wei  
168 et al., 2022). For simplicity and to reduce the degrees of freedom in the problem, we opted for the flat bottom case.  
169 We employ a non-linear bottom drag, and impose free-slip conditions on the lateral boundaries. The model utilises  
170 a linear equation of state with temperature and salinity, and vertical mixing is via a turbulent kinetic energy scheme  
171 (Gaspar et al., 1990). Atmospheric forcing is through the flux formulation, and the forcings (wind stress, penetrative  
172 solar radiation, pseudo-atmospheric temperature  $\Theta^*$  for computing sea surface temperature restoring, freshwater flux)  
173 are all zonally symmetric and with a prescribed repeating seasonal cycle and no period beyond a year, and there is no  
174 net salinity flux (see Lévy et al., 2010, Fig.1).

175 The model employs an idealised biogeochemistry model LOBSTER (see e.g., Lévy et al., 2012) with standard  
176 reference settings, and with eddy induced advective and diffusive contributions from the GM-based and Redi  
177 schemes respectively. The LOBSTER model uses nitrogen as the currency, and the six biogeochemical variables are  
178 concentrations of detritus, zooplankton, phytoplankton, nitrate, ammonium, and dissolved organic matter; variables of  
179 particular interest to the work here are phytoplankton and nitrate for their links to Net Primary Production (NPP). As  
180 LOBSTER does not represent physiological changes with changes in temperature, changes observed are solely due  
181 to changes in the transport of the biogeochemical tracers, providing a better focus on the large-scale links between  
182 physics and biogeochemistry.

183 The differences between the model employed in this work to that of Couespel et al. (2021) are the following:

- 184 • version of NEMO (NEMO v4.0.5 r14538 instead of v3.4 r4826),
- 185 • a slightly different initialisation of nitrate concentration at the start of the perturbation experiments (no averaging  
186 of the deep ocean nitrate concentration),
- 187 • the model truth is taken here to have a horizontal resolution of  $1/12^\circ$  instead of  $1/9^\circ$ .

188 The updated version of NEMO already has the GM-version of GEOMETRIC implemented from the work of Mak  
189 et al. (2022b), and it was easier to adapt the model configuration to the newer NEMO than to write the GEOMETRIC  
190 parameterisation into an older version of NEMO. Sample calculations show that the different initialisation of the  
191 nitrate concentration at the deeper ocean have no impact on the conclusions in this article. The horizontal resolution  
192 of the model truth was increased to  $1/12^\circ$  partly as a balance to resolve mesoscale processes, but without resolving  
193 too much of the submesoscale processes, so that there is a more suitable comparison between the model truth and the  
194 coarse resolution models employing the GM-based schemes, since the GM-based schemes are not designed to capture  
195 submesoscale processes. The model truth horizontal resolution of  $1/12^\circ$  was also chosen for the suggestive analogy  
196 with the global NEMO ORCA0083 (also known as ORCA12) configuration that is at a nominal horizontal resolution  
197 of  $1/12^\circ$ , and provides a benchmark reference for our future investigations into eddy permitting models. All the major  
198 conclusions of Couespel et al. (2021) are found to hold even with the present changes. A summary of key model  
199 parameters is given in Table 1, partly informed by previous works (Couespel et al., 2021; Mak et al., 2022b,a). A brief  
200 description of model sensitivities to some of these choices are given in Appendix A.

	R1 (CONST and GEOM)	R12
horizontal resolution	106 km	8.83 km
time step	30 mins	10 mins
momentum diffusion	horizontal $\nabla^2$ , $\nu = 10^5 \text{ m}^2 \text{ s}^{-1}$	horizontal $\nabla^4$ , $\nu = -3 \times 10^{10} \text{ m}^4 \text{ s}^{-1}$
tracer advection	FCT scheme	MUSCL scheme
tracer diffusion	isopycnal $\nabla^2$ , $\kappa_{\text{iso}} = 10^3 \text{ m}^2 \text{ s}^{-1}$	iso-level $\nabla^4$ , $\kappa = -10^9 \text{ m}^4 \text{ s}^{-1}$
eddy induced advection	CONST ( $\kappa_{\text{gm}} = 1000 \text{ m}^2 \text{ s}^{-1}$ ) GEOM ( $\alpha = 0.04$ , $\lambda^{-1} = 135 \text{ days}$ )	— —

Table 1: Key model parameter differences between the calculations considered in this work.

### 2.3. Experimental set up

Following the strategy of Couespel et al. (2021), the physical and biogeochemical model at the  $1^\circ$  resolution starts from model year  $-2300$ , spun up over 2000 model years to model year  $-300$  using the CONST variant with constant  $\kappa_{\text{gm}}$ . At model year  $-300$ , perturbation experiments were carried out for another 300 years to model year 0 (which is longer than the 100 years considered in Couespel et al. 2021). For the  $1^\circ$  models, the perturbation experiments are with CONST and GEOM. For the  $1/12^\circ$  model R12, the fields are simply interpolated from the  $1^\circ$  model onto the analogous  $1/12^\circ$  grid

Two sets of experiments are performed in this work. A control pre-industrial setting (tagged with a suffix CTL) integrates the aforementioned calculations for another 70 years from model year 0, subject to the same idealised atmospheric forcing. An idealised climate change scenario (tagged with a suffix CC) has the models exposed to the aforementioned idealised seasonal cycle from model year 0, but the atmospheric temperature is given an increasing linear trend of  $+0.04 \text{ }^\circ\text{C yr}^{-1}$  over 70 model years, following Couespel et al. (2021) to mimic the SSP5-8.5 scenario (e.g., Tokarska et al., 2020). Fig. 1 shows some of the summary statistics of the spinup and the adjustment under the control and idealised climate change scenario, indicating that a quasi-equilibrium has been reached at least in the upper parts of the ocean (depths less than 700 m) in the perturbation calculations during the spinup stage. Under the idealised climate change scenario, the ocean temperature increases, leading to a stronger stratification (primarily in the upper ocean; not shown) that inhibits nutrient supply and a decrease in NPP across the set of calculations, consistent with the results of Couespel et al. (2021) (see their Fig. 1d and A1).

In each of the two sets of experiments we evaluate the performance of the eddy parameterisations by examining both the physical and biogeochemical responses, critically comparing the similarities and differences between GEOM, CONST and the R12 model truth, under the CTL and CC scenario. A working hypothesis is that the physical and biogeochemical responses are improved in the GEOM calculations when compared to the CONST calculations. For evaluating the performance, we consider diagnostics calculated from data time averaged over the analysis period, taken to be the last five years of the calculation (between the start of model year 66 and the end of model year 70). Time averaged quantities are denoted by an overbar

$$\overline{(\cdot)} = \frac{1}{t_1 - t_0} \int_{t_0}^{t_1} (\cdot) dt, \quad (5)$$

with  $t_{0,1}$  being start and end of model years 66 and 70 respectively. Following the work of Couespel et al. (2021), our focus is on the responses particularly within the subpolar gyre region, which is the area that is most bioactive in the present setup. We employ the same pre-defined box utilised in Couespel et al. (2021) for our analysis, defined as the area bounded between  $y = 35^\circ \text{ N}$ ,  $45^\circ \text{ N}$  and  $z = -700 \text{ m}$ , with the boundaries marked on by the black dashed lines in the subsequent figures where appropriate. The eventual supporting evidence suggests the hypothesis is largely true (see Table 2 in §5), but there are important subtle details to be elaborated on.

Models can display multiple equilibria and/or be affected by internal modes of variability (e.g., Sérazin et al., 2017; Zanna et al., 2019a), where diagnostics would vary depending on the period of analysis. In the present model we considered ensemble experiments perturbing the initial conditions, and we found no significant internal variability beyond inter-annual periods. This could be because of the idealised model as well as the choice of forcing, which has a repeating seasonal cycle and no mode of variability longer than a year. While there is reasonable variability

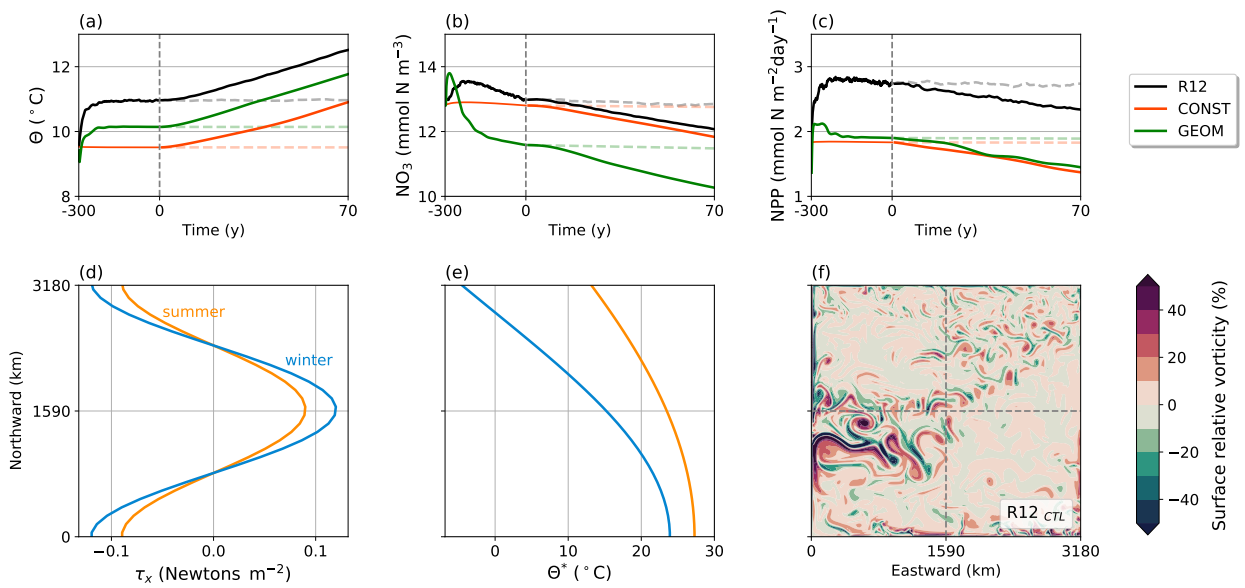


Figure 1: (Top row) Time series of various quantities from the 300-year spin-up (i.e., model years  $-300$  to  $0$  years) and the experimental period ( $0$  to  $70$  years), for the model truth (R12, black line) and coarse resolution calculations (CONST, red line; GEOM, green line) for the pre-industrial control scenario (CTL, faint dashed line) and the climate change scenario (CC, solid lines). The time axes are linear in the spinup and analysis period individually. (a) Averaged ocean temperature  $\Theta$  ( $^{\circ}\text{C}$ ) over the top  $700$  m of the model domain. (b) Model nitrate concentration ( $\text{NO}_3$ ,  $\text{mmol N m}^{-3}$ , where N is the nitrogen currency unit) over the top  $700$  m of the model domain. (c) Domain integrated Net Primary Production (NPP,  $\text{mmol N m}^{-2}\text{day}^{-1}$ ). Shown also are (d) the idealised purely zonal wind stress forcing  $\tau_x$  ( $\text{Newtons m}^{-2}$ ) with seasonal cycle limits, (e) the pseudo-atmospheric temperature  $\Theta^*$  ( $^{\circ}\text{C}$ ) with seasonal cycle limits, and (f) a snapshot of the surface relative vorticity of the model truth R12 (units of the planetary vorticity  $f_0 = 2\Omega \sin(20^{\circ})$ , where  $\Omega$  is the planetary rotation rate, and  $20^{\circ}$  is the southern edge of the domain).



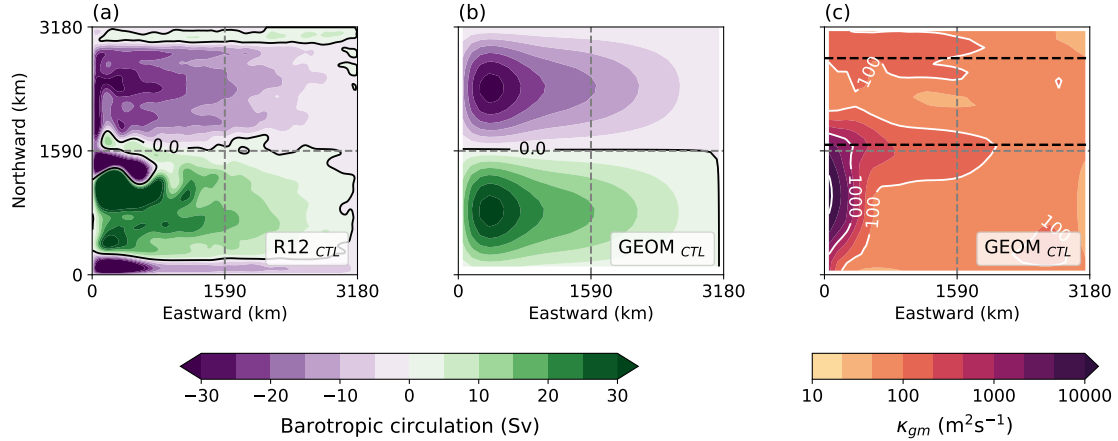


Figure 2: The barotropic streamfunction  $\Psi_{\text{baro}}$  (in  $\text{Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) of (a) R12 and (b) GEOM, with the zero contour overlaid as a black line; CONST visually looks identical to GEOM, and has been omitted. Panel (c) shows the resulting  $\kappa_{\text{gm}}$  distribution from GEOM with the choice of parameters in Table 1; the area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text.

237 particularly in the eddy resolving calculations (cf., Lévy et al., 2014), we are interested in broad scale and long time  
 238 changes, and the inter-annual variability is averaged out with a multi-year average from our diagnostics. As such, the  
 239 conclusions drawn from our diagnostics here should thus be regarded as statistically significant. However, we should  
 240 stress again that we focus on the qualitative rather than the quantitative differences. The primary interest is to see if  
 241 one parameterisation scheme performs ‘better’, and less on *how much* better; the latter is more sensitive to context  
 242 and should be quantified using more realistic models.

### 243 3. Comparison of pre-industrial controls

#### 244 3.1. Physical responses

245 Fig. 2(a, b) shows the barotropic streamfunction (where the tilde denotes a dummy integration variable)

$$\Psi_{\text{baro}}(x, y) = \int_{-H}^0 \int_0^x \bar{v}(\tilde{x}, y, z) d\tilde{x} dz \quad (6)$$

246 for the R12 and GEOM calculation; the CONST one has been omitted since visually it is indistinguishable from the  
 247 GEOM one. Both models display the familiar northern hemisphere double gyre pattern with a subtropical gyre to  
 248 the south and a subpolar gyre to the north. In the R12 calculation, because of eddy rectification effects, the modelled  
 249 Western Boundary Current is more variable and stronger through eddies converging momentum into the jet extension  
 250 (e.g., Lévy et al., 2010; Waterman et al., 2011; Waterman and Lilly, 2015). The Western Boundary Current is also  
 251 slightly south of the latitudinal center line, even though the zonal wind stress is symmetric about the same center  
 252 line (cf. Lévy et al., 2010). In addition, relatively strong re-circulation regions exist near the northern and southern  
 253 boundaries as Fofonoff gyres (e.g., Berloff, 2005; Marshall and Adcroft, 2010). All such features are absent in the  
 254 coarse resolution non-eddy models relying on the standard diffusive mesoscale parameterisations.

255 In Fig. 2(c) we show the resulting  $\kappa_{\text{gm}}(x, y)$  from GEOM. Note that  $\kappa_{\text{gm}}$  is large (on the order of a few thousand  
 256  $\text{m}^2 \text{ s}^{-1}$ ) on the Western Boundary Current in the subtropical gyre. The much smaller values of  $\kappa_{\text{gm}}$  within the subpolar  
 257 gyre and particularly its values near the northern boundary will be discussed later. The resulting domain-averaged  
 258 value of  $\kappa_{\text{gm}}$  is about  $300 \text{ m}^2 \text{ s}^{-1}$ , and we note the gyre models studied here using such a small value of  $\kappa_{\text{gm}}$  everywhere  
 259 leads to un-physical deep convection particularly along the Western Boundary Current (not shown; cf. Danabasoglu  
 260 et al., 1994). One benefit then with parameterisation schemes that allow spatial variations of  $\kappa_{\text{gm}}$  is that  $\kappa_{\text{gm}}$  can be  
 261 large only where it needs to be large, and this point will be revisited throughout the article.

262 The concentrated signal of  $\kappa_{\text{gm}}$  on the Western Boundary observed in Fig. 2(c) is also consistent with the fact  
 263 that the represented Western Boundary Current in a coarse resolution model is rather weak, which suggests that the

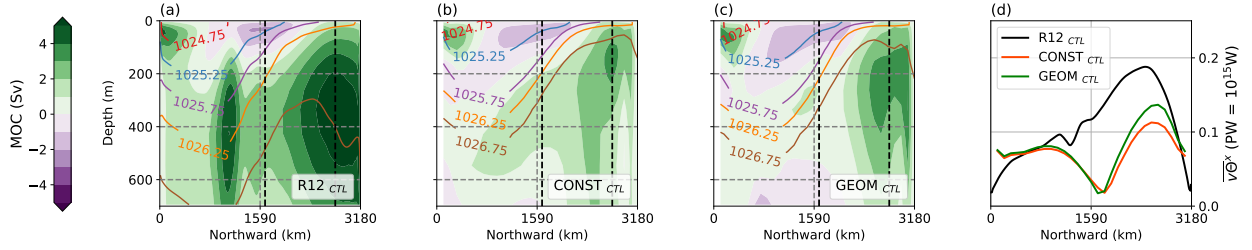


Figure 3: The diagnosed Meridional Overturning Circulation streamfunction  $\Psi_{\text{MOC}}$  from the model (shading, in  $\text{Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) and lines of constant potential density referenced to sea level (contours, in  $\text{kg m}^{-3}$ ), for (a) R12, (b) CONST, and (c) for GEOM. Panel (d) shows the diagnosed northward heat transport (in units of  $\text{PW} = 10^{15} \text{ W}$ ) for all three cases. The area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text.

264 resulting Meridional Overturning Circulation in the system is also rather weak. In Fig. 3 we show metrics relating to  
 265 the overturning circulation, namely the diagnosed Meridional Overturning Circulation streamfunction

$$\Psi_{\text{MOC}}(y, z) = \int_{-H}^z \int_0^{L_x} \bar{v}(x, y, \tilde{z}) \, dx \, d\tilde{z} \quad (7)$$

266 between different calculations, some sample isopycnals using potential density referenced to sea level, as well as the  
 267 diagnosed depth integrated and zonal-mean northward heat transport  $\bar{v}\bar{\Theta}^y$ . As shown in Fig. 3, the coarse resolution  
 268 models CONST and GEOM in general have a weaker overturning strength, partially because of a weaker modelled  
 269 Western Boundary Current arising from the more diffuse nature of the model. The particularly weak overturning in  
 270 the subtropical region of the coarse resolution models compared to the R12 model truth is consistent with a weak  
 271 Western Boundary Current, related to the structure of the displayed isopycnals via thermal wind shear relation. The  
 272 weaker overturning is reflected in the reduced northward transport of heat. The use of GEOMETRIC provides mild  
 273 improvements to the represented overturning strength particularly in the subpolar gyre, where the diagnosed  $\Psi_{\text{MOC}}$  in  
 274 GEOM is stronger than that in CONST and closer to R12 (area-weighted average root-mean-square mismatch to R12  
 275 of 1.99 Sv in GEOM compared to 2.29 Sv in CONST within the subpolar gyre box). The stronger MOC coincides  
 276 with a larger heat transport (area-weighted average heat transport of 0.094 PW in GEOM compared to 0.078 PW in  
 277 CONST, calculated from north of  $35^\circ \text{ N}$ ). This increased overturning strength is expected to have a positive effect on  
 278 the modelled biogeochemical response in the GEOM calculation, as we can expect increased nutrient transport into  
 279 the subpolar gyre by the nutrient stream or relay (e.g., Williams et al., 2017, 2011; Whitt and Jansen, 2020; Gupta  
 280 et al., 2022).

281 The double gyre model here is configured such that the downwelling is most prominent in the northern part of the  
 282 domain since this region is exposed to the coolest atmospheric temperatures, as seen in the maximum mixed layer  
 283 depths shown in Fig. 4(a-c) (diagnosed as the first depth below which  $|\sigma_\theta(z) - \sigma_\theta(z = -10 \text{ m})| > 0.01$ , where  $\sigma_\theta$   
 284 is the potential density referenced to sea level). The biggest differences between the calculations are at the northern  
 285 part of the domain, particularly in the northwestern corner of the domain. Fig. 4(d-f) shows the histogram of the  
 286 diagnosed mixed layer north of the subpolar region, where we see CONST has a notable skew towards shallower  
 287 mixed layer depths relative to GEOM and R12 (in terms of median and distribution). The more shallow mixed layer  
 288 depths observed in the CONST calculation are consistent with the decrease in the overturning strength, since the  
 289 mixed layer is correlated to the depths of deep water extent and overturning circulation. One rationalisation is that the  
 290 GM scheme flattens isopycnals and works against the steepening of isopycnals associated with deep water formation  
 291 and subsequent convective events. The CONST calculation employs a higher  $\kappa_{\text{gm}}$  value in the northern boundary  
 292 region compared with the resulting  $\kappa_{\text{gm}}$  in the GEOM calculation (see Fig. 2c), leading to a shallow bias compared  
 293 to GEOM. The causality highlights the importance of the magnitude and distribution of  $\kappa_{\text{gm}}$  in the modelled physical  
 294 mean state of coarse resolution models, where the mean transport pathways and strengths are being influenced by the  
 295 explicit or parameterised small-scale feedbacks.

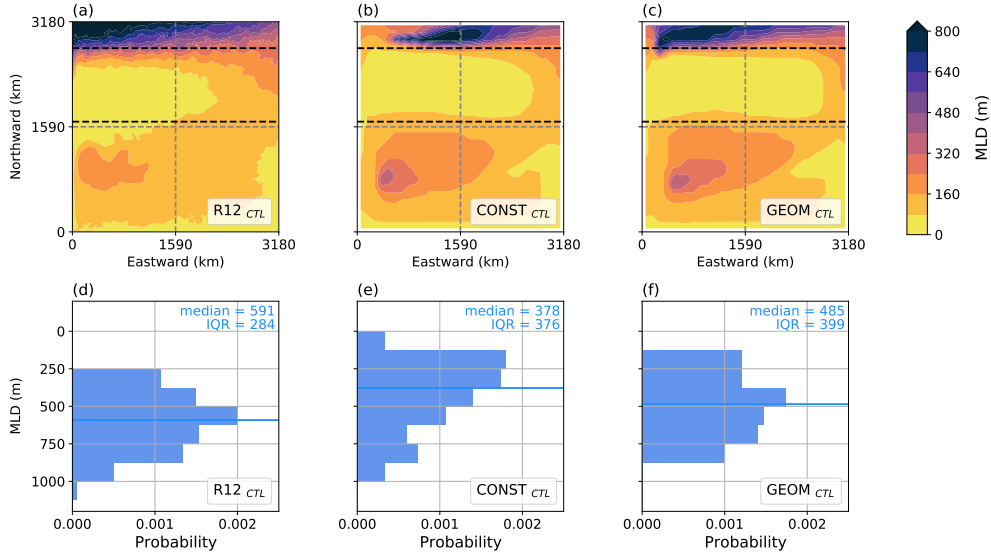


Figure 4: (Top row) Maximum mixed layer depth (m, diagnosed as the first depth below which  $|\sigma_\theta(z) - \sigma_\theta(z = -10 \text{ m})| > 0.01$ , where  $\sigma_\theta$  is the potential density referenced to sea level), for (a) R12, (b) CONST and (c) GEOM. The area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text. (Bottom row) histogram of mixed layer depth distributions and median (marked on as a line) north of the subpolar gyre region, for (d) R12, (e) CONST and (f) GEOM; the axes of the histograms have been swapped to enable ease of visual comparison.

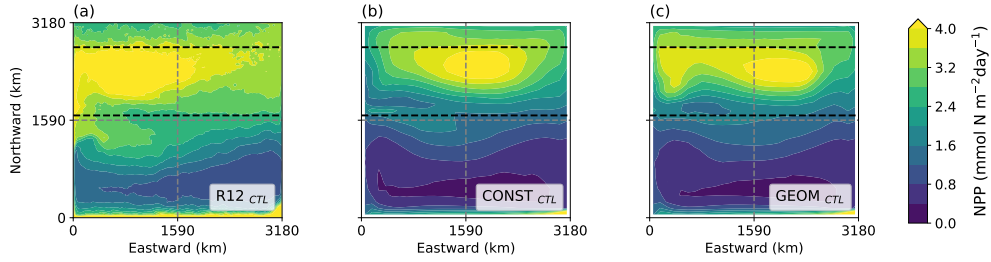


Figure 5: Vertically integrated Net Primary Production (NPP,  $\text{mmol N m}^{-2} \text{day}^{-1}$ , where N is the nitrogen currency) for (a) R12, (b) CONST and (c) GEOM. The area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text.

### 296 3.2. Biogeochemical responses

297 Since the CONST calculation is expected to have a weaker circulation (Fig. 3) and shallower mixed layer depths  
 298 (Fig. 4) relative to the GEOM case, we can expect that GEOM offers some improvements over CONST in the  
 299 biogeochemical response via changes in the nutrient transport. Fig. 5 shows the horizontal distribution of vertically  
 300 integrated NPP. For the integrated NPP averaged over the subpolar gyre region (units of  $\text{mmol N m}^{-2} \text{day}^{-1}$ , where N is  
 301 the nitrogen currency), R12 has the largest NPP at 3.67, compared to CONST at 2.76 and GEOM at 2.91 (respectively  
 302 a decrease of  $-24.8\%$  and  $-20.6\%$  relative to R12). The GEOM calculation results in NPP values closer to the model  
 303 truth R12 compared to the CONST calculation, which is consistent with our expectations, although the improvements  
 304 are somewhat modest.

305 As noted in §2, the biogeochemistry model takes no explicit account of temperature variations on the  
 306 biogeochemical activities themselves, so the changes observed are a result of the changes in the nutrient distributions.  
 307 While NPP has contributions from nitrate and ammonium, we focus our attention on nitrate as it is the dominant  
 308 form of dissolved inorganic nitrogen except in oxygen poor regions in the marine system (e.g. oxygen minimum  
 309 zones or coastal hypoxia zones). The  $f$ -ratio, the ratio between primary production arising from nitrate and total  
 310 primary production (e.g., Sarmiento and Gruber, 2006, §4), is relatively constant over the set of calculations at around

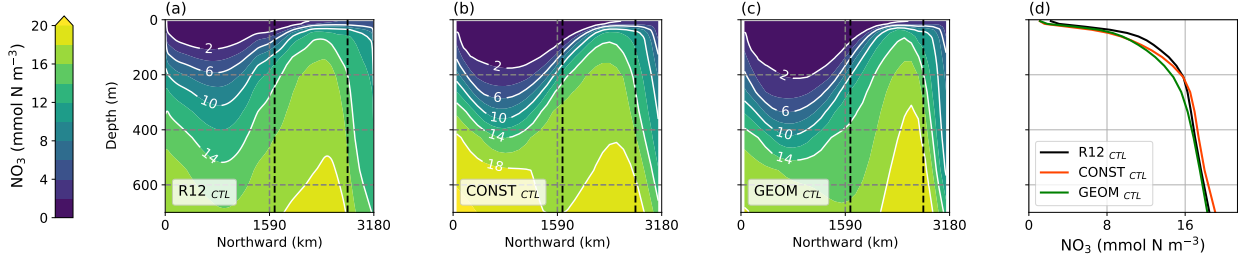


Figure 6: (a, b, c) Vertical distribution of zonally averaged nitrate concentration ( $\text{NO}_3^-$ ,  $\text{mmol N m}^{-3}$ , where N is the nitrogen currency unit), with lines of constant nitrate marked on. (d) The vertical distribution of nitrate in the predefined subpolar gyre box (see e.g. Fig. 5). The area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text.

0.43 (cf. Couespel et al., 2021). We show in Fig. 6 the zonally averaged vertical distribution of nitrate, and we see a suppression and elevation of nitrate concentration in the subtropical and subpolar gyres respectively across all models, consistent with the Ekman downwelling and upwelling from the choice of zonal wind forcing (e.g., §4 of Williams and Follows, 2011). In both the CONST and GEOM calculations, there is a strong decrease in nitrate concentration in the subtropical gyre compared with R12, possibly in line with the damped Western Boundary Current associated with the larger  $\kappa_{\text{gm}}$  values in the region. There is also an overall decrease over the whole subpolar gyre in the coarse resolution calculations. However, there is an increase in nitrate concentration in the northern parts of the subpolar gyre for GEOM compared to CONST (examined via the differences; not shown), which collectively leads to a mildly elevated NPP in the same subpolar gyre region in GEOM compared to CONST.

To analyse the transport properties of nitrate, we note that the advective contribution arises as  $\nabla \cdot (\mathbf{u}N)$ , where  $N$  denotes the nitrate concentration. Focusing on the subpolar gyre box (area enclosed by the black dashed lines in Fig. 5 and 6), noting that the box boundaries at longitudinal lines coincide with the model domain boundaries, by the divergence theorem and invoking no normal flow boundary conditions, we have

$$\int_{\text{box}} \nabla \cdot (\mathbf{u}N) \, dx \, dy \, dz = \left( \int_{\text{south}} + \int_{\text{north}} \right) vN \, dx \, dz + \int_{\text{bottom}} wN \, dx \, dy. \quad (8)$$

We can further consider the Reynolds decomposition

$$N = \bar{N} + N', \quad \bar{N}' = 0 \quad (9)$$

where the overbar is still the time average, which leads to

$$\overline{\mathbf{u}N} = \bar{\mathbf{u}}\bar{N} + \overline{\mathbf{u}'N'}, \quad (10)$$

respectively the total, the mean and the eddy advective flux of nutrients, and  $\mathbf{u}'$  is from the explicit velocity fluctuations in the case of explicit eddies, supplemented by parameterised eddy induced velocity  $\mathbf{u}^*$  when a GM-based parameterisation is active. We compute the vertical distribution of the vertical nitrate supply, i.e.,

$$\int_{y=L_s}^{y=L_n} \int_0^{L_x} (\bar{w}\bar{N} + \overline{w'N'}) \, dx \, dy, \quad (11)$$

where no vertical integration is implied, as well as the vertical cumulative integral of the horizontal nitrate supply at the southern and northern boundaries, i.e.,

$$\int_0^z \int_0^{L_x} (\bar{v}\bar{N} + \overline{v'N'}) \, dx \, dz. \quad (12)$$

Fig. 7 shows the total advective supply of nitrate into the subpolar gyre box in the vertical, at the southern boundary, and northern boundary. The dominant contribution to the total supply is in the mean component, although the eddy component is somewhat significant in the CONST case (not shown here, but see e.g. Couespel et al. 2021, Fig. A5).

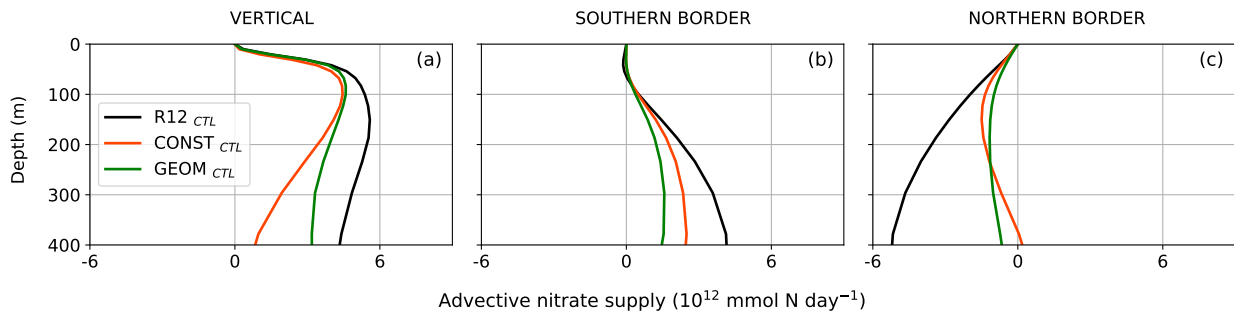


Figure 7: Total advective supply of nitrate ( $\text{NO}_3^-$ ,  $\text{mmol N day}^{-1}$ , where N is the nitrogen currency unit) into the pre-defined subpolar gyre box. (a) Vertical contribution. (b) Cumulative southern boundary contribution. (c) Cumulative northern boundary contribution. The northern boundary contribution was calculated with an extra minus sign, so positive values indicate a supply into the subpolar gyre box. Lateral and vertical diffusive contributions to nitrate flux are largely similar over the set of calculations and have been omitted.

334 In all calculations, the vertical nutrient supply rate is dominant over the top 150 m or so (magnitude of positive  
 335 values, Fig. 7a). The vertical mixing contribution dominates over the advective contribution over the top 50 m, but  
 336 is otherwise similar for the set of calculations (not shown here; cf. Couespel et al. 2021, Fig. 4). However, the  
 337 meridional nutrient supply becomes important with depth. The model truth R12 has the largest vertical gain (Fig. 7a)  
 338 throughout the depths considered, and the supply from the south and loss from the north (Fig. 7b, c) are both large,  
 339 consistent with R12 having the strongest overturning circulation out of the set of calculations considered here (cf.  
 340 Fig. 3). For the coarse resolution calculations, while the CONST case leads to a meridional gain of nutrients at depths  
 341 below 200 m via consideration of the residual of the gain from the south and loss from the north, the GEOM case has a  
 342 larger vertical supply throughout the depths considered, which is more consistent with the R12 calculation, and leads  
 343 to a larger overall total nutrient concentration, resulting in the larger diagnosed NPP values found in GEOM relative  
 344 to CONST.

345 One important point to emphasise here is that while the local eddy contributions are small, it is the eddy feedback  
 346 onto the mean state and the changes to the mean state that lead to the overall observed response (cf. Couespel et al.,  
 347 2021). Sample experiments (not shown) show that the NPP increases with decreasing  $\kappa_{\text{gm}}$ , so one might naïvely  
 348 argue that we should take  $\kappa_{\text{gm}}$  even smaller or even switch it off to improve the eddy component of nutrient supply and  
 349 increase the NPP. However, this is at the expense of introducing un-physical deep convection into the physical response  
 350 particularly along the Western Boundary Current, and the mean and eddy components are not isolated components  
 351 that one can ‘tune’ separately. The spatially varying nature of  $\kappa_{\text{gm}}$  afforded by GEOM allows the GM scheme to adjust  
 352 according to the physically modelled state, and suppressing its effects in the subpolar gyre where it is potentially  
 353 detrimental to the biogeochemical response. A calculation with a simpler prescription of  $\kappa_{\text{gm}}$  based on Treguier et al.  
 354 (1997) as implemented into NEMO (which requires a specification of a maximum  $\kappa_{\text{gm}}$  and varies in space according to  
 355 the baroclinic growth rate) was performed here and gives similar conclusions in the control calculation to GEOM (not  
 356 shown). The resulting  $\kappa_{\text{gm}}$  is not unlike that shown in Fig. 2(c), but with a much more gradual spatial variation limited  
 357 by the choice of the maximum  $\kappa_{\text{gm}}$ , taken here to be  $1000 \text{ m}^2 \text{ s}^{-1}$ . The resulting diagnostics are largely similar and  
 358 certainly improve upon the CONST case, for reasons detailed already. In that regard, it is the spatially varying nature  
 359 of  $\kappa_{\text{gm}}$  afforded by the more updated schemes that results in a modelled state that is closer to the model truth in the  
 360 selected diagnostics. However, it is known that schemes based on a prescribed maximum  $\kappa_{\text{gm}}$  limits how the models  
 361 can react to climate change scenarios (e.g., Fox-Kemper et al., 2019). The GEOM scheme and other energetically  
 362 constrained parameterisations have no such limitations, and we expect such schemes to behave in a favourable way  
 363 under the climate change scenarios.

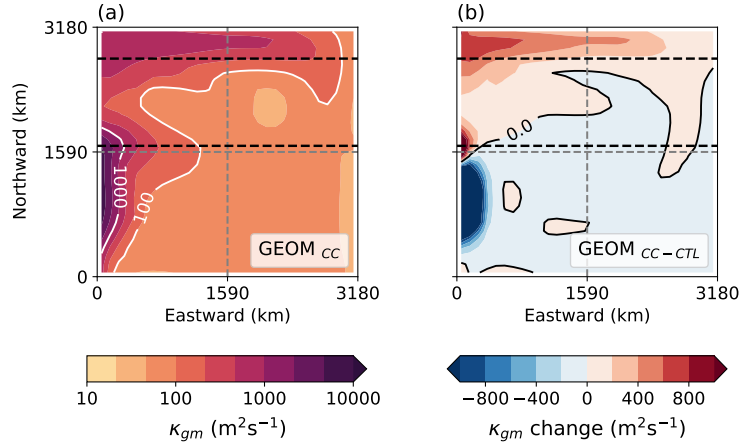


Figure 8: (a) The resulting  $\kappa_{gm}$  profile associated with GEOM under the climate change experiment, and (b) the raw differences of the  $\kappa_{gm}$  between the control and the climate change scenario (i.e., Fig. 8a minus Fig. 2c). Red colours in panel b indicate a decrease in  $\kappa_{gm}$ . The area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text.

#### 364 4. Sensitivities under idealised climate change

##### 365 4.1. Physical responses

366 Fig. 8 shows the  $\kappa_{gm}$  distribution from GEOM under the climate change scenario, and the raw differences compared  
 367 to the control scenario. The notable feature here is the increase in  $\kappa_{gm}$  towards the northern boundary in Fig. 8(b) where  
 368 the model deep water is formed. Given the discussion in the previous section, we would expect the GEOM calculation  
 369 in this case to have an over weakened overturning circulation, leading to a decrease in the nutrient supply and the NPP.  
 370 Details turn out to matter, as will be seen shortly.

371 Fig. 9(a, b, c) shows the raw differences between the overturning streamfunction under climate change and the  
 372 control case (cf. Fig. 3a, b, c). While both R12 and GEOM show a significant decrease in the very northern part of  
 373 the domain, this feature seems to be absent in the CONST case. This lack of decrease in the overturning strength  
 374 in CONST (and even a mild increase shown at the bottom right half of Fig. 3b) might have contributed to the  
 375 observed sensitivity in the diagnosed northward heat transport: the R12 and GEOM calculations both show only  
 376 small increases in the heat transport relative to the respective control scenario north of  $35^\circ$  N (+2.0% and +5.7%  
 377 increase in the area-weighted average respectively), but the CONST calculation shows a rather significant increase  
 378 in the heat transport in the same region (+23.5% increase in the area-weighted average). The response seen in R12  
 379 and GEOM are likely because of the increase in temperature offsetting the decrease in the advective velocity. On the  
 380 other hand, the magnitude of the response in CONST is quite significant and unlikely to result solely from increases in  
 381 water temperature, suggesting that the overturning response under climate change scenario in CONST is inconsistent  
 382 with the actual dynamics in the model truth with explicit eddies.

383 Fig. 10 shows the changes in the diagnosed maximum mixed layer, and all panels show that the mixed layer  
 384 depth has generally shoaled across all calculations under the climate change scenario, particularly in the region  
 385 near the northern boundary. This is consistent with the warming of the atmosphere and the associated decrease in  
 386 the ocean buoyancy loss. The shoaling is reflected in the shift of the median values, as well as a decrease in the  
 387 quartile ranges. However, note that the coarse resolution models appear to have a noticeably shallower mixed layer,  
 388 as seen in the histograms and the median values in Fig. 10(e, f) compared to the R12 calculations with explicit eddies  
 389 (Fig. 10d). From the preceding discussion, we might expect that the decrease of the maximum mixed layer depth is  
 390 more significant in GEOM given the increase in  $\kappa_{gm}$  in the region (cf. Fig. 8b), leading to a stronger flattening of  
 391 isopycnals that acts against the formation of deep mixed layers, impacting the overturning circulation.

392 We can quantify the magnitude of the overall shoaling by numerically computing the 1-Wasserstein distance  
 393  $W_1(\mu, \nu)$  (sometimes known as the earth mover's distance, e.g. Villani 2008), which measures the distance between  
 394 two discrete probability distributions  $\mu$  and  $\nu$  (i.e., a measure of the 'difference' between two histograms). Doing



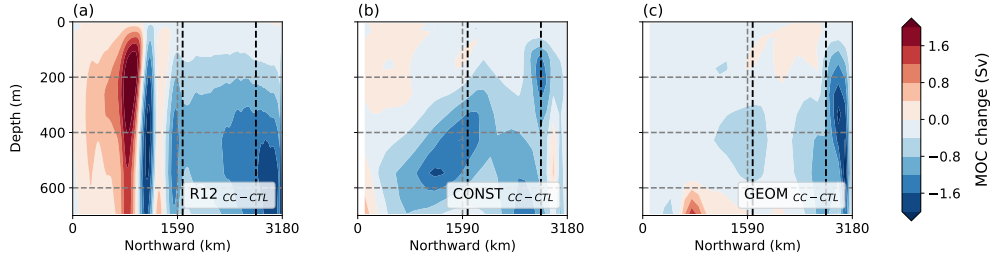


Figure 9: Raw differences of the diagnosed overturning streamfunction  $\Psi_{\text{MOC}}$  from the climate change scenario with the corresponding control scenario (see Fig. 3, for (a) R12, (b) CONST, (c) GEOM; negative values mostly correspond to a decrease in the overturning strength. The area enclosed by black dashed lines denote the boundaries of the subpolar gyre box mentioned in text. The resulting area-weighted change in the diagnosed northward heat transport north of  $35^\circ$  N given by the southern black dashed line is +2.0%, +23.5% and +5.7% respectively for R12, CONST and CONST.

395 so leads to  $W_1(\text{GEOM}_{\text{CTL}}, \text{GEOM}_{\text{CC}}) \approx 216$  while  $W_1(\text{CONST}_{\text{CTL}}, \text{CONST}_{\text{CC}}) \approx 170$  in the present mixed layer  
 396 depth diagnostic, thus supporting the conclusion that the GEOM calculation changes more within the climate change  
 397 scenario. For completeness,  $W_1(\text{R12}_{\text{CTL}}, \text{R12}_{\text{CC}}) \approx 217$ , so in this metric GEOM has a sensitivity that is more in line  
 398 with R12 than CONST.

#### 399 4.2. Biogeochemical responses

400 Fig. 11 shows the horizontal distribution of the raw differences in the vertically integrated NPP between the climate  
 401 change and control scenarios. There is a decrease in NPP across all models under climate change particularly in the  
 402 subpolar gyre, although there are isolated spots in the R12 calculation where NPP has marginally increased (south of  
 403 the Western Boundary Current separation, and at the northern boundary where downwelling occurs). The decrease in  
 404 NPP in the coarse resolution models are concentrated particularly in the east of the subpolar gyre region, and a small  
 405 patch towards the western boundary in the GEOM calculation.

406 The integrated NPP value averaged over the subpolar gyre box declines under climate change for all calculations,  
 407 with R12 at 3.16 (−13.8%), CONST at 2.13 (−22.9%) and GEOM at 2.22 (−23.6%), where the raw numbers are in  
 408 units of  $\text{mmol N m}^{-2} \text{day}^{-1}$  (N being the nitrogen currency), and the percentage difference is relative to the respective  
 409 calculations in the control scenario. The R12 model simulates the largest NPP overall, with the smallest decline  
 410 under the climate change scenario. The coarse resolution models significantly under predict the raw value of the  
 411 NPP, and also predict a more dramatic decline, in line with the previous results of Couespel et al. (2021). While it  
 412 is true that the GEOM calculation still predicts a higher NPP than the CONST calculation in both the control and  
 413 climate change scenario, the GEOM calculation displays more sensitivity to the change in forcing under the climate  
 414 change scenario, with a marginally larger NPP decline compared to CONST. Although the relative decrease in NPP  
 415 in CONST is smaller in magnitude than GEOM, we should also bear in mind that there is evidence indicating that  
 416 CONST possesses a sensitivity in the physical response that is inconsistent with the eddy resolving calculation R12  
 417 (e.g. Fig. 9), i.e., the CONST calculation might be “better” in the integrated NPP diagnostic, but not necessarily for  
 418 the right reasons.

419 The observed decline in NPP can again be attributed to the changes in the nutrient supply. We focus our attention  
 420 again on nitrate; there is a decrease in the  $f$ -ratio to around 0.40 (from around 0.43 in the control scenario) uniformly  
 421 across the set of calculations. Fig. 12 shows the differences in vertical distribution of zonally averaged nitrate between  
 422 the climate change and control scenario (see also Fig. 6). There is a decline of nitrate in the upper portions of the  
 423 subpolar gyre across all models, with a mild increase at depths, indicating a decline in upwelling consistent with the  
 424 strengthened stratification, as indicated for example by the buoyancy frequency  $N^2 \sim -\partial\rho/\partial z$  (not shown; cf. Fig.  
 425 A10 of Couespel et al. 2021). Fig. 12(d) shows the vertical distribution of the nitrate concentration averaged over the  
 426 subpolar gyre box, and it is noteworthy that the GEOM calculation has a vertical distribution change that is closer to  
 427 the model truth R12 than the CONST case.

428 The differences in the total advective fluxes into the subpolar gyre box (relative to the diagnosed supplies in Fig. 7)  
 429 are shown in Fig. 13; the dominant contribution to the total change was again found to be in the mean, although the

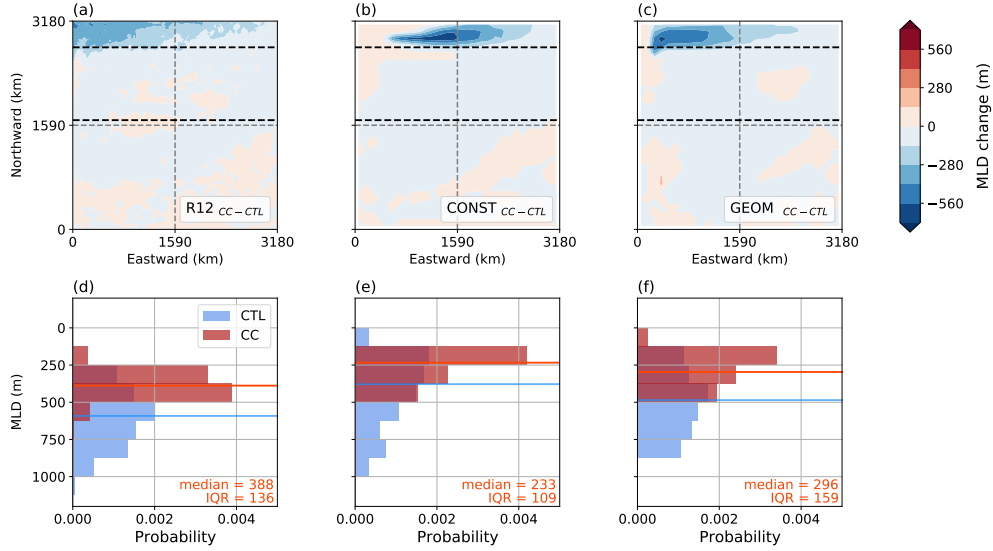


Figure 10: (Top row) Raw difference between climate change and control scenario maximum mixed layer depth (m, diagnosed as the first depth below which  $|\sigma_\theta(z) - \sigma_\theta(z = -10 \text{ m})| > 0.01$ , where  $\sigma_\theta$  is the potential density referenced to sea level), for (a) R12, (b) CONST and (c) GEOM; red colours denote a decrease in mixed layer depth (i.e., shoaling of the mixed layer). The area enclosed by black dashed lines denote the **boundaries of the subpolar gyre box** mentioned in text. (Bottom row) Histogram of the mixed layer depth distributions and median (marked on as a line) over the subpolar gyre region of both the climate change scenario (in red) and histogram of control scenario (in blue, cf. Fig. 4d, e, f) for (d) R12, (e) CONST and (f) GEOM; the axes of the histograms have been swapped to enable ease of visual comparison.

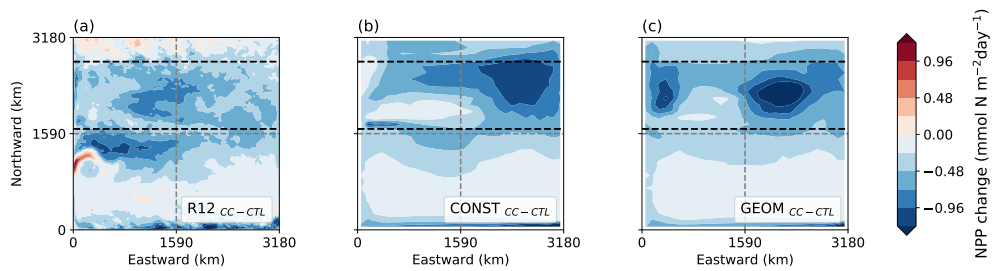


Figure 11: Raw differences of the vertically integrated Net Primary Production (NPP,  $\text{mmol N m}^{-2} \text{ day}^{-1}$ , where N is the nitrogen currency) between the climate change scenario and the control scenario (Fig. 5a, b, c) for (a) R12, (b) CONST and (c) GEOM. The area enclosed by black dashed lines denote the **boundaries of the subpolar gyre box** mentioned in text.

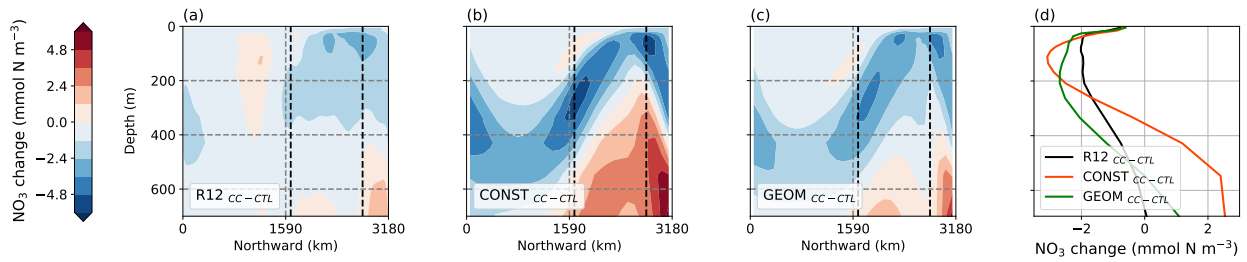


Figure 12: Raw differences between the vertical distribution of zonally averaged nitrate concentration ( $\text{NO}_3^-$ ,  $\text{mmol N m}^{-3}$ , where N is the nitrogen currency unit) between the climate change scenario and the control scenario (Fig. 6a, b, c) for (a) R12, (b) CONST and (c) GEOM; red values denote a decrease in nitrate concentration. The area enclosed by black dashed lines denote the subpolar gyre box mentioned in text. (d) Raw differences in the vertical distribution of nitrate in the predefined subpolar gyre box between the climate change scenario and the control scenario (Fig. 6d).

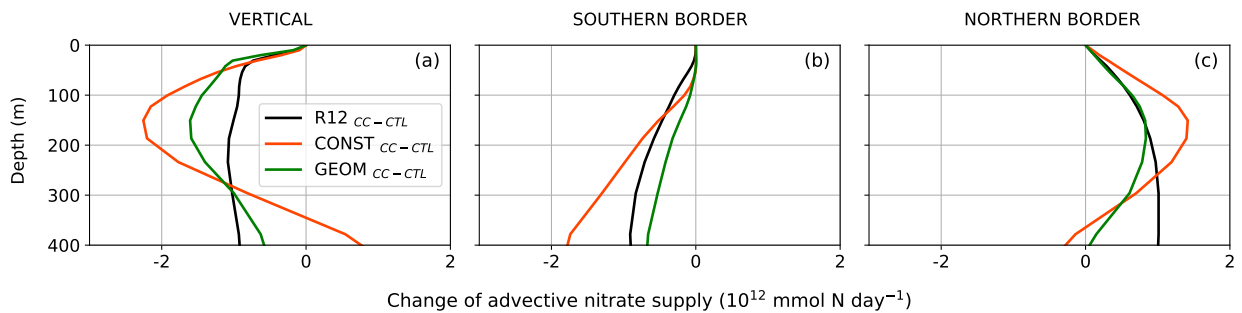


Figure 13: Differences of the total advective supply of nitrate ( $\text{NO}_3^-$ ,  $\text{mmol N day}^{-1}$ , where N is the nitrogen currency unit) into the pre-defined subpolar gyre box between the climate change and control scenario (cf. Fig. 13). (a) Vertical contribution. (b) Cumulative southern boundary contribution. (c) Cumulative northern boundary contribution. Negative values here largely mean a decrease in the supply into the subpolar gyre, while positive values mostly mean a decrease in the loss out of the subpolar gyre. Lateral and vertical diffusive contributions to nitrate flux are largely similar over the set of calculations and have been omitted.

430 eddy component is somewhat sizeable for the CONST case. Note that negative values in Fig. 13 largely mean a  
 431 decrease in the supply into the subpolar gyre, while positive values mostly mean a decrease in the loss out of the  
 432 subpolar gyre; the changes in the diffusive contributions are largely similar across the set of calculations and have  
 433 been omitted. The R12 case has a decreased vertical nutrient supply (Fig. 13a), southern boundary gain (Fig. 13b) and  
 434 northern boundary loss (Fig. 13c) under idealised climate change, broadly consistent with a decrease in the overturning  
 435 strength (cf. Fig. 9). For the coarse resolution calculations, we note that while CONST suffers a large decrease in the  
 436 vertical supply over the top 300 m (Fig. 13a), it seems to be compensated by an equally large decrease in the northern  
 437 boundary loss over the same depths (Fig. 13c). This is particularly interesting, given neither of these sensitivities are  
 438 nearly as dramatic in the R12 calculation, and suggests that this is a case where two “wrongs” happen to cancel out,  
 439 resulting in a reasonable integrated response in the NPP. On the other hand, the GEOM calculation over most panels  
 440 capture the shape of the R12 responses somewhat (and arguably substantially improve on the sensitivities displayed by  
 441 the CONST calculation), except at the northern boundary below 200 m, and there is a notable decrease in the vertical  
 442 supply in the upper 50 m of the ocean. While the GEOM calculation seems to respond in a way that is more consistent  
 443 with the model truth, it seems to (i) do things ‘wrong’ perhaps where it matters the most in the vertical nutrient supply  
 444 (upper part of the ocean where light availability and NPP are the largest), and (ii) not have the benefit of two ‘wrongs’  
 445 cancelling out as in the CONST case.

## 446 5. Conclusions and discussions

447 Numerical ocean models at non-eddy resolving to partially eddy permitting resolutions, requiring sub-grid physics  
448 parameterisation of the mesoscale processes, are going to remain the norm for the foreseeable future. Assessment of  
449 related parameterisations to highlight the possible benefits and deficiencies are required to constrain our uncertainties  
450 in the relevant conclusions and projections to be drawn from such models. To that end, this work presents an  
451 investigation of the joint physical and biogeochemical sensitivity to the choice of mesoscale eddy parameterisation in  
452 light of the recent developments in eddy parameterisation and its improvements into the modelled physical processes.  
453 The focus here is the more conventional diffusive closures utilised in coarse resolution non-eddy permitting ocean  
454 models, principally on the eddy induced advection represented by the GM scheme (e.g., Gent and McWilliams, 1990;  
455 Gent et al., 1995) and the GM version of the GEOMETRIC scheme (Marshall et al., 2012; Mak et al., 2018, 2022b).  
456 The latter takes a  $\kappa_{\text{gm}}$  that scales linearly with eddy energy (Eq. 3) and constrained by a parameterised eddy energy  
457 budget (Eq. 4). The present work highlights a need to evaluate the performance and tuning of eddy parameterisations  
458 on both the physical and biogeochemical response, and documents the performance of diffusive closures in coarse  
459 resolution models and the eddy resolving model truth as a precursor to an assessment into the eddy-permitting models  
460 as well as backscatter-type parameterisations.

461 To comprehensively assess the impacts afforded by the choice of mesoscale eddy parameterisation, this  
462 investigation employs a simplified and well-understood physical model (a double gyre configuration with a prescribed  
463 seasonal pattern leading to deep water formation near the northern boundary). Further, a simplified biogeochemistry  
464 model was chosen to focus on the chain of causality relating sensitivities afforded by the eddy parameterisation,  
465 its impact on the modelled state, its consequences for nutrient supply (e.g., Williams et al., 2017, 2011; Whitt and  
466 Jansen, 2020; Gupta et al., 2022), and in turn Net Primary Production (NPP). The choice of an idealised model with  
467 limited spatial extent allows for an eddy resolving model truth for coarse resolution models to compare against. The  
468 general model behaviours are entirely consistent with those reported in Couespel et al. (2021), where NPP decreases  
469 under the climate change scenario. This was attributed to the strengthening of upper ocean stratification, leading  
470 to a weakened overturning circulation, and thus weakening of nutrient supply into the subpolar gyre region where  
471 the NPP is strongest. The coarse resolution models display a more significant decrease in the NPP, attributed to a  
472 weaker overturning circulation in the coarse resolution models. The previous work was performed with the standard  
473 prescription of the GM scheme with a constant GM coefficient, and this work extends it in the first instance by  
474 considering a more updated GM-based eddy parameterisation, as well as critically assessing the model sensitivities as  
475 a result of the parameterisations, and in anticipation of assessing model performance in eddy permitting models.

476 A summary of the key diagnostics in this work and a comparison of the more updated GM-based eddy  
477 parameterisation with the constant case is given in Table 2. We have not found evidence for significant internal  
478 variability beyond the annual forcing period, which may be because of the choice of model and forcing set up, which  
479 only has a repeating seasonal cycle. While the reported diagnostics and conclusions should be considered statistically  
480 significant, we emphasise again that the main focus here is on the qualitative relative differences (e.g. differences  
481 in modelled state and/or sensitivities) and less on the quantitative absolute values (e.g. magnitude of differences in  
482 modelled state and/or sensitivities). The latter will be somewhat context dependent, so should be performed with a  
483 more realistic model for constraining climate projections.

484 The first main finding here is that the GM-version of the GEOMETRIC scheme (Marshall et al., 2012), which  
485 was found previously to lead to improved sensitivities in the modelled ocean mean state particularly when the domain  
486 includes a representation of the Southern Ocean (Mak et al., 2018, 2022b), leads to an improvement over the case  
487 where the GM coefficient  $\kappa_{\text{gm}}$  is set to be uniform over space, largely because the resulting  $\kappa_{\text{gm}}$  varies in space and  
488 is somewhat state-aware. The benefits afforded by a spatially varying  $\kappa_{\text{gm}}$  with reasonable properties are not entirely  
489 surprising and are somewhat known in the physical oceanography modelling community, though perhaps not so widely  
490 reported. In this particular model, the overall model response seems to be particularly sensitive to the value of  $\kappa_{\text{gm}}$  in  
491 the region with deep water formation, which is consistent with theoretical considerations through the impact on the  
492 overturning circulation (e.g., Williams and Follows, 2011). Extra calculations reported in Appendix A with prescribed  
493 spatially varying  $\kappa_{\text{gm}}$  further support the reported model responses. The observed model effect is rationalised here as  
494 the eddy induced advection acting against deep water formation and associated convective events, and a smaller  $\kappa_{\text{gm}}$  is  
495 conducive to deeper mixed layers and a stronger overturning circulation. The resulting state from using GEOMETRIC  
496 with  $\kappa_{\text{gm}} \sim E$  (where  $E$  is the total eddy energy) under the control scenario has a marginally stronger overturning

Diagnostic	R12 values	CONST values	GEOM values	improve over CONST
<b>overturning circulation (Sv)</b> (Fig. 3a, b, c and 9a, b, c)				
$L^2$ mismatch rel. R12 (CTL)	—	2.29	1.99	✓
$L^2$ mismatch rel. R12 (CC)	—	1.87	1.63	✓
<b>northward heat transport (<math>10^{15}</math> W)</b> (Fig. 3d, e, f and 9d, e, f)				
area average (CTL)	0.146	0.078	0.094	✓
area average (CC)	0.148 (+2.0%)	0.097 (+23.5%)	0.099 (+5.7%)	✓ (✓)
sensitivity ( $L^2$ )	0.004	0.020	0.006	✓
<b>northern mixed layer depth (m)</b> (Fig. 4 and 10)				
median (CTL)	592	378	486	✓
median (CC)	388 (-34.8%)	234 (-38.2%)	296 (-39.0%)	✓ (×)
quartile range (CTL)	285	376	399	×
quartile range (CC)	137 (-51.9%)	110 (-70.9%)	160 (-60.0%)	× (✓)
sensitivity (1-Wasserstein)	217	170	216	✓
<b>NPP (<math>\text{mmol N m}^{-2} \text{ day}^{-1}</math>)</b> (Fig. 5 and 11)				
area average (CTL)	3.67	2.76	2.91	✓
area average (CC)	3.16 (-13.8%)	2.13 (-22.9%)	2.22 (-23.6%)	✓ (×)
<b><math>\text{NO}_3^-</math> concentration (<math>\text{mmol N m}^{-3}</math>)</b> (Fig. 6d and 12d)				
area average (CTL)	15.61	15.57	15.01	×
area average (CC)	14.54 (-6.9%)	15.48 (-0.6%)	13.70 (-8.7%)	× (✓)
sensitivity ( $L^2$ )	1.31	2.14	1.84	✓

Table 2: Summary of diagnostics and their sensitivities for the set of calculations, where the bracketed numbers denote the percentage differences of the diagnostic between the climate change (CC) and control (CTL) scenario, and  $L^2$  denotes the area-weighted average root-mean-square difference (and has the same units as the diagnostics themselves).

497 circulation (Fig. 3) and more consistent statistics in the mixed layer depths (Fig. 4). This leads to a higher nutrient  
498 supply rate and NPP (Table 2), with modelled nutrient transport properties (Fig. 7) that are more consistent with the  
499 model truth over the CONST case. It was verified in the extra calculations with the Treguier et al. (1997) prescription  
500 of  $\kappa_{\text{gm}}$  (choosing maximum  $\kappa_{\text{gm}}$  value to be  $1000 \text{ m}^2 \text{ s}^{-1}$ ) leads to qualitatively similar results as GEOM in the control  
501 calculation (not shown). We would expect similar eddy parameterisation schemes employing mixing length arguments  
502 with  $\kappa_{\text{gm}} \sim \sqrt{K}$  (where  $K$  is the eddy kinetic energy), such as parts of MEKE (Jansen et al., 2019), to lead to qualitative  
503 similar results as GEOM here, although this has not been verified.

504 The second finding, one that is more subtle, is that a better physical response does not guarantee a better  
505 biogeochemical response, and a better biogeochemical response could arise from physically inconsistent physical  
506 responses, so there is a need to evaluate eddy parameterisations based on responses in both. The GM-version of the  
507 GEOMETRIC scheme does ‘worse’ in the integrated NPP metric to idealised climate change compared to the standard  
508 implementation, even though the model using the GEOMETRIC scheme actually seems to mostly improve on the bulk  
509 sensitivities as displayed by the model truth (e.g. nitrate concentration in Fig. 12, nutrient supply profiles in Fig. 13),  
510 and certainly more convincing and consistent than the CONST case (which was found to have a significant increase  
511 in the heat transport and different advective nutrient supply profiles). The observation here seems to stem from (i) the  
512 standard prescription of GM, while producing inconsistent sensitivities, happen to lead to cancellations (e.g. strong

513 decrease in supply of nutrient at southern boundary, Fig. 13*b*, offset by an even stronger decrease in loss of nutrient  
514 at northern boundary, Fig. 13*c*), and (ii) the GEOMETRIC scheme happens to lead to a change in regions that are  
515 particularly important to the model response (increase in the  $\kappa_{\text{gm}}$  towards the northern boundary, Fig. 8, and decrease  
516 in the vertical nutrient supply near the top of the ocean, Fig. 13*a*). Extra calculations with the Treguier et al. (1997)  
517 prescription of  $\kappa_{\text{gm}}$  with no re-tuning leads to diagnostics that are qualitatively close to the GEOM calculations, but at a  
518 lesser magnitude, so that the integrated results are somewhat better than GEOM (not shown). The better performance  
519 in NPP diagnostics however is likely because the resulting  $\kappa_{\text{gm}}$  is still artificially capped at the same value, so that  
520 the influence of  $\kappa_{\text{gm}}$  over the northern boundary region is muted compared with GEOM. We would expect similar  
521 eddy parameterisation schemes employing mixing length arguments with  $\kappa_{\text{gm}} \sim \sqrt{K}$  would do slightly better in the  
522 integrated NPP diagnostics than GEOM because of the more muted increase in  $\kappa_{\text{gm}}$  over the northern boundary region,  
523 although this has not been verified here.

524 One key point we make here is that care needs to be taken in the choice of metric to judge on the performance,  
525 and a combination of metrics might be required to highlight the intricacies of the model behaviour that are potentially  
526 masked behind a single metric, particularly when an average or integrated quantity is used. A case in hand here is that  
527 while the standard prescription of the GM scheme seems to lead to a ‘better’ response in integrated NPP, it is masking  
528 the fact that the contributing sensitivities are largely inconsistent with the model truth, i.e., two ‘wrongs’ can result  
529 in something that appears to look ‘right’. Ultimately the requirement should be that the biogeochemistry response is  
530 ‘better’ because the underlying ocean physics is ‘better’, and this work highlights a cautionary example where ocean  
531 models investigating biogeochemical responses should evaluate the modelled physical responses where possible.

532 The present use of an idealised model, in addition to providing a clean investigation into the strengths and  
533 deficiencies in the parameterisation schemes, also highlights lessons that we can learn from when extending our  
534 investigation to more complex but realistic models. If a GM-based parameterisation scheme is to be used in more  
535 realistic models, some form of tapering of  $\kappa_{\text{gm}}$  might be required as the regions of deep water formation are approached  
536 (cf. Hallberg, 2013), since this can have knock-on effects for the overturning circulation and affect biogeochemical  
537 responses in a non-local fashion. More complex biogeochemistry models are required to assess and highlight the  
538 impact of eddy parameterisations on the modelled biogeochemistry, for example carbon and oxygen budgets (e.g.,  
539 Friedlingstein et al., 2014; Berthet et al., 2019; Séférian et al., 2019; Kwiatkowski et al., 2020). Our focus here is  
540 more process oriented, and to set out a framework for evaluating the qualitative model responses and sensitivities  
541 to parameterisation, but the use of more complex and realistic models are required for quantifying uncertainties in  
542 projections, and will be a future focus.

543 Another choice made here is to focus on the eddy induced advection as represented by the GM scheme, sidelining  
544 the isoneutral diffusion as represented by the Redi scheme (e.g. Redi, 1982; Griffies, 1998). It is somewhat considered  
545 in the modelling community that the GM coefficient impacts the ventilation pathways via changes to the stratification  
546 profile, and in turn the rate of ventilation, while the isoneutral diffusion affects mostly the rate of ventilation (e.g.,  
547 England and Rahmstorf, 1999; Matear, 2001; Gnanadesikan et al., 2015; Jones and Abernathey, 2019) without  
548 significantly affecting the modelled state (but see Chouksey et al. 2022). Theoretical developments as well as  
549 numerical assessment of the GM-based schemes are somewhat more active and mature (e.g., Eden and Greatbatch,  
550 2008; Hofman and Morales Maqueda, 2011; Mak et al., 2018; Jansen et al., 2019; Bachman, 2019) compared to  
551 that of isoneutral diffusion (e.g., Smith and Marshall, 2009; Ferrari and Nikurashin, 2010; Abernathey et al., 2013;  
552 Groeskamp et al., 2021). With that in mind prior to our investigation, we have mostly focused on the GM-based  
553 schemes, but we considered simulations varying the spatially constant isoneutral diffusion coefficient  $\kappa_{\text{iso}}$ . Our sample  
554 simulations varying the spatially constant  $\kappa_{\text{iso}}$  by itself seems to have very minor to negligible impacts for this model,  
555 but there are feedback loops present if  $\kappa_{\text{gm}}$  is state-aware (see Appendix A). A systematic and comprehensive  
556 assessment of the isoneutral diffusion parameterisation schemes is a major undertaking, and we opted to postpone  
557 the related investigation.

558 While we would like to make use of mesoscale resolving models generally, these are still computationally  
559 prohibitive and likely to remain so for the foreseeable future. As a compromise, there is an increasing focus on  
560 eddy permitting models, to broadly refer to ocean models around  $1/2^\circ$  to  $1/9^\circ$  horizontal resolution, where mesoscale  
561 eddies have an explicit but incomplete representation (e.g. the explicit eddy field is substantially less energetic, from  
562 measures such as the explicit eddy kinetic energy). As noted at the beginning of this work, existing geostrophic  
563 mesoscale eddy parameterisations largely split into diffusive closures, which was the subject of this present work,



564 and backscatter approaches (e.g., Bachman, 2019; Jansen et al., 2019). The former is more targeted towards coarse  
565 resolution model without an explicit representation of eddies (e.g. models with around  $1^\circ$  horizontal resolution, such  
566 as the NEMO ORCA1 model). The latter in principle should work across models at different resolutions, but the  
567 working consensus at the time of writing seems to be that backscatter approaches work better in eddy permitting  
568 models, energising the eddies that are explicitly represented by the model itself. Given the increase in available  
569 computational power for performing global ocean models and Earth System Models at eddy permitting resolutions  
570 (normally around  $1/4^\circ$  horizontal resolution), and the benefits that result once ocean models start to become eddy  
571 permitting (e.g., Hewitt et al., 2017, 2020), an assessment into backscatter parameterisations analogous to the one  
572 carried out here is a priority, and is currently the subject of investigation.

## 573 Acknowledgements

574 This research was supported by Southern Marine Science and Engineering Guangdong Laboratory (Guangzhou)  
575 (SMSEGL20SC01) and the RGC General Research Fund 16304021. DC is funded by the Research Council of Norway  
576 under grant 275268 (COLUMBIA). ML acknowledges financial support from the Chanel Research Chair of ENS-PSL.  
577 The numerical model modifications, analyses code and sample model data including those mentioned in text but not  
578 shown may be found on Zenodo at <http://dx.doi.org/10.5281/zenodo.7612270>.

## 579 Appendix A. Model dependence on other parameters

580 Here we provide some further evidence for our assertion that the model mostly depends on  $\kappa_{\text{gm}}$ , particularly its  
581 values around the northern boundary where deep water formation occurs, and a brief description of model dependence  
582 on other key uncertain parameters within the system, namely the GEOMETRIC parameters  $\alpha$  and  $\lambda$  (see Eq. 3 and 4),  
583 and the isoneutral diffusion parameter  $\kappa_{\text{iso}}$ .

584 The dynamical argument here is that the presence of  $\kappa_{\text{gm}}$  leads to the flattening of isopycnals at the base of the  
585 mixed layer, which inhibits the deepening of mixed layers. Such an effect leads to a shallow bias of the mixed layer  
586 depths, a weakening of the overturning circulation, a reduction in nutrient supply and a reduction in NPP. The argument  
587 is in line with the previous results in the Appendix of Couespel et al. (2021), as well as our CONST experiments, where  
588 it is generally observed that the smaller the  $\kappa_{\text{gm}}$ , the higher the NPP (for precisely the aforementioned dynamical  
589 reasons, with signatures in the mixed layer depths and other physical metrics; not shown). The highest NPP occurs  
590 for the case when the GM scheme is completely switched off, but of course at the expense of introducing un-physical  
591 deep convection around the domain, as mentioned in the text.

592 It follows that varying the GEOM parameters  $\alpha$  and  $\lambda$  affect the resulting model results in a way that is  
593 consistent with varying  $\kappa_{\text{gm}}$  (varying the energy diffusion coefficient  $\eta_E$  leads to fairly weak responses in  $\kappa_{\text{gm}} \sim E$   
594 via modifying the sharpness of the modelled total eddy energy signature  $E$ ). Increasing  $\alpha$  and decreasing  $\lambda$  (or  
595 increasing the dissipation time-scale  $\lambda^{-1}$ ) both lead to increased  $\kappa_{\text{gm}}$  (consistent with Mak et al. 2017; see Marshall  
596 et al. 2017 for physical rationalisation), leading to decreases in NPP again for the aforementioned reasons. While the  
597 resulting modelled state under the control scenario differ depending on the choice of GEOM parameters, the resulting  
598 sensitivities under climate change for fixed choices of  $\alpha$  and  $\lambda$  are largely similar in magnitude, with a similar decrease  
599 in NPP, again because of the resulting increase in the  $\kappa_{\text{gm}}$  value over the northern boundary region. Although there are  
600 no strong constraints on the choice of  $\alpha$  and  $\lambda$  (but see attempts in Poulsen et al. 2019 and Mak et al. 2022a), it is at  
601 least reassuring that the conclusions regarding the sensitivity under climate change scenarios are robust.

602 The sensitivity of the modelled state to the  $\kappa_{\text{gm}}$  value at the northern boundary was further supported by results  
603 from experiments where  $\kappa_{\text{gm}}$  was artificially enhanced/suppressed under the climate change scenario, via manually  
604 modifying the CONST or GEOM  $\kappa_{\text{gm}}$  profiles in various regions. All results are consistent with the fact that increased  
605  $\kappa_{\text{gm}}$  at the northern boundary lead to decreased NPP for the physical chain of causality detailed above. Further, the  
606 results support the notion that GEOM produces ‘better’ results in the control scenario because of the spatially varying  
607  $\kappa_{\text{gm}}$ , but is perhaps over responding under the climate change scenario, as suggested in text.

608 Regarding sensitivity to the isoneutral diffusion, for lack of strong evidence to suggest which prescription functions  
609 the best, we opted to study the simple case of varying the constant diffusion coefficient  $\kappa_{\text{iso}}$ . Table A.3 documents the  
610 diagnosed NPP in the various scenarios, for both the CONST (which are results implicitly reported in the Appendix of

		NPP (CTL)	NPP (CC)	$\Delta$ NPP (self)
CONST	$\kappa_{\text{ISO}} = 500$	2.73	2.16	-20.7%
	$\kappa_{\text{ISO}} = 1000$	2.76	2.13	-22.9%
	$\kappa_{\text{ISO}} = 2000$	2.87	2.22	-22.9%
GEOM	$\kappa_{\text{ISO}} = 500$	2.75	2.06	-25.1%
	$\kappa_{\text{ISO}} = 1000$	2.91	2.22	-23.6%
	$\kappa_{\text{ISO}} = 2000$	3.11	2.48	-20.4%

Table A.3: Integrated Net Primary Production rate (NPP,  $\text{mmol N m}^{-2} \text{ day}^{-1}$ , where N is the nitrogen currency) integrated over the subpolar gyre box under the climate change scenario, for various calculations with varying  $\kappa_{\text{ISO}}$ .

611 Couespel et al. 2021) and GEOM calculations. The general conclusions here are that increasing  $\kappa_{\text{ISO}}$  leads to increased  
612 NPP, which is consistent with the increased transport of nutrients (at least in the lateral direction in the present gyre  
613 setting), certainly in the control scenario, and is suggestive in the climate change scenario. The observed sensitivity to  
614  $\kappa_{\text{ISO}}$  are stronger in the GEOM case, which arises from the nonlinear state dependence of  $\kappa_{\text{GM}}$ . In the CONST case  $\kappa_{\text{GM}}$   
615 and  $\kappa_{\text{ISO}}$  are independently prescribed, and the resulting modelled states at different  $\kappa_{\text{ISO}}$  are not so different between  
616 the experiments at least from a qualitative point of view (and consistent with the conclusions of Couespel et al. 2021).  
617 On the other hand, in the GEOM case, increases in  $\kappa_{\text{ISO}}$  leads to minor differences in the modelled state, which leads  
618 to changes in the calculated  $\kappa_{\text{GM}}$  (in this case a *decreasing*  $\kappa_{\text{GM}}$  over the northern boundary region, but with only very  
619 minor changes elsewhere in terms of the spatial pattern), which modifies onto the modelled state, leading to a modified  
620  $\kappa_{\text{GM}}$ . The claim here is that the changes in the NPP we are seeing in GEOM from changing  $\kappa_{\text{ISO}}$  arise from a positive  
621 feedback loop through its impact on  $\kappa_{\text{GM}}$  and resulting changes in the modelled stratification. The present nonlinear  
622 feedback loop between  $\kappa_{\text{ISO}}$ , the modelled state and  $\kappa_{\text{GM}}$  arising from GEOM should be studied further but is beyond  
623 the scope of the present work.

## 624 References

- 625 Abernathey, R., Ferreira, D., Klocker, A., 2013. Diagnostics of isopycnal mixing in a circumpolar channel. *Ocean Modell.* 72, 1–16.
- 626 Adcroft, A., Anderson, W., Bushuk, C. B. M., Dufour, C. O., Dunne, J. P., Griffies, S. M., Hallberg, R. W., Harrison, M. J., Held, I., Jansen, M. F.,  
627 John, J., Krasting, J. P., Langenhorst, A., Legg, S., Liang, Z., McHugh, C., Radhakrishnan, A., Reichl, B. G., Rosati, T., Samuels, B. L., Shao,  
628 A., Stouffer, R., Winton, M., Wittenberg, A. T., Xiang, B., Zadeh, N., Zhang, R., 2019. The GFDL global ocean and sea ice model OM4.0:  
629 Model description and simulation features. *J. Adv. Model. Earth Syst.* 11, 3167–3211.
- 630 Bachman, S. D., 2019. The GM+E closure: A framework for coupling backscatter with the Gent and McWilliams parameterization. *Ocean Modell.*  
631 136, 85–106.
- 632 Bachman, S. D., Marshall, D. P., Maddison, J. R., Mak, J., 2017. Evaluation of a scalar transport coefficient based on geometric constraints. *Ocean*  
633 *Modell.* 109, 44–54.
- 634 Bates, M., Tulloch, R., Marshall, J., Ferrari, R., 2014. Rationalizing the spatial distribution of mesoscale eddy diffusivity in terms of mixing length  
635 theory. *J. Phys. Oceanogr.* 44, 1523–1540.
- 636 Bellomo, K., Angeloni, M., Corti, S., von Hardenberg, J., 2021. Future climate change shaped by inter-model differences in Atlantic meridional  
637 overturning circulation response. *Nat. Commun.* 12 (1), 1–10.
- 638 Berloff, P., 2005. On rectification of randomly forced flows. *J. Mar. Res.* 63, 497–527.
- 639 Berthet, S., S  ferian, R., Bricaud, C., Chevallier, M., Voltaire, A., Eth  , C., 2019. Evaluation of an online grid-coarsening algorithm in a global  
640 eddy-admitting ocean biogeochemical model. *J. Adv. Model. Earth Syst.* 11, 1–25.
- 641 Bindoff, N. L., McDougall, T. J., 1994. Diagnosing climate change and ocean ventilation using hydrographic data. *J. Phys. Oceanogr.* 24 (6),  
642 1137–1152.
- 643 Bonan, G. B., Doney, S. C., 2018. Climate, ecosystems, and planetary futures: The challenge to predict life in Earth system models. *Science* 359,  
644 eaam8328.
- 645 Bopp, L., Resplandy, L., Orr, J. C., Doney, S. C., Dunne, J. P., Gehlen, M., annd C. Heinze, P. H., Ilyina, T., S  ferian, R., Tjiputra, J., Vichi, M.,  
646 2013. Multiple stressors of ocean ecosystems in the 21st century: projections with CMIP5 models. *Biogeosciences* 10, 6225–6245.
- 647 Bopp, L., Resplandy, L., Untersee, A., Le Mezo, P., Kageyama, M., 2017. Ocean (de)oxygenation from the Last Glacial Maximum to the twenty-  
648 first century: insights from Earth System models. *Phil. Trans. R. Soc. A* 375, 20160323.
- 649 Cessi, P., 2008. An energy-constrained parametrization of eddy buoyancy flux. *J. Phys. Oceanogr.* 38, 1807–1819.
- 650 Chelton, D. B., Schlax, M. G., Samelson, R. M., 2011. Global observations of nonlinear mesoscale eddies. *Prog. Oceanogr.* 91, 167–216.
- 651 Chouksey, A., Grisel, A., Chouksey, M., Eden, C., 2022. Changes in global ocean circulation due to isopycnal diffusion. *J. Phys. Oceanogr.* 52,  
652 2219–2235.
- 653 Couespel, D., L  vy, M., Bopp, L., 2021. Oceanic primary production decline halved in eddy-resolving simulations of global warming.  
654 *Biogeosciences* 18, 4321–4349.

655 Danabasoglu, G., McWilliams, J. C., Gent, P. R., 1994. The role of mesoscale tracer transports in the global ocean circulation. *Science* 264,  
656 1123–1126.

657 de Boer, A. M., Sigman, D. M., Toggweiler, J. R., Russell, J. L., 2007. Effect of global ocean temperature change on deep ocean ventilation.  
658 *Paleoceanography* 22, 1–15.

659 Doney, S. C., Ruckelshaus, M., Duffy, J. E., J. P. Barry, F. C., English, C. A., Galindo, H. M., Grebmeier, J. M., Hollowed, A. B., Knowlton, N.,  
660 Polovina, J., Rabalais, N. N., Sydeman, W. J., Talley, L. D., 2012. Climate change impacts on marine ecosystems. *Ann. Rev. Mar. Sci.* 4, 11–37.

661 Eden, C., Greatbatch, R. J., 2008. Towards a mesoscale eddy closure. *Ocean Modell.* 20, 223–239.

662 England, M. H., Rahmstorf, S., 1999. Sensitivity of ventilation rates and radiocarbon uptake to subgrid-scale mixing in ocean models. *J. Phys.*  
663 *Oceanogr.* 29, 2802–2828.

664 Farneti, R., Downes, S. M., Griffies, S. M., Marsland, S. J., Behrens, E., Bentsen, M., Bi, D., Biastoch, A., Böning, C. W., Bozec, A., Canuto, V. M.,  
665 Chassignet, E., Danabasoglu, G., Danilov, S., Diansky, N., Drange, H., Fogli, P. G., Gusev, A., Hallberg, R. W., Howard, A., Ilicak, M., Jung, T.,  
666 Kelley, M., Large, W. G., Leboissetier, A., Long, M., Lu, J., Masinam, S., Mishra, A., Navarra, A., Nurser, A. J. G., Patara, L., Samuels, B. L.,  
667 Sidorenko, D., Tsujino, H., Uotila, P., Wang, Q., Yeager, S. G., 2015. An assessment of Antarctic Circumpolar Current and Southern Ocean  
668 meridional overturning circulation during 1958–2007 in a suite of interannual CORE-II simulations. *Ocean Modell.* 94, 84–120.

669 Ferrari, R., Nikurashin, M., 2010. Suppression of eddy diffusivity across jets in the southern ocean. *J. Phys. Oceanogr.* 40, 1501–1519.

670 Ferreira, D., Marshall, J., Heimbach, P., 2005. Estimating eddy stresses by fitting dynamics to observations using a residual-mean ocean circulation  
671 model and its adjoint. *J. Phys. Oceanogr.* 35, 1891–1910.

672 Fox-Kemper, B., Adcroft, A. J., Böning, C. W., Chassignet, E. P., Curchitser, E. N., Danabasoglu, G., Eden, C., England, M. H., Gerdes, R.,  
673 Greatbatch, R. J., Griffies, S. M., Hallberg, R. W., Hanert, E., Heimbach, P., Hewitt, H. T., Hill, C. N., Komuro, Y., Legg, S., Le Sommer, J.,  
674 Masina, S., Marsland, S. J., Penny, S. G., Qiao, F., Ringler, T. D., Treguier, A. M., Tsujino, H., Uotila, P., Yeager, S. G., 2019. Challenges and  
675 prospects in ocean circulation models. *Front. Mar. Sci.* 6, 65.

676 Friedlingstein, P., Meinshausen, M., Arora, V. K., Jones, C. D., Anav, A., Liddicoat, S. K., Knutti, R., 2014. Uncertainties in CMIP5 climate  
677 projections due to carbon cycle feedbacks. *J. Climate* 27, 511–526.

678 Gaspar, P., Grégoris, Y., Lefevre, J., 1990. A simple eddy kinetic energy model for simulations of the oceanic vertical mixing: Tests at station papa  
679 and long-term upper ocean study site. *J. Geophys. Res.* 95, 16179–16193.

680 Gent, P. R., McWilliams, J. C., 1990. Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr.* 20, 150–155.

681 Gent, P. R., Willebrand, J., McDougall, T. J., McWilliams, J. C., 1995. Parameterizing eddy-induced tracer transports in ocean circulation models.  
682 *J. Phys. Oceanogr.* 25, 463–474.

683 Gnanadesikan, A., Pradal, M., Abernathey, R., 2015. Isopycnal mixing by mesoscale eddies significantly impacts oceanic anthropogenic carbon  
684 uptake. *Geophys. Res. Lett.* 42 (11), 4249–4255.

685 Griffies, S. M., 1998. The Gent–McWilliams skew flux. *J. Phys. Oceanogr.* 28, 831–841.

686 Groeskamp, S., LaCasce, J. H., McDougall, T. J., Rogé, M., 2021. Full-depth global estimates of ocean mesoscale eddy mixing from observations  
687 and theory. *Geophys. Res. Lett.* 47, e2020GL089425.

688 Grooms, I., 2015. A computational study of turbulent kinetic energy transport in barotropic turbulence on the  $f$ -plane. *Phys. Fluids* 27, 101701.

689 Gula, J., Molemaker, M. J., McWilliams, J. C., 2015. Topographic vorticity generation, submesoscale instability and vortex street formation in the  
690 Gulf Stream. *Geophys. Res. Lett.* 42, 4654–4062.

691 Gupta, M., Williams, R. G., Lauderdale, J. M., Jahn, O., Hill, C., Dutkiewicz, S., Follows, M. J., 2022. A nutrient relay sustains subtropical ocean  
692 productivity. *Proc. Natl Acad. Sci. USA* 119 (41), e2206504119.

693 Hallberg, R., 2013. Using a regulate function to regulate parameterizations of oceanic mesoscale eddy effects. *Ocean Modell.* 72, 92–103.

694 Helm, K. P., Bindoff, N. L., Church, J. A., 2011. Observed decreases in oxygen content of the global ocean. *Geophys. Res. Lett.* 38.

695 Hewitt, H. T., Bell, M. J., Chassignet, E. P., Czaja, A., Ferreira, D., Griffies, S. M., Hyder, P., McClean, J. L., New, A. L., Roberts, M. J., 2017.  
696 Will high-resolution global ocean models benefit coupled predictions on short-range to climate timescales? *Ocean Modell.* 120, 120–136.

697 Hewitt, H. T., Fox-Kemper, B., Pearson, B., Roberts, M., Klocke, D., 2022. The small scales of the ocean may hold the key to surprises. *Nat. Clim.*  
698 *Change* 12, 496–499.

699 Hewitt, H. T., Roberts, M., Mathiot, P., Biastoch, A., Blockley, E., Chassignet, E. P., Fox-Kemper, B., Hyder, P., Marshall, D. P., Popova, E.,  
700 Treguier, A., Zanna, L., Yool, A., Yu, Y., Beadling, R., Bell, M. J., Kuhlbrodt, T., Arsouze, T., Bellucci, A., Castruccio, F., Gan, B., Putrasahan,  
701 D., Roberts, C. D., Van Roekel, L., Zhang, Q., 2020. Resolving and parameterising the ocean mesoscale in earth system models. *Curr. Clim.*  
702 *Change Rep.* 6, 137–152.

703 Hofman, M., Morales Maqueda, M. A., 2011. The response of Southern Ocean eddies to increased midlatitude westerlies: A non-eddy resolving  
704 model study. *Geophys. Res. Lett.* 38, L03605.

705 Holmes, R. M., Groeskamp, S., Stewart, K., McDougall, T. J., 2022. Sensitivity of a coarse-resolution global ocean model to spatially variable  
706 neutral diffusion. *J. Adv. Model. Earth Syst.* 14, e2021MS002914.

707 IPBES, 2019. Global assessment report on biodiversity and ecosystem services of the Intergovernmental Science-Policy Platform on Biodiversity  
708 and Ecosystem Services.

709 IPCC, 2019. IPCC Special Report on the Ocean and Cryosphere in a Changing Climate.

710 Jackson, L., Hughes, C. W., Williams, R. G., 2006. Topographic control of basin and channel flows: the role of the bottom pressure torques and  
711 friction. *J. Phys. Oceanogr.* 36, 1786–1805.

712 Jansen, M. F., Adcroft, A., Khani, S., Kong, H., 2019. Toward an energetically consistent, resolution aware parameterization of ocean mesoscale  
713 eddies. *J. Adv. Model. Earth Syst.* 1, 1–17.

714 Jansen, M. F., Adcroft, A. J., Hallberg, R., Held, I. M., 2015. Parameterization of eddy fluxes based on a mesoscale energy budget. *Ocean Modell.*  
715 92, 28–41.

716 Jones, C. S., Abernathey, R. P., 2019. Isopycnal mixing controls deep ocean ventilation. *Geophys. Res. Lett.* 46, 13144–13151.

717 Jones, C. S., Abernathey, R. P., 2021. Modeling water-mass distributions in the modern and LGM ocean: Circulation change and isopycnal and  
718 diapycnal Mixing. *J. Phys. Oceanogr.* 51, 1523–1538.

719 Khwatiwala, S., Graven, H., Payne, S., Heimbach, P., 2018. Changes to the air-sea Flux and distribution of radiocarbon in the ocean over the 21st

720 century. *Geophys. Res. Lett.* 45, 5617–5626.

721 Klocker, A., Marshall, D. P., 2014. Advection of baroclinic eddies by depth mean flow. *Geophys. Res. Lett.* 41, L060001.

722 Kwiatkowski, L., Torres, O., Bopp, L., Aumont, O., Chamberlain, M., Christian, J. R., Dunne, J. P., Gehlen, M., Ilyina, T., John, J. G., Lenton,  
723 A., Li, H., Lovenduski, N. S., Orr, J. C., Palmieri, J., Santana-Falcón, Y., Schwinger, J., Séférian, R., Stock, C. A., Tagliabue, A., Takano, Y.,  
724 Tjiputra, J., Toyama, K., Tsujino, H., Watanabe, M., Yamamoto, A., Yool, A., Ziehn, T., 2020. Twenty-first century ocean warming, acidification,  
725 deoxygenation, and upper-ocean nutrient and primary production decline from cmip6 model projections. *Biogeosciences* 17, 3439–3470.

726 Lee, H., Moon, B., Jung, H., Park, J., Shim, S., La, N., Kim, A., Yum, S. S., Ha, J., Byun, Y., Sung, H. M., Lee, J., 2022. Development of the  
727 UKESM-TOPAZ Earth System Model (Version 1.0) and preliminary evaluation of its biogeochemical simulations. *Asia-Pac. J. Atmos. Sci.*  
728 58 (3), 379–400.

729 Lévy, M., Iovino, D., Resplandy, L., Klein, P., Madec, G., Tréguier, A., Masson, S., Takahashi, K., 2012. Large-scale impacts of submesoscale  
730 dynamics on phytoplankton: Local and remote effects. *Ocean Modell.* 43–44, 77–93.

731 Lévy, M., Klein, P., Tréguier, A., Iovino, D., Madec, G., Masson, S., Takahashi, K., 2010. Modifications of gyre circulation by sub-mesoscale  
732 physics. *Ocean Modell.* 34, 1–15.

733 Lévy, M., Resplandy, L., Lengaigne, M., 2014. Oceanic mesoscale turbulence drives large biogeochemical interannual variability at middle and  
734 high latitudes. *Geophys. Res. Lett.* 41, 2467–2474.

735 Lévy, M., Visbeck, M., Naik, N., 1999. Sensitivity of primary production to different eddy parameterizations: A case study of the spring bloom  
736 development in the northwestern Mediterranean Sea. *J. Mar. Res.* 57 (3), 427–448.

737 Li, G., Cheng, L., Zhu, J., Trenberth, K. E., Mann, M. E., Abraham, J. P., 2020. Increasing ocean stratification over the past half-century. *Nat. Clim.*  
738 *Change* 10 (12), 1116–1123.

739 Lotze, H. K., Tittensor, D. P., Bryndum-Buchholz, A., Eddy, T. D., Cheung, W. W. L., Galbraith, E. D., Barange, M., Barrier, N., Bianchi, D.,  
740 Blanchard, J. L., Bopp, L., Büchner, M., Bulman, C. M., Carozza, D. A., Christensen, V., Coll, M., Dunne, J. P., Fulton, E. A., Jennings, S.,  
741 Jones, M. C., Mackinson, S., Maury, O., Niiranen, S., Oliveros-Ramos, R., Roy, T., Fernandes, J. A., Schewe, J., Shin, Y., Silva, T. A. M.,  
742 Steenbeek, J., Stock, C. A., Verley, P., Volkholz, J., Walker, N. D., Worm, B., 2019. Global ensemble projections reveal trophic amplification of  
743 ocean biomass declines with climate change. *Proc. Natl. Acad. Sci. USA* 116 (26), 12907–12912.

744 Maddison, J. R., Marshall, D. P., 2013. The Eliassen–Palm flux tensor. *J. Fluid Mech.* 729, 69–102.

745 Madec, G., 2008. NEMO ocean engine. Note du Pôle de modélisation, Institut Pierre-Simon Laplace (IPSL), No. 27.

746 Mak, J., Avdis, A., David, T. W., Lee, H. S., Na, Y., Yan, F. E., 2022a. On constraining the mesoscale eddy energy dissipation time-scale. *J. Adv.*  
747 *Model. Earth Syst.* 14, e2022MS003223.

748 Mak, J., Maddison, J. R., Marshall, D. P., Munday, D. R., 2018. Implementation of a geometrically informed and energetically constrained  
749 mesoscale eddy parameterization in an ocean circulation model. *J. Phys. Oceanogr.* 48, 2363–2382.

750 Mak, J., Marshall, D. P., Maddison, J. R., Bachman, S. D., 2017. Emergent eddy saturation from an energy constrained parameterisation. *Ocean*  
751 *Modell.* 112, 125–138.

752 Mak, J., Marshall, D. P., Madec, G., Maddison, J. R., 2022b. Acute sensitivity of global ocean circulation and heat content to eddy energy dissipation  
753 time-scale. *Geophys. Res. Lett.* 49(8), e2021GL097259.

754 Marshall, D. P., Adcroft, A. J., 2010. Parameterization of ocean eddies: Potential vorticity mixing, energetics and Arnold’s first stability theorem.  
755 *Ocean Modell.* 32, 1571–1578.

756 Marshall, D. P., Ambaum, M. H. P., Maddison, J. R., Munday, D. R., Novak, L., 2017. Eddy saturation and frictional control of the Antarctic  
757 Circumpolar Current. *Geophys. Res. Lett.* 44, 286–292.

758 Marshall, D. P., Maddison, J. R., Berloff, P. S., 2012. A framework for parameterizing eddy potential vorticity fluxes. *J. Phys. Oceanogr.* 42,  
759 539–557.

760 Marshall, J., Adcroft, A., Hill, C., Perelman, L., Heisey, C., 1997a. A finite volume, incompressible Navier–Stokes model for studies of the ocean  
761 on parallel computers. *J. Geophys. Res.* 102, 5753–5766.

762 Marshall, J., Hill, C., Perelman, L., Adcroft, A., 1997b. Hydrostatic, quasi-hydrostatic, and non-hydrostatic ocean modelling. *J. Geophys. Res.* 102,  
763 5733–5752.

764 Martiny, A. C., Hagstrom, G. I., DeVries, T., Letscher, R. T., Britten, G. L., Garcia, C. A., Galbraith, E., Karl, D., Levin, S. A., Lomas, M. W.,  
765 Moreno, A. R., Talmy, D., Wang, W., Matsumoto, K., 2022. Marine phytoplankton resilience may moderate oligotrophic ecosystem responses  
766 and biogeochemical feedbacks to climate change. *Limnol. Oceanogr.* 67, S378–S389.

767 Matear, R. J., 2001. Effects of numerical advection schemes and eddy parameterizations on ocean ventilation and oceanic anthropogenic CO<sub>2</sub>  
768 uptake. *Ocean Modell.* 3, 217–248.

769 Matear, R. J., Hirst, A. C., McNeil, B. I., 2000. Changes in dissolved oxygen in the Southern Ocean with climate change. *Geochem. Geophys.*  
770 1 (11).

771 Munday, D. R., Johnson, H. L., Marshall, D. P., 2013. Eddy saturation of equilibrated circumpolar currents. *J. Phys. Oceanogr.* 43, 507–532.

772 Newsom, E., Zanna, L., Khatiwala, S., 2022. Relating patterns of added and redistributed ocean warming. *J. Climate* 35, 4627–4643.

773 Ni, Q., Zhai, X., Wang, G., Hughes, C. W., 2020a. Widespread mesoscale dipoles in the global ocean. *J. Geophys. Res. Oceans* 125,  
774 e2020JC016479.

775 Ni, Q., Zhai, X., Wang, G., Marshall, D. P., 2020b. Random movement of mesoscale eddies in the global ocean. *J. Phys. Oceanogr.* 50, 2341–2357.

776 Poulsen, M. B., Jochum, M., Maddison, J. R., Marshall, D. P., Nuterman, R., 2019. A geometric interpretation of Southern Ocean eddy form stress.  
777 *J. Phys. Oceanogr.* 49, 2553–2570.

778 Pradal, M., Gnanadesikan, A., 2014. How does the Redi parameter for mesoscale mixing impact global climate in an Earth system model? *J. Adv.*  
779 *Model. Earth Syst.* 6, 586–601.

780 Rahmstorf, S., 2002. Ocean circulation and climate during the past 120,000 years. *Nature* 419, 207–214.

781 Redi, M. H., 1982. Oceanic isopycnal mixing by coordinate rotation. *J. Phys. Oceanogr.* 12, 1154–1158.

782 Roberts, M. J., Jackson, L. C., Roberts, C. D., Meccia, V., Docquier, D., Koenigk, T., Ortega, P., Moreno-Chamarro, E., Bellucci, A., Coward, A. C.,  
783 Drijfhout, S., Exarchou, E., Gutjahr, O., Hewitt, H., Iovino, D., Lohmann, K., Putrasahan, D., Schiemann, R., Seddon, J., Terray, L., Xu, X.,  
784 Zhang, Q., Chang, P., Yeager, S. G., Castruccio, F. S., Zhang, S., Wu, L., 2020. Sensitivity of the Atlantic Meridional Overturning Circulation

785 to model resolution in CMIP6 HighResMIP simulations and implications for future changes. *J. Adv. Model. Earth Syst.* 12, e2019MS002014.

786 Sarmiento, J. L., Gruber, N., 2006. *Ocean Biogeochemical Dynamics*. Princeton University Press.

787 S  f  rian, R., Nabat, P., Michou, M., Saint-Martin, D., Voldoire, A., Colin, J., Decharme, B., Delire, C., Berthet, S., Chevallier, M., S  n  si, S.,

788 Franchisteguy, L., Vial, J., Mallet, M., Joetzjer, E., Geoffroy, O., Gu  r  my, J., Moine, M., Msadek, R., Ribes, A., Rocher, M., Roehrig, R.,

789 Salas-y-M  lia, D., Sanchez, E., Terray, L., Valcke, S., Waldman, R., Aumont, O., Bopp, L., Deshayes, J.,   th  , C., Madec, G., 2019. Evaluation

790 of CNRM Earth System Model, CNRM-ESM2-1: Role of Earth system processes in present-day and future climate. *J. Adv. Model. Earth. Syst.*

791 11 (12), 4182–4227.

792 S  razin, G., Jaymond, A., Leroux, S., Penduff, T., Bessi  res, L., Llovel, W., Barnier, B., Molines, J., Terray, L., 2017. A global probabilistic study

793 of the ocean heat content low-frequency variability: Atmospheric forcing versus oceanic chaos. *Geophys. Res. Lett.* 44 (11), 5580–5589.

794 Smith, K. S., Marshall, J., 2009. Evidence for enhanced eddy mixing at middepth in the Southern Ocean. *J. Phys. Oceanogr.* 39, 50–69.

795 Stewart, A. L., McWilliams, J. C., Solodoch, A., 2021. On the role of bottom pressure torques in wind-driven gyres. *J. Phys. Oceanogr.* 51,

796 1441–1464.

797 Swearer, S. E., Tremblay, E. A., Shima, J. S., 2019. *A review of biophysical models of marine larval dispersal*. CRC Press.

798 Tagliabue, A., Kwiatkowski, L., Bopp, L., Butensch  n, M., Cheung, W., Lengaigne, M., Vialard, J., 2021. Persistent uncertainties in ocean net

799 primary production climate change projections at regional scales raise challenges for assessing impacts on ecosystem services. *Front. Clim.* 3,

800 738224.

801 Takano, Y., Ito, T., Deutsch, C., 2018. Projected centennial oxygen trends and their attribution to distinct ocean climate forcings. *Global*

802 *Biogeochem. Cycles* 32, 1329–1349.

803 Tokarska, K. B., Stolpe, M. B., Sippel, S., Fischer, E. M., Smith, C. J., Lehner, F., Knutti, R., 2020. Past warming trend constrains future warming

804 in CMIP6 models. *Sci. Adv.* 6, eaaz9549.

805 Treguier, A. M., Held, I. M., Larichev, V. D., 1997. Parameterization of quasigeostrophic eddies in primitive equation ocean models. *J. Phys.*

806 *Oceanogr.* 27, 567–580.

807 Tschumi, T., Joos, F., Gehlen, M., Heinze, C., 2011. Deep ocean ventilation, carbon isotopes, marine sedimentation and the deglacial CO<sub>2</sub> rise.

808 *Clim. Past.* 7 (3), 771–800.

809 Vallis, G. K., 2006. *Atmospheric and Oceanic Fluid Dynamics*. Cambridge University Press.

810 Villani, C., 2008. *Optimal Transport: Old and New*. Springer.

811 Visbeck, M., Marshall, J., Haine, T., Spall, M., 1997. Specification of eddy transfer coefficients in coarse-resolution ocean circulation models. *J.*

812 *Phys. Oceanogr.* 27, 381–402.

813 Wang, Q., Danilov, S., Sidorenko, D., Timmermann, R., Wekerle, C., Wang, X., Jung, T., Schr  ter, J. G., 2014. The Finite Element Sea Ice-Ocean

814 Model (FESOM) v.1.4: formulation of an ocean general circulation model. *Geosci. Model Dev.* 7, 663–693.

815 Wang, Y., Stewart, A. L., 2020. Scalings for eddy buoyancy transfer across continental slopes under retrograde winds. *Ocean Modell.* 147, 101579.

816 Waterman, S., Hogg, N. G., Jayne, S. R., 2011. Eddy-mean flow interaction in the Kuroshio extension region. *J. Phys. Oceanogr.* 41, 1182–1208.

817 Waterman, S., Hoskins, B. J., 2013. Eddy shape, orientation, propagation, and mean flow feedback in western boundary current jets. *J. Phys.*

818 *Oceanogr.* 43, 1666–1690.

819 Waterman, S., Jayne, S. R., 2012. Eddy-driven recirculations from a localized transient forcing. *J. Phys. Oceanogr.* 42, 430–447.

820 Waterman, S., Lilly, J. M., 2015. Geometric decomposition of eddy feedbacks in barotropic systems. *J. Phys. Oceanogr.* 45, 1009–1024.

821 Wei, H., Wang, Y., Stewart, A. L., Mak, J., 2022. Scalings for eddy buoyancy fluxes across prograde shelf/slope fronts. *J. Adv. Model. Earth. Syst.*

822 14, e2022MS003229.

823 Whitt, D. B., Jansen, M. F., 2020. Slower nutrient stream suppresses Subarctic Atlantic Ocean biological productivity in global warming. *Proc.*

824 *Natl. Acad. Sci. USA* 117, 15504–15510.

825 Williams, R. G., Follows, M. J., 2011. *Ocean Dynamics and the Carbon Cycle: Principles and Mechanisms*. Cambridge University Press.

826 Williams, R. G., McDonagh, E., Roussenov, V. M., Torres-Valdes, S., Kind, B., Sanders, R., Hansell, D. A., 2011. Nutrient streams in the North

827 Atlantic: Advective pathways of inorganic and dissolved organic nutrients. *Global Biogeochem. Cycles* 25, GB4008.

828 Williams, R. G., Roussenov, V. M., Follows, M. J., 2017. Nutrient streams and their induction into the mixed layer. *Global Biogeochem. Cycles*

829 20, GB1016.

830 Zanna, L., Brankart, J. M., Huber, M., Leroux, S., Penduff, T., Williams, P. D., 2019a. Uncertainty and scale interactions in ocean ensembles: From

831 seasonal forecasts to multidecadal climate predictions. *Q. J. Roy. Met. Soc.* 145 (S1), 160–175.

832 Zanna, L., Khatiwala, S., Gregory, J. M., Ison, J., Heimbach, P., 2019b. Global reconstruction of historical ocean heat storage and transport. *Proc.*

833 *Natl. Acad. Sci. USA* 116, 1126–1131.

834 Zhang, Y., Vallis, G. K., 2013. Ocean heat uptake in eddying and non-eddying ocean circulation models in a warming climate. *J. Phys. Oceanogr.*

835 43, 2211–2229.