1 **MESMO 3: Flexible phytoplankton stoichiometry and refractory DOM**

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14   **Abstract**

15   We describe the third version of Minnesota Earth System Model for Ocean biogeochemistry

16   (MESMO 3), an earth system model of intermediate complexity, with a dynamical ocean, a

17   dynamic-thermodynamic sea ice, and an energy moisture balanced atmosphere. A major

18   feature of Version 3 is the flexible C:N:P ratio for the three phytoplankton functional types

19   represented in the model. The flexible stoichiometry is based on the power law formulation

20   with environmental dependence on phosphate, nitrate, temperature, and light. Other new

21   features include nitrogen fixation, water column denitrification, oxygen and temperature-

22   dependent organic matter remineralization, and $CaCO_3$ production based on the concept of

23   the residual nitrate potential growth. Also, we describe the semi-labile and refractory

24   dissolved organic pools of C, N, P, and Fe that can be enabled in MEMSO 3 as an optional

25   feature. The refractory dissolved organic matter can be degraded by photodegradation at

26   the surface and hydrothermal vent degradation at the bottom. These improvements

27   provide a basis for using MESMO 3 in further investigations of the global marine carbon

28   cycle to changes in the environmental conditions of the past, present, and future.

29

1. Introduction

Here we document the development of the third version of the Minnesota Earth System Model for Ocean biogeochemistry (MESMO 3). As described for the first two versions (Matsumoto et al., 2008, 2013), MESMO is based on the non-modular version of the Grid ENabled Integrated Earth (GENIE) system model (Lenton et al., 2006; Ridgwell et al., 2007). The computationally efficient ocean-climate model of Edwards and Marsh (Edwards and Marsh, 2005) forms the core of GENIE's physical model. MESMO is an earth system model of intermediate complexity (EMIC), which occupies a midpoint in the continuum of climate models that span high resolution, comprehensive coupled models on one end and box models at the other end (Claussen et al., 2002). MESMO has a 3D dynamical ocean model on a 36 x 36 equal-area horizontal grid with 10° increments in longitude and uniform in the sine of latitude. There are 16 vertical levels. It is coupled to a 2D energy moisture balanced model of the atmosphere and a 2D dynamic and thermodynamic model of sea ice. Thus, MESMO retains important dynamics, which allow for simulations of transient climate change, while still being computationally efficient.

Since the first version, MESMO has continued to be developed chiefly for investigations of ocean biogeochemistry (Table 1). Briefly, in MESMO 1, the main improvements over the predecessor GENIE focused on the biological production and remineralization as well as on the uptake of natural radiocarbon (^{14}C) and anthropogenic transient tracers (Matsumoto et al., 2008). The net primary production (NPP) in MESMO 1 occurred in the top two vertical levels representing the surface 100 m and depended on temperature, nutrients, light, and mixed layer depth (MLD). The nutrient dependence was based on the Michaelis-Menten uptake kinetics of phosphate (PO_4), nitrate (NO_3), and aqueous CO_2. The limiting nutrient was determined by the Liebig's rule of the minimum relative to the fixed uptake stoichiometry of C:N:P=117:16:1. A single generic phytoplankton functional type (PFT) carried out NPP, which was split between particulate organic matter (POM) and dissolved organic matter (DOM) in a globally constant ratio of 1:2. The semi-labile form of the dissolved organic carbon (DOC) was the only form of DOM simulated in MESMO 1. The POM flux across the 100 m level defined the export production. The vertical flux of POM was

60   driven by a fixed rate of sinking and a temperature-dependent, variable remineralization
61   rate.

62

63   The main aim of MESMO 2 was a credible representation of the marine silica cycle
64   (Matsumoto et al., 2013). To this end, the set of limiting nutrients (P, N, and C) in MESMO 1
65   was augmented to include iron (Fe) and silicic acid ($Si(OH)_4$) in MESMO 2 (Table 1). The
66   stable isotope of Si (^{30}Si) was also added as a state variable. The Fe cycle included an
67   aeolian flux of Fe, complexation with organic ligand, and particle scavenging of free Fe. The
68   scavenged Fe that reached the seafloor was removed from the model domain. This burial
69   flux of Fe balanced the aeolian flux at steady state. Also, a new PFT was added in MESMO 2
70   to represent chiefly diatoms. This new "large" PFT was limited by Si and characterized by a
71   high maximum growth rate and large half-saturation constants for the nutrient uptake
72   kinetics. It represented fast and opportunistic phytoplankton that do well under nutrient
73   replete conditions. In comparison, the "small" PFT was characterized with a lower
74   maximum growth rate and smaller half-saturation constants and outperformed the large
75   PFT in oligotrophic subtropical gyres. $CaCO_3$ production was associated with the "small"
76   PFT in MESMO2. The addition of Fe, Si, and the large PFT in MESMO 2 allowed it to have a
77   Fe-dependent, variable Si:N uptake ratio (Hutchins and Bruland, 1998; Takeda, 1998),
78   which is critical to simulate important features of the global ocean Si distribution.

79

80   MESMO 1 and 2 were assessed and calibrated by multi-objective tuning and extensive
81   model-data comparisons of transient tracers (anthropogenic carbon, CFCs), deep ocean
82   $\Delta^{14}C$, and nutrients (Matsumoto et al., 2008, 2013). These versions have been employed
83   successfully in a number of studies of global distributions of carbon and carbon isotopes
84   under various conditions of the past, present, and future (Cheng et al., 2018; Lee et al.,
85   2011; Matsumoto et al., 2010, 2020; Matsumoto and McNeil, 2012; Matsumoto and
86   Yokoyama, 2013; Sun and Matsumoto, 2010; Tanioka and Matsumoto, 2017; Ushie and
87   Matsumoto, 2012). Also, MESMO 1 and 2 have participated in model intercomparison
88   projects (Archer et al., 2009; Cao et al., 2009; Eby et al., 2013; Joos et al., 2013; Weaver et
89   al., 2012; Zickfeld et al., 2013).

90

91   In this contribution, we describe the third and latest version of MEMSO with a number of

92   substantial biogeochemical model modifications and new features that bring MESMO up to

93   date with the evolving and accumulating knowledge of the ocean biogeochemical cycle

94   (Table 1). There is no change in the physical model between MESMO 3 and MESMO 2. The

95   most significant new feature of MESMO 3 over the previous versions is the power law

96   formulation of flexible phytoplankton C:N:P ratio. Other new features include additional

97   PFT diazotrophs that carry out N-fixation, water column denitrification, the dependence of

98   organic matter remineralization on the dissolved oxygen (O_2) and temperature, and $CaCO_3$

99   production based on the concept of the residual nitrate potential growth. Also, we describe

100   the semi-labile DOM for P, N, and Fe (DOP_{sl}, DON_{sl}, and $DOFe_{sl}$), and the refractory DOM for

101   C, P, and N (DOC_r, DOP_r, and DON_r), which can be activated as an optional feature in MESMO

102   3. Some of these features have been described separately in different publications

103   (Matsumoto et al., 2020; Matsumoto and Tanioka, 2020; Tanioka and Matsumoto, 2017,

104   2020a). This work consolidates the descriptions of all these features in a single publication.

105

106   **2. Model Description**

107   Here we present the full set of biogeochemical equations of MESMO 3 as well as key model

108   parameters (Table 2). We describe only the biogeochemical source and sink terms and omit

109   the physical (advective and diffusive) transport terms that are calculated by the ocean

110   circulation model. We discuss the production terms first, followed by remineralization

111   terms, followed by conservation equations that incorporate both terms.

112

113   **2.1 Phytoplankton Nutrient Uptake**

114   NPP occurs in the top two vertical levels of the ocean domain above the fixed compensation

115   depth (z_c) of 100 m. Key parameter values are given in Table 2a. Nutrient uptake by

116   phytoplankton type i (Γ_i) depends on the optimal nutrient uptake timescale (τ_i), nutrients,

117   temperature (T), irradiance (I), and mixed layer depth (z_{ml}):

118

5

$$\Gamma_i = \frac{1}{\tau_i} \cdot F_{N,i} \cdot F_T \cdot F_I \cdot max\left\{1, \frac{z_c}{z_{ml}}\right\} \qquad 1$$

119

120   Subscript i refers to PFT (i =1: eukaryotes, i =2: cyanobacteria, i = 3: diazotrophs). The

121   nutrient dependence $F_{N,i}$ is given by Liebig's law of minimum combined with Michael-

122   Menten uptake kinetics of limiting nutrients: PO_4, NO_3, CO_2, (aq), total dissolved iron (sum

123   of free iron and ligand-bound iron; FeT=Fe'+FeL), and silicic acid ($Si(OH)_4$):

124

$$F_{N,i} = min\left(\frac{[PO_4]}{[PO_4]+K_{PO_4,i}} \cdot [PO_4], \frac{[NO_3]}{[NO_3]+K_{NO_3,i}} \cdot [NO_3] \cdot Q_{N,i}^{-1}, \frac{[CO_2(aq)]}{[CO_2(aq)]+K_{CO_2,i}} \cdot \right. \qquad 2$$

$$\left. [CO_2(aq)] \cdot Q_{C,i}^{-1}, \frac{[FeT]}{[FeT]+K_{FeT,i}} \cdot [FeT] \cdot Q_{Fe,i}^{-1}, \frac{[Si(OH)_4]}{[Si(OH)_4]+K_{Si(OH)_4}} \cdot [Si(OH)_4] \cdot Q_{Si}^{-1}\right),$$

125

126   where $K_{X,i}$ is the half-saturation concentration of nutrient X for PFT i. Only eukaryotes (i=1)

127   are limited by $Si(OH)_4$. Diazotrophs (i = 3) are not limited by NO_3. Nutrient uptake Γ is

128   based on the master nutrient variable P, and all other nutrient uptake is related to Γ by the

129   uptake stoichiometry $Q_{X,i}$, where X is N, Fe, Si, or C. For example, $Q_{C,i} = \frac{1}{[P:C]_i}$ for PFT i. Thus,

130    $Q_{c,i}$ is numerically equivalent to C:P for PFT i, but we write the equations in terms of P:C for

131   numerical stability and convenience. The $Q_{X,i}$ ratios represent the flexible phytoplankton

132   uptake stoichiometry and describe more fully in the following section 2.2.

133

134   The temperature dependence F_T of Equation 1 is given by:

135

$$F_T = \frac{T(^\circ C) + 2}{T(^\circ C) + 10} \qquad 3$$

136

137   which is analogous to the commonly used $Q_{10} = 2$ relationship. Light limitation F_I of

138   Equation 1 is described by a hyperbolic function:

6

$$F_T = \frac{I}{I + 20} \qquad\qquad 4$$

139

140 where I is the seasonally variable solar short-wave irradiance in W m^{-2}. Light is attenuated

141 exponentially from the ocean surface with a 20 m depth scale.

142

143 Nutrient uptake in Equation 1 has a dependence on z_{ml}, which is diagnosed using the σ_t

144 density gradient criterion (Levitus, 1982). Following the Sverdrup (1953) model of the

145 spring bloom, Equation 1 allows for the shoaling of z_{ml} relative to z_c to enhance nutrient

146 uptake.

147

148 **2.2 Phytoplankton uptake stoichiometry**

149 As noted above, all nutrients and O$_2$ are related to the main model currency P by $Q_{X,i}$. We

150 describe three different, mutually exclusive formulations in this section. The standard

151 formulation is the power law model (Matsumoto et al., 2020; Tanioka and Matsumoto,

152 2017). The other two (Linear model and Optimality-based model of stoichiometry) are

153 alternative formulations that have been coded, and the user can activate them (one at a

154 time) in place of the power law formulation. However, the alternative formulations are not

155 calibrated. Key parameter values are given in Table 2b for the power law formulation.

156

157 **2.2.1 Power law model of stoichiometry**

158 The uptake P:C and N:C ratios are calculated using the power-law formulation as a function

159 of ambient concentrations of phosphate [PO$_4$], nitrate [NO$_3$], temperature (T), and

160 Irradiance (I):

161

$$[P:C]_i = [P:C]_{0,i} \cdot \left(\frac{[PO_4]}{[PO_4]_0}\right)^{s_{PO_4,i}^{P:C}} \cdot \left(\frac{[NO_3]}{[NO_3]_0}\right)^{s_{NO_3,i}^{P:C}} \cdot \left(\frac{T}{T_0}\right)^{s_{T,i}^{P:C}} \cdot \left(\frac{I}{I_0}\right)^{s_{I,i}^{P:C}} \qquad 5$$

$$[N:C]_i = [N:C]_{0,i} \cdot \left(\frac{[PO_4]}{[PO_4]_0}\right)^{s_{PO_4,i}^{N:C}} \cdot \left(\frac{[NO_3]}{[NO_3]_0}\right)^{s_{NO_3,i}^{N:C}} \cdot \left(\frac{T}{T_0}\right)^{s_{T,i}^{N:C}} \cdot \left(\frac{I}{I_0}\right)^{s_{I,i}^{N:C}} \qquad 6$$

162

163   Equations 5 and 6 are the power-law equations that calculate the change in P:C and N:C for

164   fractional changes in environmental drivers relative to the reference P:C and N:C,

165   respectively (Matsumoto et al., 2020; Tanioka and Matsumoto, 2017). The exponents are

166   the sensitivity factors determined by a meta-analysis (Tanioka and Matsumoto, 2020a).

167   Subscript "0" indicates the reference values (Table 2b). We have hard bounds for the

168   calculated P:C and N:C ratios to be within 26.6<C:P<546.7 and 2<C:N<30 as observed

169   (Martiny et al., 2013).

170

171   The P:C and N:C ratios from Equations 5 and 6 can then be converted to $Q_{N,i}$ and $Q_{C,i}$ for use

172   in Equation 2.

173

$$Q_{C,i} = \frac{1}{[P:C]_i} \qquad 7$$

$$Q_{N,i} = \frac{1}{[P:N]_i} = \frac{[N:C]_i}{[P:C]_i} \qquad 8$$

174

175   **2.2.2 Linear model of stoichiometry by Galbraith & Martiny**

176   A much simpler, alternative formulation for P:C and N:C is the model of Galbraith & Martiny

177   (2015) where P:C is a linear function of [PO_4] (in μM), and N:C is a Holling type 2 functional

178   form with a frugality behavior only at very low [NO_3] (in μM). The same P:C and N:C values

179   are applied to all three PFTs.

180

$$[P{:}C] = \frac{6.9 \cdot [PO_4] + 6.0}{1000} \qquad\qquad 9$$

$$[N{:}C] = 0.125 + \frac{0.03 \cdot [NO_3]}{0.32 + [NO_3]} \qquad\qquad 10$$

181

182 **2.2.3 Optimality-based model of stoichiometry**

183 The optimality-based model of phytoplankton growth is based on the chain model, which

184 connects the cellular P, N, and C acquisition by a chain of limitations, where the P quota

185 limits N assimilation and the N quota drives carbon fixation (Pahlow et al., 2013; Pahlow

186 and Oschlies, 2009, 2013). Resource-allocation of cellular P, N, and C among different

187 cellular compartments are derived from balancing energy gain from gross carbon fixation

188 and energy loss due to nutrient acquisition and light-harvesting. The optimality-based

189 model by Pahlow et al. (2013) computes C:N and C:P as a function of nutrient availability

190 (PO$_4$ and NO$_3$), irradiance, and day length. Temperature dependence was added by

191 Arteaga et al. (2014) following the simple logarithmic temperature dependence on

192 maximum nutrient uptake rate following (Eppley, 1972).

193

194 Different versions of this optimality-based model have previously been successfully

195 implemented in global ocean biogeochemical models, such as the Pelagic Interactions

196 Scheme for Carbon and Ecosystem Studies (PISCES) (Kwiatkowski et al., 2018, 2019) and

197 the University of Victoria Earth System Model (UVIC) (Chien et al., 2020; Pahlow et al.,

198 2020). However, as we are not describing any results in this paper, we will only mention

199 here that there is an option to calculate C:N:P using this stoichiometry model in MESMO 3.

200 The full description of the optimality-based stoichiometry model and its parameter

201 calibration are presented specifically for the UVic model elsewhere (Chien et al., 2020;

202 Pahlow et al., 2020).

203

204 **2.2.4 Stoichiometry of iron and silica**

205 Iron uptake stoichiometry $Q_{Fe,i}$ is calculated as a function of FeT following the power-law

206 formulation of Ridgwell (2001). Key parameter values are given in Table 2c.

9

207

$$Q_{Fe,i} = [Fe:P]_i = [Fe:C]_i \cdot Q_{C,i} \qquad \qquad 11$$

$$[Fe:C]_i = 1.0/([C:Fe]_{min,i} + [C:Fe]_{ref,i} \cdot [FeT]^{-s^{Fe:C_i}}) \qquad 12$$

208

209   For all PFTs, the power law exponent $s^{Fe:C}$ in Equation 12 is -0.65. The allowable Fe:C ratio
210   is bounded at the low end by the hard-bound minimum Fe:C of 1:220,000. The scaling
211   constant or $[C:Fe]_{ref,i}$ is set differently for PFTs, with eukaryotes having a higher base
212    $[C:Fe]_{ref,i}$ than cyanobacteria and diazotrophs (115,623:1 and 31,805:1, respectively). The
213   high end of the allowable Fe:C ratio is bounded by $[C:Fe]_{min,i}$ (i.e., maximim Fe:C) of
214   15,000:1 for eukaryotes and 20,000:1 for cyanobacteria/diazotrophs. These parameters
215   directly follow Ridgwell (2001), who fitted power-law functions to the experimental data
216   (Sunda and Huntsman, 1995).

217

218   Silica uptake stoichiometry by eukaryotes Q_{Si} is a power law of total dissolved iron [FeT]
219   and increases with a decrease in [FeT] (Brzezinski, 2002). The power law exponent $s^{Si:N}$ is
220   set to 0.7. The Si:N ratio is limited to a maximum of 18 and a minimum of 1.

221

$$Q_{Si} = [Si:P] = [Si:N] \cdot Q_{N,1} \qquad \qquad 13$$

$$[Si:N] = min\left([Si:N]_{max}, max\left([Si:N]_{min}, \left(\frac{[FeT]}{0.5 \; nmol \; kg^{-1}} \right)^{-s^{Si:N}} \right) \right) \qquad 14$$

222

223   O$_2$ liberated by phytoplankton during photosynthesis per PO$_4$ consumed ($Q_{-O_2,i}$) is
224   calculated from the uptake C:P and N:P ratios (Tanioka and Matsumoto, 2020b):

225

$$Q_{-O_2,i} = 1.1Q_{C,i} + 2Q_{N,i} \qquad \qquad 15$$

226

227   **2.3 Production of POM and DOM**

10

In the top 100 m of the model domain, where phytoplankton P uptake occurs (i.e., $\Gamma_i > 0$, see section 2.1), NPP is immediately routed to POM and DOM pools (Figure 1). The production fluxes of POM, DOM_{sl}, and DOM_r from NPP are given as Jprod. Here we write the equations in terms of P, which is the master nutrient variable:

$$Jprod_{POP_i} = (1 - fDOM) \cdot \Gamma_i \qquad 16$$

$$Jprod_{DOP_{sl}} = \sum_i (1 - fDOM_r) \cdot fDOM \cdot \Gamma_i \qquad 17$$

$$Jprod_{DOP_r} = \sum_i fDOM_r \cdot fDOM \cdot \Gamma_i \qquad 18$$

The term fDOM denotes the fraction of NPP that is routed to DOM as opposed to POM. Likewise, $fDOM_r$ is the fraction of DOM that is routed to DOM_r as opposed to DOM_{sl}. The value of $fDOM_r$ is not well known but estimated to be ~1% (Hansell, 2013), which we tentatively adopt in MESMO 3. If DOM_r is not selected in the model run, $fDOM_r = 0$. In previous versions of MESMO, fDOM was assigned a constant value of 0.67. In reality, a large variability is observed locally for this ratio, ranging from 0.01-0.2 in temperate waters to 0.1-0.7 in the Southern Ocean (Dunne et al., 2005; Henson et al., 2011; Laws et al., 2000). In MESMO 3, fDOM is calculated as a function of the ambient temperature following Laws et al. (Laws et al., 2000):

$$fDOM = 1.0 - min\big(0.72, max\big(0.04, 0.62 - 0.02 \cdot T(°C)\big)\big) \qquad 19$$

This formulation gives low export efficiency (i.e., high fDOM) in the warmer regions compared to the colder high latitude regions. Locally, we impose fixed fDOM upper and lower bounds of 0.96 and 0.28, respectively, as estimated from a previous study (Dunne et al., 2005).

11

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253　In MEMSO 3, a new DOM production pathway below the production layer is available as an

254　option. In previous MESMO versions, sinking POM was respired in the water column with

255　the loss of O_2 directly to the dissolved inorganic forms (i.e., POC-->DIC, POP-->PO$_4$, and

256　POP-->NO$_3$). In the new "deep POC split" pathway, sinking POM is simply broken down into

257　DOM without the loss of O_2 <u>as in the production layer</u> (Figure 1). If DOM$_r$ is selected in the

258　model, the broken-down POM is further routed to both DOM$_{sl}$ and DOM$_r$ according to

259　fDOM$_r$. If not, all of the broken down POM is converted to DOM$_{sl}$. Thus, when the deep POC

260　split is activated, the presence of DOM in the deep ocean can be accounted for by *in situ*

261　production of DOM and DOM$_r$ in addition to DOM transport from the surface. Thus, the

262　deep POC split pathway offers an alternative means to control deep ocean DOM

263　distribution.

264

265　**2.4 Production of CaCO$_3$ and opal by eukaryotes**

266　In MESMO 2, opal production was associated with the large PFT and CaCO$_3$ production was

267　associated with the "small" PFT. We recognize that coccolithophorids and diatoms, which

268　are the producers of these biogenic tests, are both eukaryotes. Therefore, in MEMSO 3, we

269　associate both CaCO$_3$ and opal production with the POM production by the same eukaryote

270　PFT (Jprod$_{POM1}$) :

271

$$Jprod_{CaCO3} = r^{CaCO_3:POC} \cdot Jprod_{POM1} \cdot Q_{C,1} \qquad\qquad 20$$

$$Jprod_{opal} = Jprod_{POM1} \cdot Q_{Si} \qquad\qquad 21$$

272

273　The concept of the residual nitrate potential growth (RNPG) (Balch et al., 2016) is useful in

274　allowing competition between diatoms and non-siliceous phytoplankton within the same

275　PFT (Matsumoto et al., 2020). Typically, in the real ocean, non-Si phytoplankton are able to

276　grow faster and dominate the community if Si concentration is low and diatom growth is Si

277　limited. Otherwise, diatoms are more competitive, as they have higher intrinsic growth

278　rates. The RNPG index recasts the ambient concentrations of NO$_3$ and Si(OH)$_4$ into potential

279　algal growth rates:

280

$$RNPG = \frac{[NO_3]}{[NO_3] + K_{NO_3,1}} - \frac{[Si(OH)_4]}{[Si(OH)_4] + K_{Si(OH)_4}}$$ 22

281

282   If RNPG is more positive, the index indicates that nitrate-dependent growth exceeds silica-

283   dependent growth. Thus, non-Si phytoplankton are more competitive, and this leads to

284   higher $CaCO_3$ production. On the other hand, a more negative RNPG implies that silica

285   limitation for diatoms is relieved, leading to enhanced diatom growth and reduced $CaCO_3$

286   production. The RNPG index is incorporated in the calculation of the rain ratio $r^{CaCO_3:POC}$

287   presented in Equation 20 as:

288

$$r^{CaCO_3:POC} = r_0^{CaCO_3:POC} \cdot (\Omega - 1)^\eta \cdot min(1, max(0.1, RNPG)) \cdot k_{T,CaCO_3}$$ 23

289

290   Equation 23 indicates the base rain ratio $r_0^{CaCO_3:POC}$ (set to 0.30) is also modified by the

291   carbonate ion saturation state Ω by η (set to 1.28) by as well as by temperature (see

292   Ridgwell et al. (2007) and references therein):

293

$$\Omega = \frac{[Ca^{2+}][CO_3^{2-}]}{K_{sp}}$$ 24

$$k_{T,CaCO_3} = min\left(1.0, \frac{T(^\circ C) + 2}{T(^\circ C) + 8}\right)$$ 25

294

295   K_{sp} is the solubility product of $CaCO_3$. The temperature dependency of $CaCO_3$ formation

296   ($k_{T,CaCO_3}$) is similar to that of Moore et al. (2004) where warmer temperatures favor the

297   growth of carbonate-bearing phytoplankton.

298

299   **2.5 Remineralization of POM and DOM**

300   Once produced, both POM and DOM undergo remineralization throughout the water

301   column. Key remineralization parameter values are given in Table 2d. Previously, POM

13

304   remineralization had a temperature dependence and decayed exponentially with depth

305   (Yamanaka et al., 2004). In MESMO 3, we incorporate an additional dependency on

306   dissolved oxygen following Laufkötter et al. (2017):

307

$$R_{POM_i} = V_{POM} \cdot e^{k_R \cdot T \cdot \frac{[O_2]}{[O_2] + K_{O_2}}} \cdot [POM_i] \qquad 26$$

308

309   V_{POM} is the base remineralization rate, parameter k_R expresses the temperature sensitivity

310   of remineralization, and K_{O_2} is half-saturation constant for oxygen-dependent

311   remineralization. When the sediment model is not coupled, any POM that reaches the

312   seafloor dissolves completely to its inorganic form and is returned to the overlying water.

313

314   In MESMO 3, all forms of semi-labile DOM remineralize at the same rate. It is represented

315   by τ_{sl}, the inverse of the time scale of DOM_{sl} decay, which has been estimated previously to

316   be ~1.5 years (Hansell, 2013):

317

$$R_{DOM_{sl}} = \tau_{sl} \cdot [DOM_{sl}] \qquad 27$$

318

319   All forms of DOM_r also remineralize at the same rate in MESMO 3. In total, there are three

320   optional, additive sinks of DOM_r in the model: slow background decay, photodegradation,

321   and degradation via hydrothermal vents (Figure 1). Observations clearly indicate that the

322   [14]C age of deep ocean DOC_r is 10^3 years (e.g., Druffel et al., 1992), much older than DI[14]C.

323   Also, the deep ocean DOC_r concentration decreases modestly along the path of the deep

324   water from the deep North Atlantic to the deep North Pacific (Hansell and Carlson, 1998).

325   Thus, it is understood that there is a slow DOM_r background decay in the deep ocean. We

326   represent this <u>ubiquitous</u> process with τ_{bg}, which is the inverse of the background decay

327   time scale, estimated to be ~16,000 years (Hansell, 2013).

328

329   Observations to date indicate that photodegradation is a major sink of DOM_r (e.g., Mopper

330   et al., 1991). This process is believed to convert DOM_r that is upwelled from the ocean

14

331   interior into the euphotic zone into more labile forms of DOM. We represent

332   photodegradation with τ_{photo}, the inverse of the decay time scale, estimated to be ∼70 years

333   (Yamanaka and Tajika, 1997). <u>This occurs only in the surface.</u>

334

335   Finally, observations of DOM emanating from different types of hydrothermal vents

336   indicate that they have variable impacts on the deep sea DOM_r (Lang et al., 2006). However,

337   the off-axis vents circulate far more seawater through the fractured oceanic crust than the

338   high temperature and diffuse vents and thus believed to determine the overall impact of

339   the vents on the deep sea DOM_r as a net sink (Lang et al., 2006). Here we assume simply

340   that seawater that circulates through the vents loses all DOM_r (i.e., $1/\tau_{vent} < \Delta t$, where Δt is

341   the biogeochemical model time step of 0.05 year). This means that the more seawater

342   circulates through the vents, the more DOM_r is removed: the total removal rate depends on

343   the vent flux of seawater H_{flux}. We implement the vent degradation of DOM_r in MESMO 3 by

344   first identifying the wet grid boxes located immediately above known mid-ocean ridges. We

345   then distribute the annual global H_{flux} of 4.8×10^{16} kg yr^{-1} (Lang et al., 2006) equally among

346   those ridge-associated grid boxes. The grid cells contain a mass of seawater much greater

347   than the mass that circulates through vents in Δt (10^{21} kg vs. 10^{13} kg). Therefore, the

348   seawater mass in the vent grid cells that does not circulate through the vents in Δt is

349   subject only to background degradation in MESMO 3.

350

351   The three DOM_r sinks are not mutually exclusive. They can thus be combined to yield the

352   total DOM_r remineralization rate:

353

$$R_{DOM_r} = \left(\tau_{bg} + \tau_{photo} + \tau_{vent} \cdot \frac{SW_{flux\_local}}{SW_{grid}} \right) \cdot [DOM_r] \qquad 28$$

354

355   where SW_{flux_local} is the mass of seawater that circulates through the vents in each grid box

356   in Δt, and SW_{grid} is the total mass of seawater in the same grid box.

357

15

358 The amount of O_2 respired as a result of these POM and DOM remineralization processes is
359 related to the organic carbon pools by the respiratory quotients of POC and DOC, $r_{-O_2:POC}$
360 and $r_{-O_2:DOC}$, respectively. These are molar ratios of O_2 consumed per unit organic carbon
361 respired. They are variable and calculated from the ambient POM and DOM concentration
362 (Tanioka and Matsumoto, 2020b):
363

$$r_{-O_2:POC} = 1.1 + \frac{2[PON]}{[POC]} \qquad\qquad 29$$

$$r_{-O_2:DOC} = 1.1 + \frac{2[DON]}{[DOC]} \qquad\qquad 30$$

364
365 **2.6 Remineralization of CaCO$_3$ and opal**
366 Remineralization of CaCO$_3$ and opal particles occurs as they sink through the water column
367 and remains the same as in MESMO 2. Key parameter values are given in Table 2d.
368 Remineralization of CaCO$_3$ is a function of temperature similar to that of particulate organic
369 matter remineralization but without oxygen dependency. The temperature dependence
370 term k_R modifies the base remineralization rate V_{CaCO_3}:
371

$$R_{CaCO_3} = V_{CaCO_3} \cdot e^{k_R \cdot T} \cdot [CaCO_3] \qquad\qquad 31$$

372
373 Opal remineralization in the water column follows Ridgwell et al. (2002). The rate of opal
374 remineralization R_{opal} is given by the product of normalized dissolution rate (r_{opal}), base
375 opal dissolution rate (k_{opal}), and opal concentration [opal]:
376

$$R_{opal} = r_{opal} \cdot k_{opal} \cdot [opal] \qquad\qquad 32$$

$$r_{opal} = 0.16 \cdot \left(1 + \frac{T(°C)}{15}\right) \cdot u_{opal} + 0.55 \cdot \left(\left(1 + \frac{T(°C)}{400}\right)^{4} \cdot u_{opal}\right)^{9.25} \qquad 33$$

$$u_{opal} = \frac{[Si(OH)_4]_{eq} - [Si(OH)_4]}{[Si(OH)_4]_{eq}} \qquad\qquad 34$$

377

378   r_{opal} is a function of temperature (T) and the degree of under-saturation (u_{opal}), which in

379   turn is calculated from the ambient $[Si(OH)_4]$ and $[Si(OH)_4]$ at equilibrium. The

380   equilibrium concentration is a function of ambient temperature:

381

$$log_{10}\left([Si(OH)_4]_{eq}\right) = 6.44 - \frac{968}{T(K)} \qquad\qquad 35$$

382

383   Without the sediment module of MESMO activated, both $CaCO_3$ and opal particles that

384   reach the seafloor are completely dissolved back to inorganic forms.

385

386   **2.7 Conservation of organic matter and biogenic tests**

387   The time rate of change of the biogenic organic matter and tests are given by the sum of the

388   production terms (i.e., sources) and the remineralization terms (i.e., sinks). The circulation-

389   related transport terms are omitted as noted above, but the vertical transport due to

390   particle sinking is included here. The sinking speed w is the same for all particles. The sum

391   of POM_i of all the PFTs give the total POM concentrations:

392

17

$$\frac{\partial [POP]_i}{\partial t} = Jprod_{POP_i} - \frac{\partial}{\partial z}(w[POP]_i) - R_{POP,i} \qquad 36$$

$$\frac{\partial [POC]_i}{\partial t} = Jprod_{POP_i} \cdot Q_{C,i} - \frac{\partial}{\partial z}(w[POC]_i) - R_{POC,i} \qquad 37$$

$$\frac{\partial [PON]_i}{\partial t} = Jprod_{POP_i} \cdot Q_{N,i} - \frac{\partial}{\partial z}(w[PON]_i) - R_{PON,i} \qquad 38$$

$$\frac{\partial [POFe]_i}{\partial t} = Jprod_{POP_i} \cdot Q_{Fe,i} - \frac{\partial}{\partial z}(w[POFe]_i) - R_{POFe,i} \qquad 39$$

$$[POM] = \sum_i [POM]_i \qquad 40$$

393

394 The time rate of change of CaCO₃ and opal is expressed in much the same way as POM:

395

$$\frac{\partial [CaCO_3]}{\partial t} = Jprod_{CaCO3} - \frac{\partial}{\partial z}(w[CaCO_3]) - R_{CaCO_3} \qquad 41$$

$$\frac{\partial [opal]}{\partial t} = Jprod_{opal} - \frac{\partial}{\partial z}(w[opal]) - R_{opal} \qquad 42$$

396

397 The DOM pools have the production and remineralization terms without the particle

398 sinking term:

399

$$\frac{\partial [DOP_{sl}]}{\partial t} = Jprod_{DOP_{sl}} - R_{DOP_{sl}} \tag{43}$$

$$\frac{\partial [DON_{sl}]}{\partial t} = Jprod_{DON_{sl}} - R_{DON_{sl}} \tag{44}$$

$$\frac{\partial [DOC_{sl}]}{\partial t} = Jprod_{DOC_{sl}} - R_{DOC_{sl}} \tag{45}$$

$$\frac{\partial [DOFe_{sl}]}{\partial t} = Jprod_{DOFe_{sl}} - R_{DOFe_{sl}} \tag{46}$$

$$\frac{\partial [DOP_{r}]}{\partial t} = Jprod_{DOP_{r}} - R_{DOP_{r}} \tag{47}$$

$$\frac{\partial [DON_{r}]}{\partial t} = Jprod_{DON_{r}} - R_{DON_{r}} \tag{48}$$

$$\frac{\partial [DOC_{r}]}{\partial t} = Jprod_{DOC_{r}} - R_{DOC_{r}} \tag{49}$$

400

2.8 Conservation of inorganic nutrients

The time rate of change of the inorganic nutrients have organic carbon production as sink terms and remineralization as source terms. The production terms (J_{prod}) are zero below the upper ocean production layer. Nutrients have a unit of mol element kg^{-1} in the model.

405

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$$\frac{\partial [PO_4]}{\partial t} = -\sum_i \Gamma_i + \sum_i R_{POP,i} + R_{DOP_{sl}} + R_{DOP_r} \tag{50}$$

$$\frac{\partial [NO_3]}{\partial t} = -\sum_i \Gamma_i \cdot Q_{N,i} + \sum_i R_{PON,i} + R_{DON_{sl}} + R_{DON_r} + Fix_N - Den_N \tag{51}$$

$$\frac{\partial [DIC]}{\partial t} = -\left(\sum_i \Gamma_i Q_{C,i} + Jprod_{CaCO3}\right) + \sum_i R_{POC,i} + R_{DOC_{sl}} + R_{DOC_r} + R_{CaCO_3} \\ + F_{gas,CO_2} \tag{52}$$

$$\frac{\partial [ALK]}{\partial t} = -\left(2 \cdot Jprod_{CaCO3} - \sum_i \Gamma_i Q_{N,i}\right) - \sum_i R_{PON,i} - R_{DON_{sl}} - R_{DON_r} - Fix_N \\ + Den_N + 2 \cdot R_{CaCO_3} \tag{53}$$

$$\frac{\partial [FeT]}{\partial t} = -\sum_i \Gamma_i Q_{Fe,i} + \sum_i R_{POFe,i} + R_{DOFe_{sl}} + R_{POM_{Fe}} + Aeolian_{Fe} \tag{54}$$

$$\frac{\partial [Si(OH)_4]}{\partial t} = -Jprod_{opal} + R_{opal} \tag{55}$$

$$\frac{\partial [O_2]}{\partial t} = \sum_i \Gamma_i \cdot Q_{-O_2,i} - \left(r_{-O_2:DOC} \cdot \left(R_{DOC_{sl}} + R_{DOC_r}\right) + \sum_i r_{-O_2:POC,i} \cdot R_{POC,i}\right) \\ + 1.25\, Den_N + F_{gas,O_2} \tag{56}$$

409

410   In Equation 51, Fix_N is the N-fixation carried out by diazotrophs, and Den_N is the water

411   column denitrification. There is an air-sea gas exchange term F_{gas} in Equations 52 and 56 for

412   gaseous CO_2 and O_2, respectively. In Equation 53, alkalinity increases with decreasing

413   nitrate concentrations and increasing $CaCO_3$ dissolution. Equation 54 contains R_{POMFe},

414   which is an iron source that represents remineralization of the Fe' scavenged by sinking

415   particles. These terms are explained in the following sections.

416

417   **2.9 Prognostic nitrogen cycle**

418 Biological production by diazotrophs is stimulated when the ambient NO_3 is low. Nitrogen
419 fixed by diazotrophs during their growth is added to the marine NO_3 pool. The prognostic
420 nitrogen fixation model employed here is similar to that used in the HAMOCC
421 biogeochemical module (Paulsen et al., 2017):
422

$$Fix_N = \Gamma_3 \cdot Q_{N,3} \cdot I_{NO_3} \, , \qquad 57$$

$$I_{NO_3} = \left(1.0 - \frac{[NO_3]^2}{K_{N_2}^{\,2} + [NO_3]^2} \right), \qquad 58$$

423
424 where Fix_N is the nitrogen fixation rate and I_{NO_3} is the nitrate dependency term in
425 quadratic Michaelis-Menten kinetics form with the half-saturation constant K_{N_2}. See Table
426 2e for the values related to the N cycle.
427
428 Water-column denitrification is formulated in an approach similar to that of the original
429 GENIE model (Ridgwell et al., 2007), in which 2 moles of NO_3 are converted to 1 mole of N_2
430 and liberating 2.5 moles of O_2 as a byproduct:
431

$$2NO_3^- + 2H^+ \rightarrow 2.5O_2 + N_2 + H_2O \qquad 59$$

432
433 Denitrification takes place in grid boxes, in which O_2 concentration is below a threshold
434 concentration ($O_{2,def}$) and is stimulated if the total global inventory of NO_3 relative to PO_4 is
435 high. In other words, denitrification can effectively act as negative feedback to nitrogen
436 fixation. The threshold O_2 concentration ($O_{2,def}$) takes the minimum of the hard-bound O_2
437 threshold concentration ($O_{2,crit}$) and the NO_3 to PO_4 ratio, scaled by a parameter k_D. The
438 parameters $O_{2,crit}$ and k_D are calibrated to give the global denitrification rate of roughly
439 100 Tg N yr^{-1}, which balances the total nitrogen fixation rate in the model.
440

21

$$Den_N = 0.8 \, [yr^{-1}] \cdot max \left(([O_2]_{def} - [O_2]), 0 \right) \qquad 60$$

$$[O_2]_{def} = min \left(O_{2,crit}, k_D \cdot \frac{[NO_3]_{inventory}}{[PO_4]_{inventory}} \right) \qquad 61$$

441

442 **2.10 Prognostic iron cycle**

443 The iron cycle in MESMO 3 remains the same as in MESMO 2. Key parameter values are

444 given in Table 2e. The two species of dissolved iron (Fe' and FeL) are partitioned according

445 to the following equilibrium relationship:

446

$$K_{ligand} = \frac{[FeL]}{[Fe'] \cdot [L]} \qquad 62$$

447

448 where [L] is the ligand concentration, and K_{ligand} is the conditional stability constant. The

449 sum of ligand and FeL is set at a constant value of 1 nmol kg^{-1} everywhere. Iron is

450 introduced into the model domain by a constant fraction (3.5 weight %) of aeolian dust

451 deposition at the surface (F_{in}) following the prescribed modern flux pattern (Mahowald et

452 al., 2006) with constant solubility (β):

453

$$S_{Fe} = \beta \cdot F_{in} \qquad 63$$

454

455 Particle-scavenged iron POM$_{Fe}$ (note the difference from POFe) is produced below the

456 productive layer when sinking POM scavenges Fe' to sinking POM:

457

$$J_{Fe} = -\tau_{sc} \cdot K_o \cdot [POC]^{0.58} \cdot [Fe'] \qquad 64$$

458

459 where τ_{sc} and K_o and are empirical parameters that determine the strength of scavenging.

460 Remineralization of Fe scavenged to POM (POM$_{Fe}$) is identical in form to that of POM

461 remineralization:

22

462

$$R_{POM_{Fe}} = V_{POM} \cdot e^{k_R T \cdot \frac{[O_2]}{[O_2] + K_{O_2}}} \cdot [POM_{Fe}]$$ 65

463

464   The conservation equation of the particle scavenged iron is thus expressed as :

465

$$\frac{\partial [POM_{Fe}]}{\partial t} = J_{Fe} - \frac{\partial}{\partial z}(w[POM_{Fe}]) - R_{POM_{Fe}}$$ 66

466

467   Any scavenged iron that escapes remineralization in the water column reaching the

468   seafloor is removed from the model domain in order to keep the total Fe inventory at a

469   steady state.

470

471   **2.11 Air-sea gas Exchange**

472   The air-sea gas exchange formulation remains the same as in MESMO 2 and follows

473   Ridgwell et al. (2007). It is the function of gas transfer velocity, the ambient dissolved gas

474   concentration, and saturation gas concentration. The flux of CO_2 and O_2 gases across the

475   air-sea interface is given by:

476

$$F_{gas,CO_2} = k \cdot ([CO_2]_{sat} - [CO_2]) \cdot (1 - A)$$ 67   **Deleted:** $\cdot \rho$

$$F_{gas,O_2} = k \cdot ([O_2]_{sat} - [O_2]) \cdot (1 - A)$$ 68   **Deleted:** $\cdot \rho$

477

478   where k is the gas transfer velocity, $[CO_2]_{sat}$ and $[O_2]_{sat}$ are saturation concentrations, and   **Deleted:** ρ is the density of seawater,

479   A is the fractional ice-covered area that is calculated by the physical model. Gas transfer

480   velocity k is a function of wind speed (u) following Wanninkhof (1992) where Sc is the

481   Schmidt Number for a specific gas:

482

$$k = 0.31 \cdot u^2 \cdot \left(\frac{Sc}{660}\right)^{-0.5} \qquad \qquad \textit{69}$$

486

3 Results and Discussion

488 All new results from MESMO 3 presented here are from the steady state. On a single

489 compute core at the Minnesota Supercomputing Institute, it takes approximately 1 hour to

490 complete 1,000 years of MESMO 3 simulation. The "standard" MESMO 3 has the power law

491 model of flexible stoichiometry but no DOM_r. The results from the standard model

492 (hereafter just MESMO 3) are presented in Section 3.1, and the results from the DOM_r-

493 enabled model are presented in Section 3.2. In Table 3, we summarize and compare key

494 biogeochemical diagnostics from MESMO 3 against those from MESMO 2 and available

495 observational constraints. The global NPP, as well as global export production of POC and

496 opal, are comparable or somewhat lower in MESMO 3 than MESMO 2.

497

498 We relied on experience to calibrate MESMO 3 with the primary goal of reasonably

499 simulating the phytoplankton community composition and C:N:P ratio (e.g., abundant

500 cyanobacteria and high ratio for all PFTs in oligotrophic gyres). We tried to improve or at

501 least preserve the gains that we achieved in earlier versions of MESMO in terms of the

502 global distributions of PO_4, NO_3, O_2, $Si(OH)_4$, and FeT (Supplemental Figures S1, S2, S3, S4,

503 and S5). Overall, there is a stronger nutrient depletion in the new model. For example, the

504 surface PO_4 and NO_3 in MESMO 3 are now sufficiently depleted in the subtropical gyres but

505 too low in the eastern equatorial Pacific when compared to the World Ocean Atlas (Figure

506 S1; see RMSE in Table 3). It is a challenge for MESMO and other coarse resolution models to

507 simulate narrow dynamical features such as equatorial upwelling and reproduce

508 biogeochemical features with sharp gradients. The spatial pattern of POC export that drives

509 this surface nutrient pattern is similar in the two models (Figure S2). In the 1D global

510 profile, there is a marked improvement in the subsurface distribution of O_2 in MESMO 3

511 over MESMO 2. Whereas the depth of the oxygen minimum was ~300 m in MEMOS 2, it is

512 ~1000 m in both MESMO 3 and the World Ocean Atlas (Figure S3). At 1000 m, the O_2

513 minimum is located in the far North Pacific in MESMO 3, whereas in the World Ocean Atlas

530   it occurs in both the Northeast Pacific and the Arabian Sea. In contrast, the world ocean at

531   1000 m is too well-oxygenated in MESMO 2. We believe that the improved match in the O_2

532   minimum depth would help simulate denitrification in the correct depth range, and there is

533   a modest improvement in the data-model O_2 mismatch in terms of RMSE in MESMO 3 over

534   MESMO 2 (Table 3). The deepening of the O_2 minimum was achieved largely by increasing

535   the particle sinking speed, which tends to strengthen the biological pump and deplete the

536   surface nutrients. This also helps MESMO 3 preserves MESMO 2's surface $Si(OH)_4$ depletion

537   in much of the world ocean except in the North Pacific and Southern Ocean (Figure S4).

538   This is a feature captured by $Si^* < 0$ ($Si^* = [Si(OH)_4] - [NO_3]$) in observations (Sarmiento et al.,

539   2004) and simulated previously by MESMO 2 and now MESMO 3. Finally, surface FeT is also

540   depleted more strongly in MESMO 3 over MESMO2, except the North Atlantic, where

541   aeolian deposition of dust from the Sahara maintains a steady Fe supply (Figure S5).

542

543   In MESMO 3, we made no effort to calibrate all the semi-labile pools of DOM: DOC_{sl}, DOP_{sl},

544   DON_{sl}, and $DOFe_{sl}$. We note only that the surface DOC_{sl} concentration of 58 µmol kg^{-1} and

545   DOC export production of 1.4 Pg C yr^{-1} in MESMO 3 are higher than in MESMO 2 (24 µmol

546   kg^{-1} and 0.4 Pg C yr^{-1}, respectively). The higher surface concentration is due to the longer τ_{sl}

547   in MESMO 3 (Table 2d). The global average of the temperature dependent fDOM in MESMO

548   3 is 0.69, which is slightly higher than the spatially uniform value of 0.67 in MESMO 2.

549

550   **3.1 Novel features of MESMO 3**

551   An important new feature of MESMO 3 is the representation of the primary producers by

552   three PFTs (Figure 2). The eukaryotes are characterized by the highest maximum growth

553   rate and high half-saturation constants. Thus, the eukaryotes are more dominant than the

554   other PFTs in the more eutrophic waters of the equatorial and polar regions (Figure 2a).

555   The cyanobacteria have smaller half-saturation constants and thus are more dominant in

556   the oligotrophic subtropical gyres (Figure 2c). The diazotrophs do not have NO_3 limitation

557   but have the lowest maximum growth rate. Thus it is much lower in abundance than the

558   other two PFTs generally, and outcompeted in transient blooms and thus excluded in

559   higher latitudes (Figure 2e).

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25

568

569 Figure 2 also indicates that all three PFTs show Fe limitation in the Southern Ocean.

570 Outside the Southern Ocean, the eukaryotes are primarily limited by $Si(OH)_4$ (Figure 2b).

571 As far as organic carbon is concerned, we consider the eukaryotes to basically represent

572 diatoms, which are arguably the most important agent of organic carbon export. In this

573 context, the widespread silica limitation for eukaryotes would be consistent with the

574 notions that silica uptake by diatoms should be limited in ~60% of the world surface ocean

575 (Ragueneau et al. 2000) and that much the world ocean thermocline is filled with silica-

576 depleted water (Si*<0 as noted above). On the other hand, the cyanobacteria are largely

577 limited by NO_3 outside the Southern Ocean (Figure 2d). The diazotrophs are limited by iron

578 in much of the world ocean except in the Atlantic basin (Figure 2f), where surface PO_4 is

579 strongly depleted in both observations (Mather et al., 2008) and in our model (Figure S1).

580

581 Figure 3 illustrates the influence of the RNPG index, which was implemented in MESMO 3

582 to allow for the effect of competition between diatoms and coccolithophores within the

583 same PFT (Equations 22 and 23). The eukaryote NPP (Figure 3a) is effectively split into two

584 parts: one is associated with diatoms and opal production (Figure 3b), and the other is

585 associated with coccolithophores and $CaCO_3$ production (Figure 3c). According to the RNPG

586 index, opal production is simulated more in the higher latitudes of the Southern Ocean and

587 the North Pacific, where surface $[Si(OH)_4]$ is abundant. Elsewhere, $CaCO_3$ production is

588 relatively larger. The decoupling is prominent in the North Indian. Note that the spatial

589 pattern of $CaCO_3$ production is quite different in MEMOS 3 (Figure 3c) and MESMO 2

590 (Figure 3d), because $CaCO_3$ production was associated in MESMO 2 with the "small" PFT,

591 which corresponds to the cyanobacteria PFT in MESMO 3.

592

593 The global pattern of the mean C:P uptake ratio in the production layer is shown in Figure

594 4. Consistent with observations (Martiny et al., 2013), the simulated C:P ratio of the

595 phytoplankton community is elevated in the oligotrophic subtropical gyres and low in the

596 eutrophic polar waters (Figure 4a). The community C:P ratio exceeds 200 in the gyres and

597 reaches as low as 40-50 in the Southern Ocean. The community C:P has contributions from

598 both physiological effects (i.e., environmental drivers acting on each PFT's C:P ratio) and

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604 taxonomic effects (i.e., the shift in the community composition changes the weighting of
605 each PFT's C:P ratio). Figure 4b shows that the community C:P is high in oligotrophic gyres
606 because cyanobacteria and to a lesser extent diazotrophs dominate the community, and
607 their C:P ratio is high. Conversely, the community C:P is low in the polar waters because the
608 eukaryotes dominate and their C:P ratio is low. For both eukaryotes and cyanobacteria,
609 their C:P is high in oligotrophic subtropical gyres because PO_4 is low (Figure 4c and d). This
610 physiological effect is larger in eukaryotes than cyanobacteria because the former has
611 greater sensitivity (i.e., larger sensitivity factor $s_{PO_4}^{P:C}$, Equation 5 , Table 2b). However, the
612 cyanobacteria PFT's C:P ratio has an additional sensitivity to temperature (i.e., $s_T^{P:C} \neq 0$)
613 that elevates their C:P in the lower latitudes. We do not show the C:P ratio for diazotrophs
614 because it is very similar to that of cyanobacteria (Figure 4b, d).
615
616 In order to gain more insights into the spatial patterns of the C:P ratio (Figure 4), we
617 examined the relationships between the C:P and C:N ratios and the four possible
618 environmental drivers for eukaryotes and cyanobacteria (Figure 5; again, diazotrophs are
619 not shown). The red plots show that there is a causal relationship between the ratios and
620 the drivers as formulated by the power law model (Equations 5 and 6). The black plots
621 show the absence of a causal relationship. For example, the C:P ratio of both eukaryotes
622 and cyanobacteria are strongly correlated with PO_4 because there is a causal relationship
623 (Figure 5a, b shown in red). Similarly, the C:N ratio of the same two PFTs have a strong
624 correlation with PO_4 (Figure 5c, d in black), but there is actually not a causal relationship
625 (i.e., $s_{PO_4}^{N:C} = 0$, Table 2b). The C:N-PO_4 correlation exists, simply because the nutrients are
626 well correlated. Similarly, because temperature and photosynthetically active radiation
627 (PAR) tend to be correlated via latitude, the stoichiometry has a similar correlation to the
628 two drivers. For example, cyanobacteria C:P has a strong correlation with both
629 temperature and PAR (Figure 5j, 4n), but only the temperature is a real driver. Figure 5
630 indicates which are the dominant drivers of the C:N:P ratio in MESMO 3. For the eukaryote
631 C:P ratio, it is PO_4. For the cyanobacteria C:P ratio, the important drivers are temperature
632 and PO_4. For the C:N ratio for both eukaryotes and cyanobacteria, NO_3 is more important
633 than PAR. Figure 5 also serves to remind us that correlation does not indicate causation.

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637

638   Figure 6 shows the community C:P and C:N ratios plotted against the four environmental

639   drivers. Unlike Figure 5, which reflected the individual PFT's physiological response, Figure

640   6 includes the effect of taxonomy as well. Still, the effects of PO_4 and temperature are

641   clearly visible on the community C:P ratio. Both low $[PO_4]$ and warmer waters are found in

642   the lower latitudes, so the P frugality and temperature effects are additive. The effect of

643   NO_3 on the community C:N ratio is also very clear, but the effect of PAR is not as clear. Thus

644   overall, the physiological effects seen in the PFT-specific C:N:P are obvious in the

645   community C:N:P ratio.

646

647   **3.2 DOM_r-enabled MESMO 3**

648   In MESMO 2, DOC_{sl} was a standard state variable. In MESMO 3, other forms of DOM are

649   available as options. They are the semi-labile forms of DOM: DOP_{sl}, DON_{sl}, and $DOFe_{sl}$; and

650   the refractory forms of DOM: DOC_r, DOP_r, and DON_r. MESMO 3 is not yet calibrated with

651   respect to all the DOM variables, but here we demonstrate their potential use in future

652   biogeochemical investigations by presenting steady state DOM results from the model

653   experiment *LV* (experiment ID: 210310m). In this run, all three sinks of DOM_r are activated:

654   slow background decay, photodegradation, and degradation in hydrothermal vents.

655

656   The experiment name *LV* stands for "literature values." In *LV*, we use the literature values

657   for the key DOM remineralization model parameters (Table 2d) and $fDOM_r$ = 0.01 (Hansell,

658   2013). All other model parameter values in the *LV* run are identical to the standard MESMO

659   3 model (Table 2). The black lines in Figure 7 show the global mean vertical profiles of the

660   total DOC ($DOC_t = DOC_{sl} + DOC_r$) in solid line and DOC_r in dashed line. Qualitatively, the

661   simulated profiles are consistent with the observations, showing a near-uniform DOC_r

662   concentration and a DOC_{sl} profile that rapidly with depth in the top few hundred meters

663   (Hansell, 2013). However, the simulated values reach 130 μmol kg^{-1} in the surface, which is

664   approximately twice the observations. More typically, the observed DOC_r is 30~40 μmol kg^{-1}

665   ^{1}, and the observed DOC_{sl} attenuates with depth from 30~40 μmol kg^{-1} near the surface. So

28

666   their sum, which is represented by DOC_t, is approximately 60-80 μmol kg^{-1} at the surface in

667   observations.

668

669   Figure 8 adds a lateral perspective to Figure 7. The rapid DOC_t attenuation in the vertical is

670   strong in the lower latitudes where stratification is generally stronger. The transport of

671   DOC_{sl} from the surface to deeper waters is evident in the high latitudes of the North

672   Atlantic and the Southern Ocean. The DOC_t change in the deep ocean is limited.

673   Observations of deep ocean DOC_t indicates a reduction by 29% or 14 μmol kg^{-1} from the

674   deep North Atlantic to the deep North Pacific (Hansell and Carlson, 1998). Figure 8 shows

675   that the deep ocean DOC_t gradient in *LV* is approximately 10 μmol kg^{-1} from 70-75 μmol kg$^-$

676   ^{1} in the North Atlantic to <65 μmol kg^{-1} in the North Pacific.

677

678   The horizontal DOC_t distributions from the *LV* run can also be compared to a global

679   extrapolation based on an artificial neural network (ANN) of the available DOC_t data

680   (Roshan and DeVries, 2017). At the surface, the extrapolation indicates higher DOC_t

681   concentrations in the subtropical gyres (Figure 9a), while our simulation does not clearly

682   delineate the gyres (Figure 9c). In our model, fDOM is temperature-dependent and strongly

683   controls the production of DOM. The surface DOC_t is thus more elevated in the lower

684   latitudes. Interestingly, the ANN study diagnosed higher rates of DOM production in the

685   subtropical gyres. Since the oligotrophic subtropical gyres have low NPP, the diagnosis

686   would thus suggest that somehow fDOM is higher in the gyres. At depths, both the

687   extrapolated and simulated DOC_t show a gradual decline in concentrations from the North

688   Atlantic to the North Pacific (Figure 9b, d). The highest deep DOC_t in the *LV* run is seen just

689   south of Greenland, where convection occurs in the model.

690

691   Finally, we show that the deep ocean radiocarbon aging is larger in DIC than in DOC_t in the

692   model (Figure 10). The North Pacific-North Atlantic Δ^{14}C gradient is roughly -100‰ for

693   DIC and -70‰ for DOC_t. The oldest DOC_t Δ^{14}C is approximately -430‰ in the North Pacific.

694   If ^{14}C decay were the only mechanism of change along the path of the deepwater

695   circulation, the Δ^{14}C gradient should be quite similar between DIC and DOC_t, which are both

696 dissolved phases and transported passively by the same circulation. The one potentially

697 important difference is that the addition of the relatively young $DI^{14}C$ and $DO^{14}C$ to the

698 deep ocean by the "deep POC split" (see Section 2.3) impacts DOC_t $\Delta^{14}C$ more than DIC $\Delta^{14}C$,

699 because DOC_t is two orders of magnitude lower in concentration than DIC.

700

701 In observations, the aging of DIC and DOC_t is reportedly similar in the Antarctic Bottom

702 Water (below 4000 m) of the deep Pacific (Druffel et al., 2019). This may be explained by

703 the fact that there would not be much deep POC split occurring so deep in the ocean. The

704 North Pacific-North Atlantic $\Delta^{14}C$ gradient, accounting for thermonuclear bomb ^{14}C, may be

705 as large as -100‰ for DOC_t (about -550‰ in the deep Pacific and -456‰ in the deep

706 Atlantic) (Druffel et al., 2019). This gradient is not rigorously determined, because there is

707 not enough data to do an objective analysis. Therefore, the equivalent $\Delta^{14}C$ gradient for DIC

708 cannot be determined. However, the DIC $\Delta^{14}C$ endmember values by inspection (about -

709 250‰ in the deep Pacific and -70‰ in the deep Atlantic) (Matsumoto and Key, 2004)

710 indicate a clearly larger $\Delta^{14}C$ gradient for DIC than DOC_t as simulated by the experiment *LV*.

711

712 One lesson from the data-*LV* run mismatch in the overall DOC_t concentration (Figure 7) and

713 surface DOC_t pattern (Figure 9) is that the parameter values from the literature do not fully

714 capture the DOC cycle and/or MESMO 3 is still lacking some important DOC process. In *LV*,

715 the surface DOC_t is too high because DOC_r is too high, while DOC_{sl} is not unreasonable

716 (Figure 7). DOC_r is too high because there is too much DOC_r production (e.g., $fDOM_r$=1% is

717 too large), there is too little DOC_r degradation (e.g., one of the DOM decay mechanisms is

718 too slow; Equation 28 and Table 2d), or some combination of both. For example, $fDOM_r$ is a

719 key parameter that is not well constrained by observations. Had we used 0.2% instead of

720 1% for $fDOM_r$, the global mean surface DOC_t drops to 76 μmol kg^{-1} (red line, Figure 7),

721 consistent with observations. For achieving a better surface DOC_t pattern, we may need a

722 different formulation of fDOM that is, for example, negatively related to nutrient

723 concentrations so that fDOM increases in the oligotrophic subtropical gyres (Roshan and

724 DeVries, 2017).

725

726 Another lesson from the DOM modeling exercise is that it is important to simulate DOP_r

727 reasonably well in order to preserve the favorable results we achieved in MESMO 3 with

728 respect to biological production and the phytoplankton C:N:P ratio. We find that in the

729 experiment *LV*, the global mean DOP_r concentration becomes steady at 0.45 µmol-P kg^{-1}. In

730 observations, the mean DOC_r is about 40 µmol-C kg^{-1}, and the DOC_r:DOP_r ratio is estimated

731 to be ~1370:1 (Letscher and Moore, 2015), so DOP_r concentration should only be roughly

732 0.03 µmol-P kg^{-1}. Thus, the simulated DOP_r=0.45 µmol-P kg^{-1} is an order of magnitude too

733 high. Because there is more P in the form of DOP_r in *LV*, the oceanic inventory of PO_4

734 declines, causing a nearly 10% drop in export production compared to the standard

735 MESMO 3. In *LV*, the decline in the surface ocean PO_4 that accompanies the change in the

736 PO_4 inventory acts on the phytoplankton physiology (i.e., P effect on C:P in Equation 5),

737 which leads to a large rise in the global mean phytoplankton community C:P export ratio

738 from 113:1 to 127:1. The implementation of preferential remineralization of DOP (and

739 DON) over DOC (Letscher and Moore, 2015) is one way to deal with the problem of too high

740 DOP_r concentrations.

741

742 **3.3 Large-scale patterns of N_2 fixation and denitrification**

743 The modeled habitat of diazotrophs is concentrated in tropical and subtropical waters

744 between 40°S and 40°N and limited by iron (Figure 1e, f). Most noticeably in North Pacific

745 subtropical gyre, diazotrophs constitute ~40% of total NPP. The latitudinal extent of

746 diazotrophs is mainly determined by surface nitrate availability and physical factors such

747 as surface temperature and irradiance. Low nitrate availability in subtropical gyres gives

748 diazotrophs a competitive advantage over small cyanobacteria. Warm temperature and

749 high irradiance also critical physical factors that drive the growth of diazotrophs in the

750 model.

751

752 The modeled global depth-integrated N_2 fixation is 101 Tg N yr^{-1} (Table 3), and this value

753 falls well within the range of observational and geochemical constraints of 80 – 200 Tg N

754 yr^{-1} (Landolfi et al., 2018). In MESMO 3, N_2 fixation occurs in the North Pacific and mid-to-

755 low latitudes of the Atlantic basin (Supplementary Figure S6), where diazotrophs are

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31

762   generally more abundant (Figure 2e). The elevated N_2 fixation rate in the North Pacific,

763   where nitrate limits eukaryotes and cyanobacteria (Figure 2b, d), can be explained by the

764   healthy growth of diazotrophs, which is not limited by N. In the subtropical and tropical

765   Atlantic and the Indian Ocean, high N_2 fixation is driven by elevated C:P and N:P ratio

766   (Figure 4), exemplified by low phosphate availability and warm surface temperature. This

767   spatial pattern agrees with a recent inverse model study (Wang et al., 2019), which showed

768   an elevated N_2 fixation rate in subtropical gyres.

769

770   Global water-column denitrification is 101 Tg N yr^{-1} (Table 3) and is equal to the global N_2

771   fixation because the model has reached steady state. Denitrification is restricted to the

772   subpolar North Pacific (Figure S6), where sub-surface oxygen concentration is significantly

773   depleted (Figure S3d). Enhanced denitrification in this region is in qualitative agreement

774   with a previous modeling study (Bianchi et al., 2018), which showed the anaerobic niche

775   due to particle microenvironments can significantly expand the hypoxic expanses in the

776   North Pacific. However, the extent of denitrification in our model does not include the

777   eastern equatorial Pacific and northern Indian, which are important hotspots for

778   denitrification (Codispoti, 2007; Deutsch et al., 2007). This issue is typical of coarse-

779   resolution global ocean biogeochemistry models that lack spatial resolution in reproducing

780   intense upwelling (Marchal et al., 1998; Najjar et al., 1992; Yamanaka and Tajika, 1997).

781

782   Finally, the ratio of the global inventories of NO_3 and PO_4 in MESMO 3 is just about 16 at

783   steady state, consistent with observations (Gruber and Sarmiento, 1997). One key model

784   parameter in this regard is the nitrate uptake half saturation constant of diazotrophs, $K_{NO3,3}$

785   in Equation 2. A large value of $K_{NO3,3}$ will make it hard for diazotrophs to obtain fixed N

786   from NO_3, which would facilitate N_2 fixation and pushes up the global N/P ratio. With a

787   smaller value of $K_{NO3,3}$, diazotrophs will more easily uptake NO_3, thus depressing N_2

788   fixation, lowering the global N/P ratio.

789

790   **4. Conclusions**

791   MESMO 3, the third and latest version of MESMO, is comprehensively described here. With

792   a fully flexible C:N:P ratio in three PFTs, a prognostic N cycle, and more mechanistic

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804 schemes of organic matter production and remineralization, MESMO 3 reflects the evolving
805 and accumulating knowledge of the ocean biogeochemistry. The model thus remains an
806 effective tool for investigations of the global biogeochemical cycles especially on long time
807 scale given the model's computational efficiency. In particular, MESMO 3 holds promise for
808 studying the marine DOM cycle. The optional features of MESMO 3 include the semi-labile
809 and refractory pools of C, P, N, and Fe. The fact that the literature values regarding the
810 present marine DOM cycle are unable to simulate key observations indicates an
811 opportunity for MESMO 3 to contribute to an improved understanding of the marine DOM
812 cycle.
813
814 **Code availability**
815 The complete code of MESMO version 3.0 and results presented here are available at
816 GitHub https://github.com/gaia3intc/mesmo.git and have a DOI:
817 10.5281/zenodo.4403605.
818
819 **Author contribution**
820 KM, TT, and JZ developed the model code. KM performed the simulations, carried out
821 analyses, and archived the model code and results. KM and TT wrote the paper.
822

827

33

828   **Tables**

829

830   **Table 1.** MESMO Development

831   PFT = phytoplankton functional types. MESMO2 PFTs are LG = large/diatoms and SM = small.
832   MESMO 3 PFTs are Eu = eukaryotes, Cy = cyanobacteria, and Dz = diazotrophs. OM = organic
833   matter. RNPG = residual nitrate potential growth. T = temperature. PAR = photosynthetically
834   active radiation. fDOM = fraction of NPP routed to dissolved organic matter (DOM). The two
835   types of DOM are semi-labile (DOC, DOP, DON, and DOFe) and refractory (DOCr, DOPr, and
836   DONr). Carbon isotopes (^{12}C, ^{13}C, and ^{14}C) are calculated separately for DOC and DOCr. The
837   run ID is 210310m for the MESMO 3 experiment LVR and 210310o for the experiment LVR
838   with fDOM$_r$=0.2%.

839

840

841   **Table 2.** MESMO 3 Biogeochemical Model Parameters Values

842

843   **Table 3.** Key Biogeochemical Model Diagnostics

844   [a]NPP for MESMO 2 was unavailable as a model output and therefore estimated from POC
845   and fDOM=0.67. [b]NPP (in terms of C) is needed in the calculation of the PFT abundance. The
846   root mean square error (RMSE) of the simulated P, N, Si, and O$_2$ distributions from MESMO
847   2 and 3 was calculated relative to the World Ocean Atlas 2018 gridded data (Garcia et al.,
848   2018, 2019). The model-data comparison is made in the top 100 m for nutrients and below
849   100 m for O$_2$. WOA18 was regridded to the MESMO 3 grid to calculate the RMSE.

850

851

852   References for independent constraints: (1) global NPP (Carr et al., 2006); (2) global POC
853   export (DeVries and Weber, 2017); (3) global DOC export assumed to be 20% of total
854   carbon export (Hansell et al., 2009; Roshan and DeVries, 2017); (4) global opal (Dunne et
855   al., 2007); (5) global CaCO$_3$ export (Berelson et al., 2007); (6) global N fixation and
856   denitrification rates (Landolfi et al., 2018); (7) uptake C:N:P ratio is based on POM
857   measurements (Martiny et al., 2013); (8) export C:N:P ratio is assumed to equal the
858   subsurface remineralization ratio (Anderson and Sarmiento, 1994); (9) Deep O$_2$ from
859   WOA18 below 100 m (Garcia et al., 2019).

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Figures

Figure 1. Schematic diagram of DOM cycling in MESMO 2 versus MESMO 3. In the new model, DOM_r can be activated. DOM_r is produced from POM breakdown, which can occur in the production layer or throughout the water column in the "deep POC split." Possible DOM_r remineralization mechanisms are the slow background degradation that occurs everywhere, thermal degradation in hydrothermal vents, and photodegradation in the surface. See text for details.

Figure 2. NPP-based surface phytoplankton functional type (PFT) abundance and nutrient limitation in MESMO 3. Fractional abundance and nutrient limitation for eukaryotes (a, b), cyanobacteria (c, d), and diazotrophs (e, f).

Figure 3. Eukaryote production in MESMO 3 and $CaCO_3$ export in MESMO 2. In MESMO 3, eukaryote NPP (a) is linked to both opal export (b) and $CaCO_3$ export (c) but the two export productions are differentiated by the residual nitrate potential growth (RNPG). Compare $CaCO_3$ export in MESMO 3 (c) to MESMO 2 (d). Unit = mol m^{-2} $year^{-1}$.

Figure 4. Uptake C:P ratio in the top 100 m in MESMO 3: (a) phytoplankton community C:P, (b) zonal mean C:P of all three PFTs and phytoplankton community, (c) eukaryote C:P, and (d) cyanobacteria C:P. The colors in (b) indicate: black = community C:P, red = eukaryote C:P, green = cyanobacteria C:P, and blue = diazotroph C:P. Also, (b) shows the range of observed C:P ratio binned by latitude (Martiny et al., 2013).

Figure 5. Scatter plots of surface ocean eukaryote and cyanobacteria C:P and C:N vs. environmental drivers in MESMO 3. Columns: 1 = eukaryote C:P, 2 = cyanobacteria C:P, 3 = eukaryote C:N, and 4 = cyanobacteria C:N. Rows: 1 = PO_4, 2 = NO_3, 3 = temperature, and 4 = PAR. Red indicates causal relationship according to the power law formulation of flexible C:N:P ratio. PAR = photosynthetically active radiation in W m^{-2}.

Deleted: The effect of the residual nitrate potential growth (RNPG) on the eukaryote production in MESMO 3. Eukaryote NPP (a), opal export (b), and $CaCO_3$ export (c) in MESMO 3. $CaCO_3$ export in MESMO 2 (d). Unit = mol m^{-2} $year^{-1}$.

905   **Figure 6.** Scatter plots of surface ocean community C:P and C:N vs environmental drivers in
906   MESMO 3.

907

908   **Figure 7.** Global mean vertical profiles of DOC from the DOM_R-enabled MESMO 3. DOC_t
909   ($DOC_{sl}+DOC_r$, black line) and DOC_r (black dashed line) from the *LV* run. Red line is DOC_t
910   after reducing $fDOM_r$ from 1% in *LV* to 0.2% (Experiment 210310o). Unit = µmol kg^{-1}.

911

912   **Figure 8.** Global depth-latitude transect of DOC_t from the DOM_R-enabled MESMO 3 *LV* run.
913   Transects are N-S along 25°W in the Atlantic, E-W along 60°S in the Southern Ocean, and N-
914   S along 165°E in the Pacific. Unit = µmol kg^{-1}.

915

916   **Figure 9.** Assessment of surface and deep ocean DOC_t from the DOM_R-enabled MESMO 3 *LV*
917   run. Data-derived DOC_t distributions in the top 100 m (a) and 2500-4000 m (b). Model-
918   simulated DOC_t distributions in the top 100 m (c) and 2500-4000 m (d). Date-derived DOC_t
919   are from Roshan and DeVries (Roshan and DeVries, 2017). Unit = µmol kg^{-1}.

920

921   **Figure 10.** $\Delta^{14}C$ of deep ocean DIC (a) and DOC_t (b) from the DOM_R-enabled MESMO 3 *LV*
922   run. Vertical average over 2500-4000 m water depth. Unit = ‰.

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924 **References:**

925 Anderson, L. A. and Sarmiento, J. L.: Redfield ratios of remineralization determined by
926   nutrient data analysis, Global Biogeochem. Cycles, 8(1), 65–80,
927   doi:10.1029/93GB03318, 1994.

928 Archer, D., Eby, M., Brovkin, V., Ridgwell, A., Cao, L., Mikolajewicz, U., Caldeira, K.,
929   Matsumoto, K., Munhoven, G., Montenegro, A. and Tokos, K.: Atmospheric Lifetime of
930   Fossil Fuel Carbon Dioxide, Annu. Rev. Earth Planet. Sci., 37(1), 117–134,
931   doi:10.1146/annurev.earth.031208.100206, 2009.

932 Arteaga, L., Pahlow, M. and Oschlies, A.: Global patterns of phytoplankton nutrient and light
933   colimitation inferred from an optimality-based model, Global Biogeochem. Cycles,
934   28(7), 648–661, doi:10.1002/2013GB004668, 2014.

935 Balch, W. M., Bates, N. R., Lam, P. J., Twining, B. S., Rosengard, S. Z., Bowler, B. C., Drapeau, D.
936   T., Garley, R., Lubelczyk, L. C., Mitchell, C. and Rauschenberg, S.: Factors regulating the
937   Great Calcite Belt in the Southern Ocean and its biogeochemical significance, Global
938   Biogeochem. Cycles, 30(8), 1124–1144, doi:10.1002/2016GB005414, 2016.

939 Berelson, W. M., Balch, W. M., Najjar, R., Feely, R. A., Sabine, C. and Lee, K.: Relating
940   estimates of $CaCO_3$ production, export, and dissolution in the water column to
941   measurements of $CaCO_3$ rain into sediment traps and dissolution on the sea floor:A
942   revised global carbonate budget, Global Biogeochem. Cycles, 21(1), 1–15,
943   doi:10.1029/2006GB002803, 2007.

944 Bianchi, D., Weber, T. S., Kiko, R. and Deutsch, C.: Global niche of marine anaerobic
945   metabolisms expanded by particle microenvironments, Nat. Geosci., 11(4), 263–268,
946   doi:10.1038/s41561-018-0081-0, 2018.

947 Brzezinski, M. A.: A switch from $Si(OH)_4$ to NO_3^- depletion in the glacial Southern Ocean,
948   Geophys. Res. Lett., 29(12), doi:10.1029/2001gl014349, 2002.

949 Cao, L., Eby, M., Ridgwell, A., Caldeira, K., Archer, D., Ishida, A., Joos, F., Matsumoto, K.,
950   Mikolajewicz, U., Mouchet, A., Orr, J. C., Plattner, G.-K., Schlitzer, R., Tokos, K.,
951   Totterdell, I., Tschumi, T., Yamanaka, Y. and Yool, A.: The role of ocean transport in the
952   uptake of anthropogenic CO_2, Biogeosciences, 6(3), 375–390, doi:10.5194/bg-6-375-
953   2009, 2009.

954 Carr, M. E., Friedrichs, M. A. M., Schmeltz, M., Noguchi Aita, M., Antoine, D., Arrigo, K. R.,

955   Asanuma, I., Aumont, O., Barber, R., Behrenfeld, M., Bidigare, R., Buitenhuis, E. T.,

956   Campbell, J., Ciotti, A., Dierssen, H., Dowell, M., Dunne, J., Esaias, W., Gentili, B., Gregg,

957   W., Groom, S., Hoepffner, N., Ishizaka, J., Kameda, T., Le Quéré, C., Lohrenz, S., Marra, J.,

958   Mélin, F., Moore, K., Morel, A., Reddy, T. E., Ryan, J., Scardi, M., Smyth, T., Turpie, K.,

959   Tilstone, G., Waters, K. and Yamanaka, Y.: A comparison of global estimates of marine

960   primary production from ocean color, Deep. Res. Part II Top. Stud. Oceanogr., 53(5–7),

961   741–770, doi:10.1016/j.dsr2.2006.01.028, 2006.

962 Cheng, H., Edwards, R. L., Southon, J., Matsumoto, K., Feinberg, J. M., Sinha, A., Zhou, W., Li,

963   H., Li, X., Xu, Y., Chen, S., Tan, M., Wang, Q., Wang, Y., Ning, Y., Lawrence Edwards, R.,

964   Southon, J., Matsumoto, K., Feinberg, J. M., Sinha, A., Zhou, W., Li, H., Li, X., Xu, Y., Chen,

965   S., Tan, M., Wang, Q., Wang, Y. and Ning, Y.: Atmospheric 14C/12C changes during the

966   last glacial period from hulu cave, Science (80-.)., 362(6420), 1293–1297,

967   doi:10.1126/science.aau0747, 2018.

968 Chien, C. Te, Pahlow, M., Schartau, M. and Oschlies, A.: Optimality-based non-Redfield

969   plankton-ecosystem model (OPEM v1.1) in UVic-ESCM 2.9 - Part 2: Sensitivity analysis

970   and model calibration, Geosci. Model Dev., 13(10), 4691–4712, doi:10.5194/gmd-13-

971   4691-2020, 2020.

972 Claussen, M., Mysak, L., Weaver, A., Crucifix, M., Fichefet, T., Loutre, M. F., Weber, S., Alcamo,

973   J., Alexeev, V., Berger, A., Calov, R., Ganopolski, A., Goosse, H., Lohmann, G., Lunkeit, F.,

974   Mokhov, I., Petoukhov, V., Stone, P. and Wang, Z.: Earth system models of intermediate

975   complexity: Closing the gap in the spectrum of climate system models, Clim. Dyn.,

976   18(7), 579–586, doi:10.1007/s00382-001-0200-1, 2002.

977 Codispoti, L. A.: An oceanic fixed nitrogen sink exceeding 400 Tg N a-1 vs the concept of

978   homeostasis in the fixed-nitrogen inventory, Biogeosciences, 4(2), 233–253,

979   doi:10.5194/bg-4-233-2007, 2007.

980 Deutsch, C., Sarmiento, J. L., Sigman, D. M., Gruber, N. and Dunne, J. P.: Spatial coupling of

981   nitrogen inputs and losses in the ocean, Nature, 445(7124), 163–167,

982   doi:10.1038/nature05392, 2007.

983 DeVries, T. and Weber, T.: The export and fate of organic matter in the ocean: New

984   constraints from combining satellite and oceanographic tracer observations, Global

985   Biogeochem. Cycles, 31(3), 535–555, doi:10.1002/2016GB005551, 2017.

38

986   Druffel, E. R. M., Williams, P. M., Bauer, J. E. and Ertel, J. R.: Cycling of dissolved and
987   particulate organic matter in the open ocean, J. Geophys. Res., 97(C10), 15639,
988   doi:10.1029/92JC01511, 1992.

989   Druffel, E. R. M., Griffin, S., Wang, N., Garcia, N. G., McNichol, A. P., Key, R. M. and Walker, B.
990   D.: Dissolved Organic Radiocarbon in the Central Pacific Ocean, Geophys. Res. Lett.,
991   46(10), 5396–5403, doi:10.1029/2019GL083149, 2019.

992   Dunne, J. P., Armstrong, R. A., Gnanadesikan, A. and Sarmiento, J. L.: Empirical and
993   mechanistic models for the particle export ratio, Global Biogeochem. Cycles, 19(4),
994   n/a-n/a, doi:10.1029/2004GB002390, 2005.

995   Dunne, J. P., Sarmiento, J. L. and Gnanadesikan, A.: A synthesis of global particle export from
996   the surface ocean and cycling through the ocean interior and on the seafloor, Global
997   Biogeochem. Cycles, 21(4), 1–16, doi:10.1029/2006GB002907, 2007.

998   Eby, M., Weaver, A. J., Alexander, K., Zickfeld, K., Abe-Ouchi, A., Cimatoribus, A. A., Crespin,
999   E., Drijfhout, S. S., Edwards, N. R., Eliseev, A. V., Feulner, G., Fichefet, T., Forest, C. E.,
1000   Goosse, H., Holden, P. B., Joos, F., Kawamiya, M., Kicklighter, D., Kienert, H., Matsumoto,
1001   K., Mokhov, I. I., Monier, E., Olsen, S. M., Pedersen, J. O. P., Perrette, M., Philippon-
1002   Berthier, G., Ridgwell, A., Schlosser, A., Schneider von Deimling, T., Shaffer, G., Smith, R.
1003   S., Spahni, R., Sokolov, A. P., Steinacher, M., Tachiiri, K., Tokos, K., Yoshimori, M., Zeng,
1004   N. and Zhao, F.: Historical and idealized climate model experiments: an
1005   intercomparison of Earth system models of intermediate complexity, Clim. Past, 9(3),
1006   1111–1140, doi:10.5194/cp-9-1111-2013, 2013.

1007   Edwards, N. R. and Marsh, R.: Uncertainties due to transport-parameter sensitivity in an
1008   efficient 3-D ocean-climate model, Clim. Dyn., 24(4), 415–433, doi:10.1007/s00382-
1009   004-0508-8, 2005.

1010   Eppley, R. W.: Temperature and phytoplankton growth in the sea, Fish. Bull., 70(4), 1063–
1011   85, 1972.

1012   Galbraith, E. D. and Martiny, A. C.: A simple nutrient-dependence mechanism for predicting
1013   the stoichiometry of marine ecosystems, Proc. Natl. Acad. Sci., 112(27), 8199–8204,
1014   doi:10.1073/pnas.1423917112, 2015.

1015   Garcia, H., Weathers, K. W., Paver, C. R., Smolyar, I., Boyer, T. P., Locarnini, R. A., Zweng, M.
1016   M., Mishonov, A. V., Baranova, O. K., Seidov, D. and Reagan, J. R.: World Ocean Atlas

1017   2018. Volume 4: Dissolved Inorganic Nutrients (phosphate, nitrate and nitrate+nitrite,

1018   silicate), NOAA Atlas NESDIS 84, 84(July), 35, 2018.

1019   Garcia, H. E., Weathers, K., Paver, C. R., Smolyar, I., Boyer, T. P., Locarnini, R. A., Zweng, M.

1020   M., Mishonov, A. V., Baranova, O. K., Seidov, D. and Reagan, J. R.: World Ocean Atlas

1021   2018, Volume 3: Dissolved Oxygen, Apparent Oxygen Utilization, and Oxygen

1022   Saturation, NOAA Atlas NESDIS, 3(83), 38 pp., 2019.

1023   Gruber, N. and Sarmiento, J. L.: Global patterns of marine nitrogen fixation and

1024   denitrification, Global Biogeochem. Cycles, 11(2), 235–266, doi:10.1029/97GB00077,

1025   1997.

1026   Hansell, D. A.: Recalcitrant Dissolved Organic Carbon Fractions, Ann. Rev. Mar. Sci., 5, 421–

1027   445, doi:10.1146/annurev-marine-120710-100757, 2013.

1028   Hansell, D. A. and Carlson, C. A.: Deep-ocean gradients in the concentration of dissolved

1029   organic carbon, Nature, 395(6699), 263–266, doi:10.1038/26200, 1998.

1030   Hansell, D. A., Carlson, C. A., Repeta, D. J. and Schlitzer, R.: Dissolved Organic Matter in the

1031   Ocean: A Controversy stimulates new insights, Oceanography, 22(4), 202–211,

1032   doi:https://doi.org/10.5670/oceanog.2009.109, 2009.

1033   Henson, S. A., Sanders, R., Madsen, E., Morris, P. J., Le Moigne, F. and Quartly, G. D.: A

1034   reduced estimate of the strength of the ocean's biological carbon pump, Geophys. Res.

1035   Lett., 38(4), n/a-n/a, doi:10.1029/2011GL046735, 2011.

1036   Hutchins, D. A. and Bruland, K. W.: Iron-limited growth and Si:N ratios in a costal upwelling

1037   regime, Nature, 393(June), 561–564, 1998.

1038   Joos, F., Roth, R., Fuglestvedt, J. S., Peters, G. P., Enting, I. G., Von Bloh, W., Brovkin, V., Burke,

1039   E. J., Eby, M., Edwards, N. R., Friedrich, T., Frölicher, T. L., Halloran, P. R., Holden, P. B.,

1040   Jones, C., Kleinen, T., Mackenzie, F. T., Matsumoto, K., Meinshausen, M., Plattner, G. K.,

1041   Reisinger, A., Segschneider, J., Shaffer, G., Steinacher, M., Strassmann, K., Tanaka, K.,

1042   Timmermann, A. and Weaver, A. J.: Carbon dioxide and climate impulse response

1043   functions for the computation of greenhouse gas metrics: A multi-model analysis,

1044   Atmos. Chem. Phys., 13(5), 2793–2825, doi:10.5194/acp-13-2793-2013, 2013.

1045   Kwiatkowski, L., Aumont, O., Bopp, L. and Ciais, P.: The Impact of Variable Phytoplankton

1046   Stoichiometry on Projections of Primary Production, Food Quality, and Carbon Uptake

1047   in the Global Ocean, Global Biogeochem. Cycles, 516–528,

40

1048   doi:10.1002/2017GB005799, 2018.

1049 Kwiatkowski, L., Aumont, O. and Bopp, L.: Consistent trophic amplification of marine

1050   biomass declines under climate change, Glob. Chang. Biol., 25(1), 218–229,

1051   doi:10.1111/gcb.14468, 2019.

1052 Landolfi, A., Kähler, P., Koeve, W. and Oschlies, A.: Global marine N2 fixation estimates:

1053   From observations to models, Front. Microbiol., 9(SEP), 1–8,

1054   doi:10.3389/fmicb.2018.02112, 2018.

1055 Lang, S. Q., Butterfield, D. A., Lilley, M. D., Paul Johnson, H. and Hedges, J. I.: Dissolved

1056   organic carbon in ridge-axis and ridge-flank hydrothermal systems, Geochim.

1057   Cosmochim. Acta, 70(15), 3830–3842, doi:10.1016/j.gca.2006.04.031, 2006.

1058 Laufkötter, C., John, J. G., Stock, C. A. and Dunne, J. P.: Temperature and oxygen dependence

1059   of the remineralization of organic matter, Global Biogeochem. Cycles, 31(7), 1038–

1060   1050, doi:10.1002/2017GB005643, 2017.

1061 Laws, E. A., Falkowski, P. G., Smith, W. O., Ducklow, H. W. and McCarthy, J. J.: Temperature

1062   effects on export production in the open ocean, Global Biogeochem. Cycles, 14(4),

1063   1231–1246, doi:10.1029/1999GB001229, 2000.

1064 Lee, S. Y., Chiang, J. C. H., Matsumoto, K. and Tokos, K. S.: Southern Ocean wind response to

1065   North Atlantic cooling and the rise in atmospheric CO 2 : Modeling perspective and

1066   paleoceanographic implications, Paleoceanography, 26, 1–16,

1067   doi:10.1029/2010PA002004, 2011.

1068 Lenton, T. M., Williamson, M. S., Edwards, N. R., Marsh, R., Price, A. R., Ridgwell, A. J.,

1069   Shepherd, J. G. and Cox, S. J.: Millennial timescale carbon cycle and climate change in an

1070   efficient Earth system model, Clim. Dyn., 26(7–8), 687–711, doi:10.1007/s00382-006-

1071   0109-9, 2006.

1072 Letscher, R. T. and Moore, J. K.: Preferential remineralization of dissolved organic

1073   phosphorus and non-Redfield DOM dynamics in the global ocean: Impacts on marine

1074   productivity, nitrogen fixation, and carbon export, Global Biogeochem. Cycles, 29(3),

1075   325–340, doi:10.1002/2014GB004904, 2015.

1076 Levitus, S.: Climatological atlas of the world ocean, US Department of Commerce, National

1077   Oceanic and Atmospheric Administration., 1982.

1078 Mahowald, N. M., Muhs, D. R., Levis, S., Rasch, P. J., Yoshioka, M., Zender, C. S. and Luo, C.:

41

1079   Change in atmospheric mineral aerosols in response to climate: Last glacial period,

1080   preindustrial, modern, and doubled carbon dioxide climates, J. Geophys. Res. Atmos.,

1081   111(D10), n/a-n/a, doi:10.1029/2005JD006653, 2006.

1082 Marchal, O., Stocker, T. F. and Joos, F.: A latitude-depth, circulation-biogeochemical ocean

1083   model for palaeoclimate studies. Development and sensitivities, Tellus, Ser. B Chem.

1084   Phys. Meteorol., 50B(3), 290–316, doi:10.1034/j.1600-0889.1998.t01-2-00006.x,

1085   1998.

1086 Martiny, A. C., Pham, C. T. A. A., Primeau, F. W., Vrugt, J. A., Moore, J. K., Levin, S. A. and

1087   Lomas, M. W.: Strong latitudinal patterns in the elemental ratios of marine plankton

1088   and organic matter, Nat. Geosci., 6(4), 279–283, doi:10.1038/ngeo1757, 2013.

1089 Mather, R. L., Reynolds, S. E., Wolff, G. A., Williams, R. G., Torres-Valdes, S., Woodward, E. M.

1090   S., Landolfi, A., Pan, X., Sanders, R. and Achterberg, E. P.: Phosphorus cycling in the

1091   North and South Atlantic Ocean subtropical gyres, Nat. Geosci., 1(7), 439–443,

1092   doi:10.1038/ngeo232, 2008.

1093 Matsumoto, K. and Key, R. M.: Natural radiocarbon distribution in the deep ocean, in Global

1094   Environmental Change in the Ocean and on Land, edited by M. Shiyomi, H. Kawahata,

1095   H. Koizumi, A. Tsuda, and Y. Awaya, pp. 45–58, Terrapub, Tokyo. [online] Available

1096   from: http://svr4.terrapub.co.jp/e-library/kawahata/pdf/045.pdf, 2004.

1097 Matsumoto, K. and McNeil, B.: Decoupled response of ocean acidification to variations in

1098   climate sensitivity, J. Clim., 26, 1764–1771, doi:10.1175/JCLI-D-12-00290.1, 2012.

1099 Matsumoto, K. and Tanioka, T.: Shifts in regional production as a driver of future global

1100   ocean production stoichiometry, Environ. Res. Lett., 15(124027), doi:10.1088/1748-

1101   9326/abc4b0, 2020.

1102 Matsumoto, K. and Yokoyama, Y.: Atmospheric Δ 14 C reduction in simulations of Atlantic

1103   overturning circulation shutdown, Global Biogeochem. Cycles, 27(2), 296–304,

1104   doi:10.1002/gbc.20035, 2013.

1105 Matsumoto, K., Tokos, K. S., Price, A. R. and Cox, S. J.: First description of the Minnesota

1106   Earth System Model for Ocean biogeochemistry (MESMO 1.0), Geosci. Model Dev., 1(1),

1107   1–15, doi:10.5194/gmd-1-1-2008, 2008.

1108 Matsumoto, K., Tokos, K., Chikamoto, M. and Ridgwell, A.: Characterizing post-industrial

1109   changes in the ocean carbon cycle in an Earth system model, Tellus B Chem. Phys.

1110   Meteorol., 62(4), 296–313, doi:10.1111/j.1600-0889.2010.00461.x, 2010.

1111   Matsumoto, K., Tokos, K., Huston, A. and Joy-Warren, H.: MESMO 2: a mechanistic marine

1112   silica cycle and coupling to a simple terrestrial scheme, Geosci. Model Dev., 6(2), 477–

1113   494, doi:10.5194/gmd-6-477-2013, 2013.

1114   Matsumoto, K., Rickaby, R. and Tanioka, T.: Carbon Export Buffering and CO2 Drawdown by

1115   Flexible Phytoplankton C:N:P Under Glacial Conditions, Paleoceanogr.

1116   Paleoclimatology, 35(7), 1–22, doi:10.1029/2019PA003823, 2020.

1117   Moore, J. K., Doney, S. C. and Lindsay, K.: Upper ocean ecosystem dynamics and iron cycling

1118   in a global three-dimensional model, Global Biogeochem. Cycles, 18(4), 1–21,

1119   doi:10.1029/2004GB002220, 2004.

1120   Mopper, K., Zhou, X., Kieber, R. J., Kieber, D. J., Sikorski, R. J. and Jones, R. D.: Photochemical

1121   degradation of dissolved organic carbon and its impact on the oceanic carbon cycle,

1122   Nature, 353(6339), 60–62, doi:10.1038/353060a0, 1991.

1123   Najjar, R. G., Sarmiento, J. L. and Toggweiler, J. R.: Downward transport and fate of organic

1124   matter in the ocean: Simulations with a general circulation model, Global Biogeochem.

1125   Cycles, 6(1), 45–76, doi:10.1029/91GB02718, 1992.

1126   Pahlow, M. and Oschlies, A.: Chain model of phytoplankton P, N and light colimitation, Mar.

1127   Ecol. Prog. Ser., 376(2), 69–83, doi:10.3354/meps07748, 2009.

1128   Pahlow, M. and Oschlies, A.: Optimal allocation backs droop's cell-quota model, Mar. Ecol.

1129   Prog. Ser., 473(2010), 1–5, doi:10.3354/meps10181, 2013.

1130   Pahlow, M., Dietze, H. and Oschlies, A.: Optimality-based model of phytoplankton growth

1131   and diazotrophy, Mar. Ecol. Prog. Ser., 489, 1–16, doi:10.3354/meps10449, 2013.

1132   Pahlow, M., Chien, C. Te, Arteaga, L. A. and Oschlies, A.: Optimality-based non-Redfield

1133   plankton-ecosystem model (OPEM v1.1) in UVic-ESCM 2.9 - Part 1: Implementation

1134   and model behaviour, Geosci. Model Dev., 13(10), 4663–4690, doi:10.5194/gmd-13-

1135   4663-2020, 2020.

1136   Paulsen, H., Ilyina, T., Six, K. D. and Stemmler, I.: Incorporating a prognostic representation

1137   of marine nitrogen fixers into the global ocean biogeochemical model HAMOCC, J. Adv.

1138   Model. Earth Syst., 9(1), 438–464, doi:10.1002/2016MS000737, 2017.

1139   Ridgwell, A.: Glacial-interglacial perturbations in the global carbon cycle, Ph, D. thesis, 134

1140   pp., Univ. East Anglia, Norwich, U. K. Ridgwell, A. J., U. Edwards, 2001.

43

1141   Ridgwell, A., Hargreaves, J. C., Edwards, N. R., Annan, J. D., Lenton, T. M., Marsh, R., Yool, A.
1142   and Watson, A.: Marine geochemical data assimilation in an efficient Earth System
1143   Model of global biogeochemical cycling, Biogeosciences, 4(4), 87–104, doi:10.5194/bg-
1144   4-87-2007, 2007.

1145   Ridgwell, A. J., Watson, A. J. and Archer, D. E.: Modeling the response of the oceanic Si
1146   inventory to perturbation, and consequences for atmospheric CO_2, Global
1147   Biogeochem. Cycles, 16(4), 19-1-19–25, doi:10.1029/2002GB001877, 2002.

1148   Roshan, S. and DeVries, T.: Efficient dissolved organic carbon production and export in the
1149   oligotrophic ocean, Nat. Commun., 8(1), 2036, doi:10.1038/s41467-017-02227-3,
1150   2017.

1151   Sarmiento, J. L., Gruber, N., Brzezinski, M. A. and Dunne, J. P.: High-latitude controls of
1152   thermocline nutrients and low latitude biological productivity, Nature, 427(6969), 56–
1153   60, doi:10.1038/nature02127, 2004.

1154   Sun, X. and Matsumoto, K.: Effects of sea ice on atmospheric pCO_2: A revised view and
1155   implications for glacial and future climates, J. Geophys. Res., 115(G02015),
1156   doi:10.1029/2009JG001023, 2010.

1157   Sunda, W. G. and Huntsman, S. A.: Iron uptake and growth limitation in oceanic and coastal
1158   phytoplankton, Mar. Chem., 50(1–4), 189–206, doi:10.1016/0304-4203(95)00035-P,
1159   1995.

1160   Sverdrup, H. U.: On the conditions for the vernal blooming of phytoplankton, J. Cons. Perm.
1161   Int. Pour l'Exploration La Mer, 18, 287–195, 1953.

1162   Takeda, S.: Influence of iron availability on nutrient consumption ratio, Nature, 393(JUNE),
1163   774–777, 1998.

1164   Tanioka, T. and Matsumoto, K.: Buffering of Ocean Export Production by Flexible Elemental
1165   Stoichiometry of Particulate Organic Matter, Global Biogeochem. Cycles, 31(10), 1528–
1166   1542, doi:10.1002/2017GB005670, 2017.

1167   Tanioka, T. and Matsumoto, K.: A meta-analysis on environmental drivers of marine
1168   phytoplankton C : N : P, Biogeosciences, 17(11), 2939–2954, doi:10.5194/bg-17-2939-
1169   2020, 2020a.

1170   Tanioka, T. and Matsumoto, K.: Stability of Marine Organic Matter Respiration
1171   Stoichiometry, Geophys. Res. Lett., 47(1), 1–10, doi:10.1029/2019GL085564, 2020b.

44

1172   Ushie, H. and Matsumoto, K.: The role of shelf nutrients on glacial-interglacial CO 2 : A

1173   negative feedback, Global Biogeochem. Cycles, 26, 1–10, doi:10.1029/2011GB004147,

1174   2012.

1175   Wang, W.-L., Moore, J. K., Martiny, A. C. and Primeau, F. W.: Convergent estimates of marine

1176   nitrogen fixation, Nature, 566(7743), 205–211, doi:10.1038/s41586-019-0911-2,

1177   2019.

1178   Wanninkhof, R.: Relationship between wind speed and gas exchange over the ocean, J.

1179   Geophys. Res., 97(C5), 7373–7382, doi:10.1029/92JC00188, 1992.

1180   Weaver, A. J., Sedlá, J., Eby, M., Alexander, K., Crespin, E., Fichefet, T., Philippon-berthier, G.,

1181   Joos, F., Kawamiya, M., Matsumoto, K., Steinacher, M., Tachiiri, K., Tokos, K., Yoshimori,

1182   M. and Zickfeld, K.: Stability of the Atlantic meridional overturning circulation : A

1183   model intercomparison, Geophys. Res. Lett., 39, 1–7, doi:10.1029/2012GL053763,

1184   2012.

1185   Yamanaka, Y. and Tajika, E.: Role of dissolved organic matter in the marine biogeochemical

1186   cycle: Studies using an ocean biogeochemical general circulation model, Global

1187   Biogeochem. Cycles, 11(4), 599–612, doi:10.1029/97GB02301, 1997.

1188   Yamanaka, Y., Yoshie, N., Fujii, M., Aita, M. N. and Kishi, M. J.: An Ecosystem Model Coupled

1189   with Nitrogen-Silicon-Carbon Cycles Applied to Station A7 in the Northwestern Pacific,

1190   J. Oceanogr., 60(2), 227–241, doi:10.1023/B:JOCE.0000038329.91976.7d, 2004.

1191   Zickfeld, K., Eby, M., Weaver, A. J., Alexander, K., Crespin, E., Edwards, N. R., Eliseev, A. V.,

1192   Feulner, G., Fichefet, T., Forest, C. E., Friedlingstein, P., Goosse, H., Holden, P. B., Joos, F.,

1193   Kawamiya, M., Kicklighter, D., Kienert, H., Matsumoto, K., Mokhov, I. I., Monier, E.,

1194   Olsen, S. M., Pedersen, J. O. P., Perrette, M., Philippon-Berthier, G., Ridgwell, A.,

1195   Schlosser, A., Von Deimling, T. S., Shaffer, G., Sokolov, A., Spahni, R., Steinacher, M.,

1196   Tachiiri, K., Tokos, K. S., Yoshimori, M., Zeng, N. and Zhao, F.: Long-Term climate

1197   change commitment and reversibility: An EMIC intercomparison, J. Clim., 26(16),

1198   5782–5809, doi:10.1175/JCLI-D-12-00584.1, 2013.

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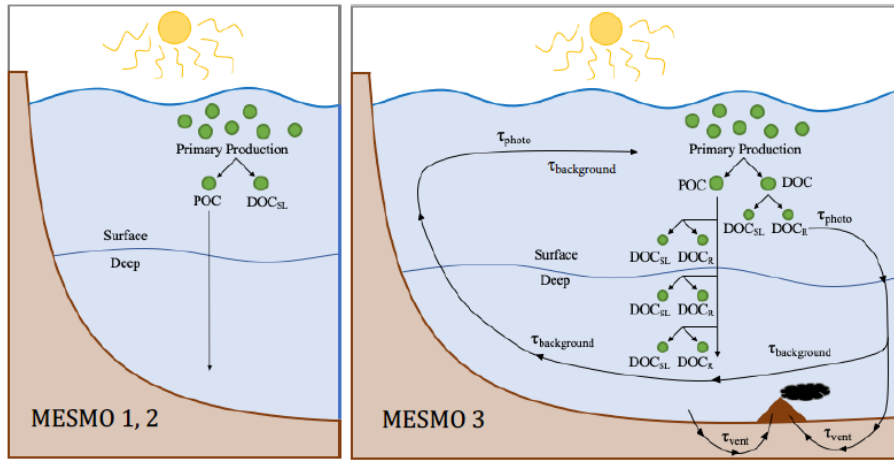
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1203 **Figure 1.** Schematic diagram of DOM cycling in MESMO 2 versus MESMO 3. In the new

1204 model, DOM_r can be activated. DOM_r is produced from POM breakdown, which can occur in

1205 the production layer or throughout the water column in the "deep POC split." Possible

1206 DOM_r remineralization mechanisms are the slow background degradation that occurs

1207 everywhere, thermal degradation in hydrothermal vents, and photodegradation in the

1208 surface. See text for details.

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Figure 2. NPP-based surface phytoplankton functional type (PFT) abundance and nutrient limitation in MESMO 3. Fractional abundance and nutrient limitation for eukaryotes (a, b), cyanobacteria (c, d), and diazotrophs (e, f).

(a) MESMO3 Eukaryotes NPP (mol-C m^{-2} yr^{-1})

(b) MESMO 3 Opal Export (mol-Si m^{-2} yr^{-1})

(c) MESMO3 CaCO$_3$ Export (mol-C m^{-2} yr^{-1})

(d) MESMO2 CaCO$_3$ Export (mol-C m^{-2} yr^{-1})

Figure 3. Eukaryote production in MESMO 3 and CaCO$_3$ export in MESMO 2. In MESMO 3, eukaryote NPP (a) is linked to both opal export (b) and CaCO$_3$ export (c) but the two export productions are differentiated by the residual nitrate potential growth (RNPG). Compare CaCO$_3$ export in MESMO 3 (c) to MESMO 2 (d). Unit = mol m^{-2} year^{-1}.

Figure 4. Uptake C:P ratio in the top 100 m in MESMO 3: (a) phytoplankton community C:P, (b) zonal mean C:P of all three PFTs and phytoplankton community, (c) eukaryote C:P, and (d) cyanobacteria C:P. The colors in (b) indicate: black = community C:P, red = eukaryote C:P, green = cyanobacteria C:P, and blue = diazotroph C:P. Also, (b) shows the range of observed C:P ratio binned by latitude (Martiny et al., 2013).

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Figure 5. Scatter plots of surface ocean eukaryote and cyanobacteria C:P and C:N vs.

environmental drivers in MESMO 3. Columns: 1 = eukaryote C:P, 2 = cyanobacteria C:P, 3 =

eukaryote C:N, and 4 = cyanobacteria C:N. Rows: 1 = PO$_4$, 2 = NO$_3$, 3 = temperature, and 4 =

PAR. Red indicates causal relationship according to the power law formulation of flexible

C:N:P ratio. PAR = photosynthetically active radiation in W m^{-2}.

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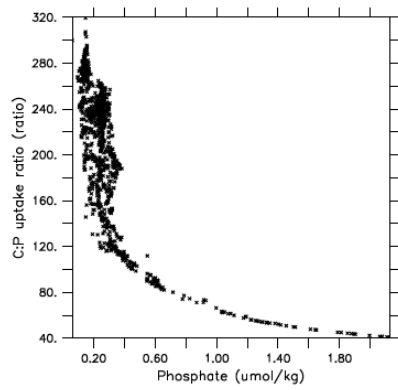
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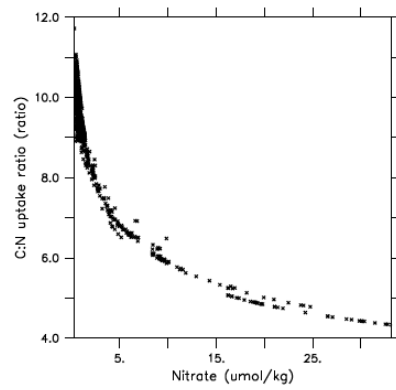
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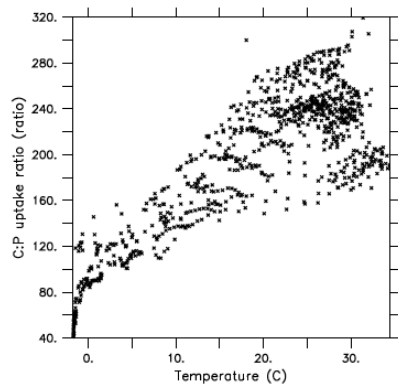
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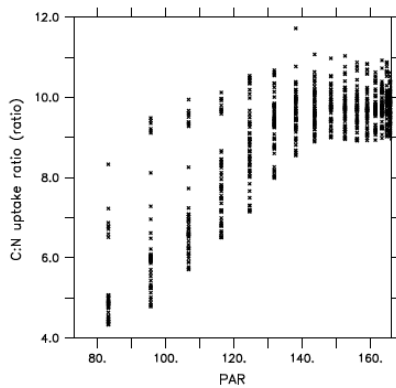
com C:N vs PAR

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1249 **Figure 6.** Scatter plots of surface ocean community C:P and C:N vs environmental drivers in

1250 MESMO 3.

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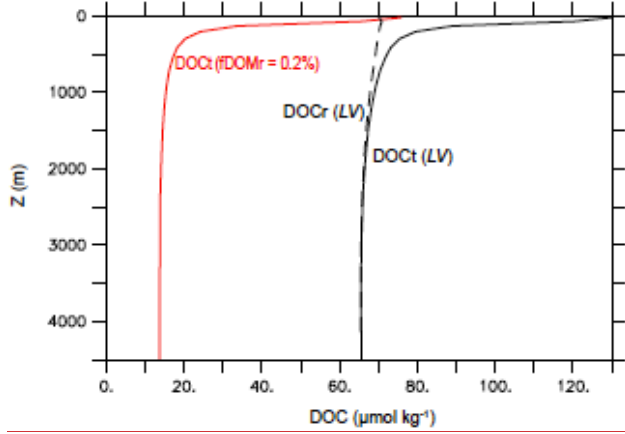
1254 **Figure 7.** Global mean
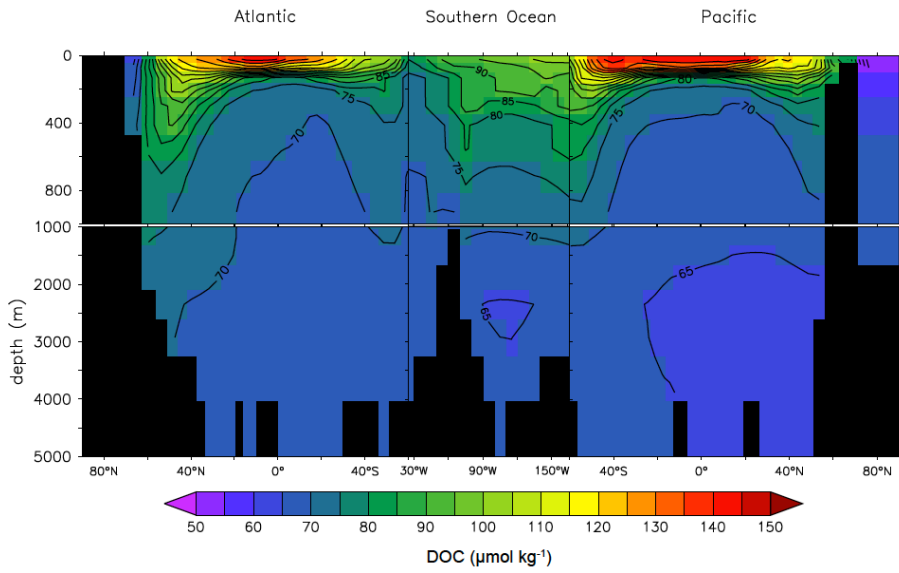
1255 vertical profiles of DOC from the DOM_R-enabled MESMO 3. DOC_t ($DOC_{sl}+DOC_r$, black line)

1256 and DOC_r (black dashed line) from the *LV* run (Experiment 210310m). Red line is DOC_t

1257 after reducing $fDOM_r$ from 1% in *LV* to 0.2% (Experiment 210310o). Unit = μmol kg^{-1}.
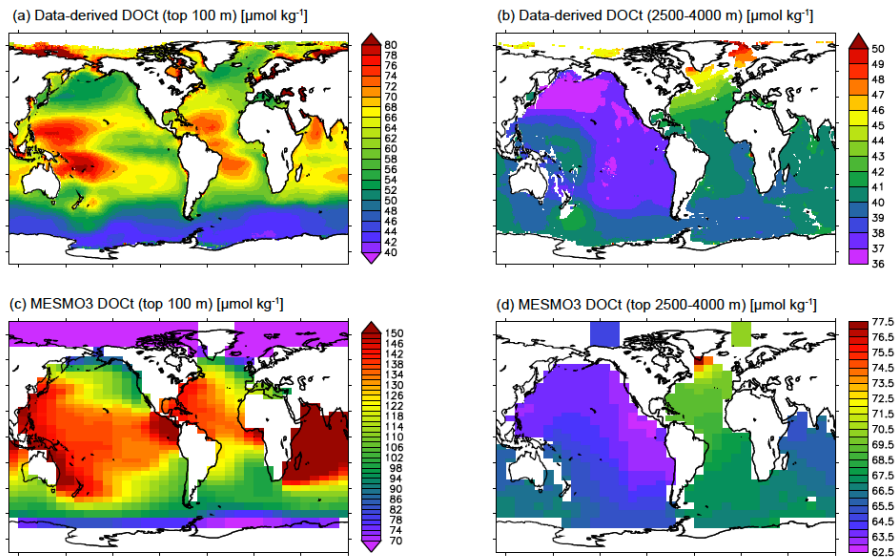
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Figure 8. Global depth-latitude transect of DOC$_t$ from the DOM$_R$-enabled MESMO 3 *LV* run. Transects are N-S along 25°W in the Atlantic, E-W along 60°S in the Southern Ocean, and N-S along 165°E in the Pacific. Unit = μmol kg^{-1}.

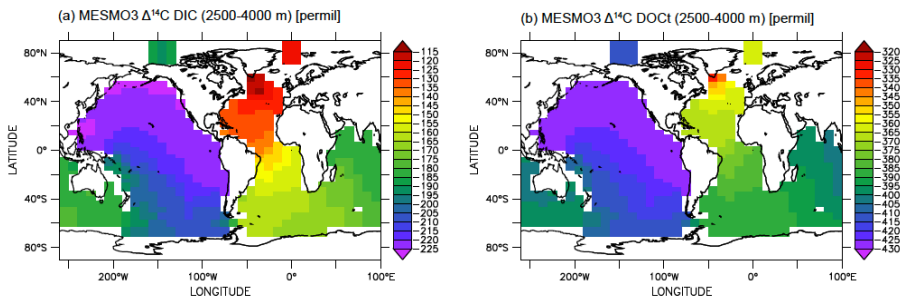**Figure 9.** Assessment of surface and deep ocean DOC_t from the DOM_R-enabled MESMO 3 *LV* run. Data-derived DOC_t distributions in the top 100 m (a) and 2500-4000 m (b). Model-simulated DOC_t distributions in the top 100 m (c) and 2500-4000 m (d). Date-derived DOC_t are from Roshan and DeVries (Roshan and DeVries, 2017). Unit = μmol kg^{-1}.

Figure 10. $\Delta^{14}C$ of deep ocean DIC (a) and DOC_t (b) from the DOM_R-enabled MESMO 3 *LV* run. Vertical average over 2500-4000 m water depth. Unit = ‰.