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

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

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

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

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:

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

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

2.2.3 Optimality-based model of stoichiometry

182

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

2.2.4 Stoichiometry of iron and silica

204

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.

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

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

226

227   **2.3 Production of POM and DOM**

228 In the top 100 m of the model domain, where phytoplankton P uptake occurs (i.e., $\Gamma_i > 0$,

229 see section 2.1), NPP is immediately routed to POM and DOM pools (Figure 1). The

230 production fluxes of POM, DOM$_{sl}$, and DOM$_r$ from NPP are given as Jprod. Here we write the

231 equations in terms of P, which is the master nutrient variable:

232

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

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

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

233

234 The term fDOM denotes the fraction of NPP that is routed to DOM as opposed to POM.

235 Likewise, fDOM$_r$ is the fraction of DOM that is routed to DOM$_r$ as opposed to DOM$_{sl}$. The

236 value of fDOM$_r$ is not well known but estimated to be ~1% (Hansell, 2013), which we

237 tentatively adopt in MESMO 3. If DOM$_r$ is not selected in the model run, fDOM$_r$ = 0. In

238 previous versions of MESMO, fDOM was assigned a constant value of 0.67. In reality, a large

239 variability is observed locally for this ratio, ranging from 0.01-0.2 in temperate waters to

240 0.1-0.7 in the Southern Ocean (Dunne et al., 2005; Henson et al., 2011; Laws et al., 2000). In

241 MESMO 3, fDOM is calculated as a function of the ambient temperature following Laws et

242 al. (Laws et al., 2000):

243

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

244

245 This formulation gives low export efficiency (i.e., high fDOM) in the warmer regions

246 compared to the colder high latitude regions. Locally, we impose fixed fDOM upper and

247 lower bounds of 0.96 and 0.28, respectively, as estimated from a previous study (Dunne et

248 al., 2005).

249

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

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

252 the loss of O_2 directly to the dissolved inorganic forms (i.e., POC-->DIC, POP-->PO_4, and

253 POP-->NO_3). In the new "deep POC split" pathway, sinking POM is simply broken down into

254 DOM without the loss of O_2 as in the production layer (Figure 1). If DOM_r is selected in the

255 model, the broken-down POM is further routed to both DOM_{sl} and DOM_r according to

256 $fDOM_r$. If not, all of the broken down POM is converted to DOM_{sl}. Thus, when the deep POC

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

258 production of DOM and DOM_r in addition to DOM transport from the surface. Thus, the

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

260 distribution.

261

262 **2.4 Production of $CaCO_3$ and opal by eukaryotes**

263 In MESMO 2, opal production was associated with the large PFT and $CaCO_3$ production was

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

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

266 associate both $CaCO_3$ and opal production with the POM production by the same eukaryote

267 PFT ($Jprod_{POM1}$) :

268

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

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

269

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

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

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

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

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

275 rates. The RNPG index recasts the ambient concentrations of NO_3 and $Si(OH)_4$ into potential

276 algal growth rates:

277

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

278

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

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

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

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

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

284   presented in Equation 20 as:

285

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

286

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

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

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

290

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

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

291

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

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

294   growth of carbonate-bearing phytoplankton.

295

296   **2.5 Remineralization of POM and DOM**

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

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

13

299   remineralization had a temperature dependence and decayed exponentially with depth

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

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

302

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

303

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

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

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

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

308

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

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

311   be ~1.5 years (Hansell, 2013):

312

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

313

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

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

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

317   [14]C age of deep ocean DOC_r is 10^3 years (e.g., Druffel et al., 1992), much older than $DI^{14}C$.

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

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

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

321   represent this ubiquitous process with τ_{bg}, which is the inverse of the background decay

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

323

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

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

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

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

328   (Yamanaka and Tajika, 1997). This occurs only in the surface.

329

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

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

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

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

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

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

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

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

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

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

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

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

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

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

344   subject only to background degradation in MESMO 3.

345

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

347   total DOM_r remineralization rate:

348

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

349

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

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

352

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

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

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

359

2.6 Remineralization of CaCO$_3$ and opal

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

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

367

368   Opal remineralization in the water column follows Ridgwell et al. (2002). The rate of opal
369   remineralization R_{opal} is given by the product of normalized dissolution rate (r_{opal}), base
370   opal dissolution rate (k_{opal}), and opal concentration [opal]:
371

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

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

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

372

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

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

375   equilibrium concentration is a function of ambient temperature:

376

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

377

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

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

380

381   **2.7 Conservation of organic matter and biogenic tests**

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

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

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

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

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

387

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

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

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

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

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

388

389   The time rate of change of CaCO$_3$ and opal is expressed in much the same way as POM:

390

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

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

391

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

393   sinking term:

394

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

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

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

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

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

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

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

395

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.

400

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

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

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

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

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

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

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

401

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

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

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

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

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

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

408

409   **2.9 Prognostic nitrogen cycle**

410   Biological production by diazotrophs is stimulated when the ambient NO_3 is low. Nitrogen

411   fixed by diazotrophs during their growth is added to the marine NO_3 pool. The prognostic

412   nitrogen fixation model employed here is similar to that used in the HAMOCC

413   biogeochemical module (Paulsen et al., 2017):

414

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

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

415

416   where Fix_N is the nitrogen fixation rate and I_{NO_3} is the nitrate dependency term in

417   quadratic Michaelis-Menten kinetics form with the half-saturation constant K_{N_2}. See Table

418   2e for the values related to the N cycle.

419

420   Water-column denitrification is formulated in an approach similar to that of the original

421   GENIE model (Ridgwell et al., 2007), in which 2 moles of NO_3 are converted to 1 mole of N_2

422   and liberating 2.5 moles of O_2 as a byproduct:

423

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

424

425   Denitrification takes place in grid boxes, in which O_2 concentration is below a threshold

426   concentration ($O_{2,def}$) and is stimulated if the total global inventory of NO_3 relative to PO_4 is

427   high. In other words, denitrification can effectively act as negative feedback to nitrogen

428   fixation. The threshold O_2 concentration ($O_{2,def}$) takes the minimum of the hard-bound O_2

429   threshold concentration ($O_{2,crit}$) and the NO_3 to PO_4 ratio, scaled by a parameter k_D. The

430   parameters $O_{2,crit}$ and k_D are calibrated to give the global denitrification rate of roughly

431   100 Tg N yr^{-1}, which balances the total nitrogen fixation rate in the model.

432

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

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

433

434 **2.10 Prognostic iron cycle**

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

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

437 to the following equilibrium relationship:

438

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

439

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

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

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

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

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

445

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

446

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

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

449

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

450

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

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

453 remineralization:

454

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

455

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

457

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

458

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

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

461   steady state.

462

463   **2.11 Air-sea gas Exchange**

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

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

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

467   air-sea interface is given by:

468

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

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

469

470   where k is the gas transfer velocity, $[CO_2]_{sat}$ and $[O_2]_{sat}$ are saturation concentrations, and

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

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

473   Schmidt Number for a specific gas:

474

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

475

3 Results and Discussion

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

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

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

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

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

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

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

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

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

486

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

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

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

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

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

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

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

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

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

496 simulate narrow dynamical features such as equatorial upwelling and reproduce

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

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

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

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

501 ~1000 m in both MESMO 3 and the World Ocean Atlas (Figure S3). At 1000 m, the O_2

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

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

504   1000 m is too well-oxygenated in MESMO 2. We believe that the improved match in the O_2

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

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

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

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

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

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

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

512   2004) and simulated previously by MESMO 2 and now MESMO 3.Finally, surface FeT is also

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

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

515

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

517   DON_{sl}, and $DOFe_{sl}$. We note only that the surface DOC_{sl} concentration of 58 $\mu mol\ kg^{-1}$ and

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

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

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

521   3 is 0.69, which is slightly higher than the spatially uniform value of 0.67 in MESMO 2.

522

523   **3.1 Novel features of MESMO 3**

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

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

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

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

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

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

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

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

532   higher latitudes (Figure 2e).

533

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

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

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

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

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

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

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

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

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

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

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

545

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

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

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

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

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

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

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

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

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

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

556 which corresponds to the cyanobacteria PFT in MESMO 3.

557

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

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

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

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

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

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

564   taxonomic effects (i.e., the shift in the community composition changes the weighting of

565   each PFT's C:P ratio). Figure 4b shows that the community C:P is high in oligotrophic gyres

566   because cyanobacteria and to a lesser extent diazotrophs dominate the community, and

567   their C:P ratio is high. Conversely, the community C:P is low in the polar waters because the

568   eukaryotes dominate and their C:P ratio is low. For both eukaryotes and cyanobacteria,

569   their C:P is high in oligotrophic subtropical gyres because PO_4 is low (Figure 4c and d). This

570   physiological effect is larger in eukaryotes than cyanobacteria because the former has

571   greater sensitivity (i.e., larger sensitivity factor $s_{PO_4}^{P:C}$, Equation 5 , Table 2b). However, the

572   cyanobacteria PFT's C:P ratio has an additional sensitivity to temperature (i.e., $s_T^{P:C} \neq 0$)

573   that elevates their C:P in the lower latitudes. We do not show the C:P ratio for diazotrophs

574   because it is very similar to that of cyanobacteria (Figure 4b, d).

575

576   In order to gain more insights into the spatial patterns of the C:P ratio (Figure 4), we

577   examined the relationships between the C:P and C:N ratios and the four possible

578   environmental drivers for eukaryotes and cyanobacteria (Figure 5; again, diazotrophs are

579   not shown). The red plots show that there is a causal relationship between the ratios and

580   the drivers as formulated by the power law model (Equations 5 and 6). The black plots

581   show the absence of a causal relationship. For example, the C:P ratio of both eukaryotes

582   and cyanobacteria are strongly correlated with PO_4 because there is a causal relationship

583   (Figure 5a, b shown in red). Similarly, the C:N ratio of the same two PFTs have a strong

584   correlation with PO_4 (Figure 5c, d in black), but there is actually not a causal relationship

585   (i.e., $s_{PO_4}^{N:C} = 0$, Table 2b). The C:N-PO_4 correlation exists, simply because the nutrients are

586   well correlated. Similarly, because temperature and photosynthetically active radiation

587   (PAR) tend to be correlated via latitude, the stoichiometry has a similar correlation to the

588   two drivers. For example, cyanobacteria C:P has a strong correlation with both

589   temperature and PAR (Figure 5j, 4n), but only the temperature is a real driver. Figure 5

590   indicates which are the dominant drivers of the C:N:P ratio in MESMO 3. For the eukaryote

591   C:P ratio, it is PO_4. For the cyanobacteria C:P ratio, the important drivers are temperature

592   and PO_4. For the C:N ratio for both eukaryotes and cyanobacteria, NO_3 is more important

593   than PAR. Figure 5 also serves to remind us that correlation does not indicate causation.

594

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

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

597 6 includes the effect of taxonomy as well. Still, the effects of PO_4 and temperature are

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

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

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

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

602 community C:N:P ratio.

603

604 **3.2 DOM_r-enabled MESMO 3**

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

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

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

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

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

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

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

612

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

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

615 2013). All other model parameter values in the *LV* run are identical to the standard MESMO

616 3 model (Table 2). The black lines in Figure 7 show the global mean vertical profiles of the

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

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

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

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

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

622 ^{1}, and the observed DOC_{sl} attenuates with depth from 30~40 μmol kg^{-1} near the surface. So

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

624   observations.

625

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

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

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

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

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

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

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

633   ^{1} in the North Atlantic to <65 µmol kg^{-1} in the North Pacific.

634

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

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

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

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

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

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

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

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

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

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

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

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

647

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

649   model (Figure 10). The North Pacific-North Atlantic $\Delta^{14}C$ gradient is roughly -100‰ for

650   DIC and -70‰ for DOC_t. The oldest DOC_t $\Delta^{14}C$ is approximately -430‰ in the North Pacific.

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

652   circulation, the $\Delta^{14}C$ gradient should be quite similar between DIC and DOC_t, which are both

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

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

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

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

657

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

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

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

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

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

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

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

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

666   250‰ in the deep Pacific and -70‰ in the deep Atlantic) (Matsumoto and Key, 2004)

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

668

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

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

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

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

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

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

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

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

677   1% for $fDOM_r$, the global mean surface DOC_t drops to 76 µmol kg^{-1} (red line, Figure 7),

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

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

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

681   DeVries, 2017).

682

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

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

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

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

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

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

689　　0.03 μmol-P kg^{-1}. Thus, the simulated DOP_r=0.45 μmol-P kg^{-1} is an order of magnitude too

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

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

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

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

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

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

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

697　　DOP_r concentrations.

698

699　　**3.3 Large-scale patterns of N_2 fixation and denitrification**

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

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

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

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

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

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

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

707　　model.

708

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

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

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

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

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

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

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

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

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

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

719   an elevated N_2 fixation rate in subtropical gyres.

720

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

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

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

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

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

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

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

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

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

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

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

732

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

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

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

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

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

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

739   fixation, lowering the global N/P ratio.

740

741   **4. Conclusions**

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

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

744   schemes of organic matter production and remineralization, MESMO 3 reflects the evolving

745   and accumulating knowledge of the ocean biogeochemistry. The model thus remains an

746   effective tool for investigations of the global biogeochemical cycles especially on long time

747   scale given the model's computational efficiency. In particular, MESMO 3 holds promise for

748   studying the marine DOM cycle. The optional features of MESMO 3 include the semi-labile

749   and refractory pools of C, P, N, and Fe. The fact that the literature values regarding the

750   present marine DOM cycle are unable to simulate key observations indicates an

751   opportunity for MESMO 3 to contribute to an improved understanding of the marine DOM

752   cycle.

753

754   **Code availability**

755   The complete code of MESMO version 3.0 and results presented here are available at

756   GitHub https://github.com/gaia3intc/mesmo.git and have a DOI:

757   10.5281/zenodo.4403605.

758

759   **Author contribution**

760   KM, TT, and JZ developed the model code. KM performed the simulations, carried out

761   analyses, and archived the model code and results. KM and TT wrote the paper.

762

763   **Acknowledgements**

767

Tables

Table 1. MESMO Development

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

Table 2. MESMO 3 Biogeochemical Model Parameters Values

Table 3. Key Biogeochemical Model Diagnostics

[a]NPP for MESMO 2 was unavailable as a model output and therefore estimated from POC and fDOM=0.67. [b]NPP (in terms of C) is needed in the calculation of the PFT abundance. The root mean square error (RMSE) of the simulated P, N, Si, and O_2 distributions from MESMO 2 and 3 was calculated relative to the World Ocean Atlas 2018 gridded data (Garcia et al., 2018, 2019). The model-data comparison is made in the top 100 m for nutrients and below 100 m for O_2. WOA18 was regridded to the MESMO 3 grid to calculate the RMSE.

References for independent constraints: (1) global NPP (Carr et al., 2006); (2) global POC export (DeVries and Weber, 2017); (3) global DOC export assumed to be 20% of total carbon export (Hansell et al., 2009; Roshan and DeVries, 2017); (4) global opal (Dunne et al., 2007); (5) global $CaCO_3$ export (Berelson et al., 2007); (6) global N fixation and denitrification rates (Landolfi et al., 2018); (7) uptake C:N:P ratio is based on POM measurements (Martiny et al., 2013); (8) export C:N:P ratio is assumed to equal the subsurface remineralization ratio (Anderson and Sarmiento, 1994); (9) Deep O_2 from WOA18 below 100 m (Garcia et al., 2019).

804   **Figures**

805

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

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

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

809   DOM_r remineralization mechanisms are the slow background degradation that occurs

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

811   surface. See text for details.

812

813   **Figure 2.** NPP-based surface phytoplankton functional type (PFT) abundance and nutrient

814   limitation in MESMO 3. Fractional abundance and nutrient limitation for eukaryotes (a, b),

815   cyanobacteria (c, d), and diazotrophs (e, f).

816

817   **Figure 3.** Eukaryote production in MESMO 3 and $CaCO_3$ export in MESMO 2. In MESMO 3,

818   eukaryote NPP (a) is linked to both opal export (b) and $CaCO_3$ export (c) but the two export

819   productions are differentiated by the residual nitrate potential growth (RNPG). Compare

820    $CaCO_3$ export in MESMO 3 (c) to MESMO 2 (d). Unit = mol m^{-2} $year^{-1}$.

821

822   **Figure 4.** Uptake C:P ratio in the top 100 m in MESMO 3: (a) phytoplankton community C:P,

823   (b) zonal mean C:P of all three PFTs and phytoplankton community, (c) eukaryote C:P, and

824   (d) cyanobacteria C:P. The colors in (b) indicate: black = community C:P, red = eukaryote

825   C:P, green = cyanobacteria C:P, and blue = diazotroph C:P. Also, (b) shows the range of

826   observed C:P ratio binned by latitude (Martiny et al., 2013).

827

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

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

830   eukaryote C:N, and 4 = cyanobacteria C:N. Rows: 1 = PO_4, 2 = NO_3, 3 = temperature, and 4 =

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

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

833

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

835   MESMO 3.

836

837   **Figure 7.** Global mean vertical profiles of DOC from the DOM_R-enabled MESMO 3. DOC_t

838   (DOC_{sl}+DOC_r, black line) and DOC_r (black dashed line) from the *LV* run. Red line is DOC_t

839   after reducing $fDOM_r$ from 1% in *LV* to 0.2% (Experiment 210310o). Unit = μmol kg^{-1}.

840

841   **Figure 8.** Global depth-latitude transect of DOC_t from the DOM_R-enabled MESMO 3 *LV* run.

842   Transects are N-S along 25°W in the Atlantic, E-W along 60°S in the Southern Ocean, and N-

843   S along 165°E in the Pacific. Unit = μmol kg^{-1}.

844

845   **Figure 9.** Assessment of surface and deep ocean DOC_t from the DOM_R-enabled MESMO 3 *LV*

846   run. Data-derived DOC_t distributions in the top 100 m (a) and 2500-4000 m (b). Model-

847   simulated DOC_t distributions in the top 100 m (c) and 2500-4000 m (d). Date-derived DOC_t

848   are from Roshan and DeVries (Roshan and DeVries, 2017). Unit = μmol kg^{-1}.

849

850   **Figure 10.** Δ^{14}C of deep ocean DIC (a) and DOC_t (b) from the DOM_R-enabled MESMO 3 *LV*

851   run. Vertical average over 2500-4000 m water depth. Unit = ‰.

852

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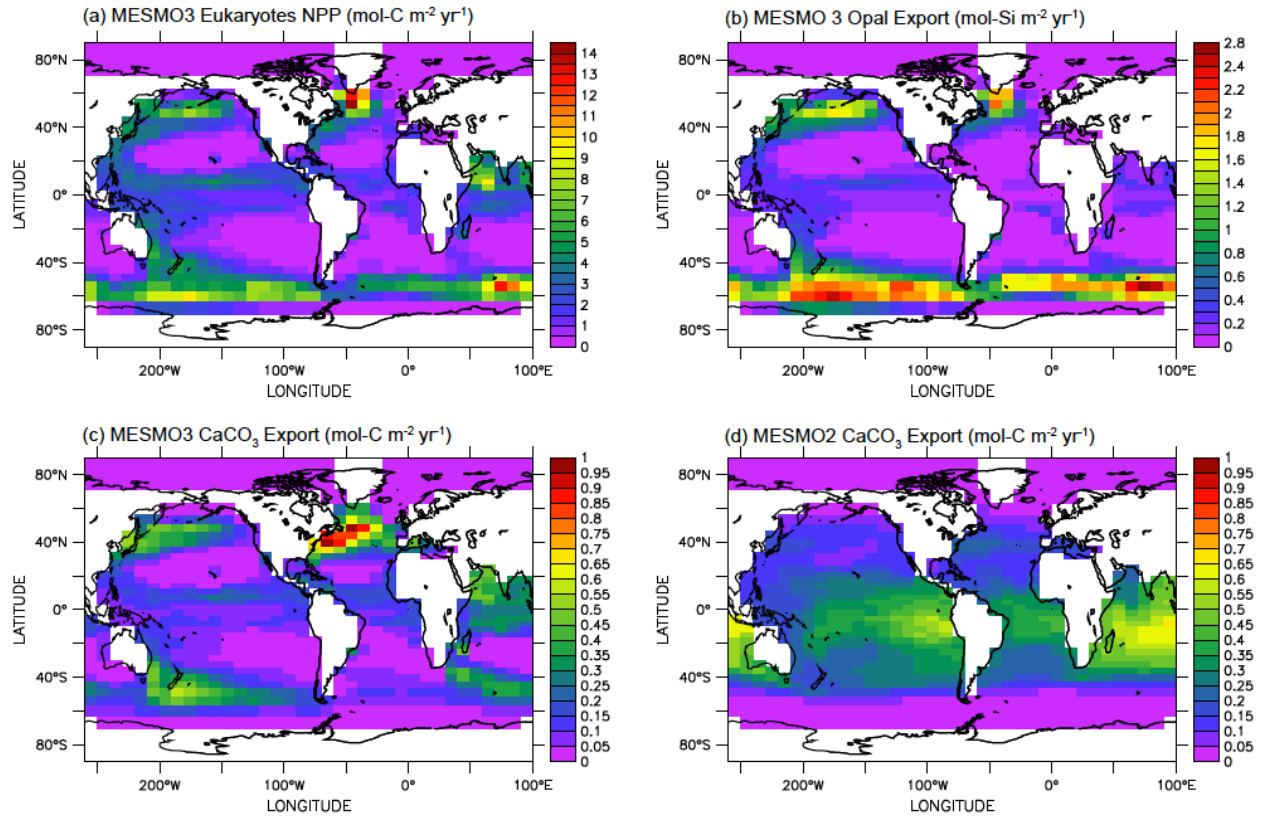1127   5782–5809, doi:10.1175/JCLI-D-12-00584.1, 2013.

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1132 **Figure 1.** Schematic diagram of DOM cycling in MESMO 2 versus MESMO 3. In the new

1133 model, DOM_r can be activated. DOM_r is produced from POM breakdown, which can occur in

1134 the production layer or throughout the water column in the "deep POC split." Possible

1135 DOM_r remineralization mechanisms are the slow background degradation that occurs

1136 everywhere, thermal degradation in hydrothermal vents, and photodegradation in the

1137 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).

Figure 3. Eukaryote production in MESMO 3 and CaCO₃ export in MESMO 2. In MESMO 3, eukaryote NPP (a) is linked to both opal export (b) and CaCO₃ export (c) but the two export productions are differentiated by the residual nitrate potential growth (RNPG). Compare CaCO₃ export in MESMO 3 (c) to MESMO 2 (d). Unit = mol m⁻² year⁻¹.

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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}.

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

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1176   **Figure 7.** Global mean

1177   vertical profiles of DOC from the DOM_R-enabled MESMO 3. DOC_t (DOC_{sl}+DOC_r, black line)

1178   and DOC_r (black dashed line) from the *LV* run (Experiment 210310m). Red line is DOC_t

1179   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}.
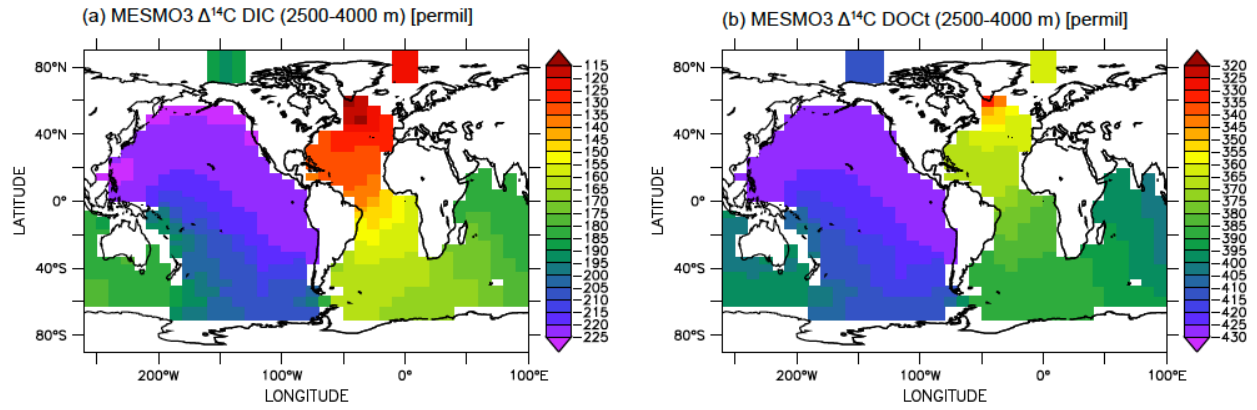
Figure 9. Assessment of surface and deep ocean DOC$_t$ from the DOM$_R$-enabled MESMO 3 *LV* run. Data-derived DOC$_t$ distributions in the top 100 m (a) and 2500-4000 m (b). Model-simulated DOC$_t$ distributions in the top 100 m (c) and 2500-4000 m (d). Date-derived DOC$_t$ are from Roshan and DeVries (Roshan and DeVries, 2017). Unit = µmol kg^{-1}.

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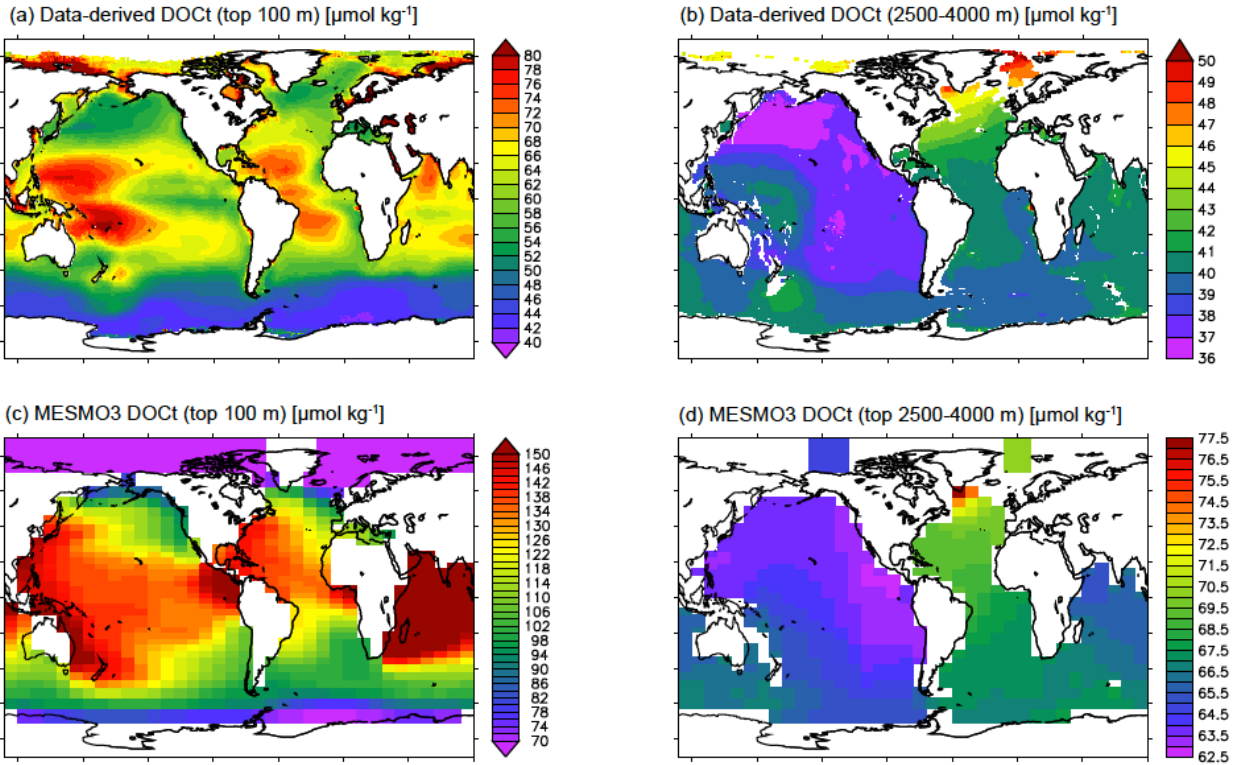
Figure 10. $\Delta^{14}C$ of deep ocean DIC (a) and DOC$_t$ (b) from the DOM$_R$-enabled MESMO 3 *LV*

1197   run. Vertical average over 2500-4000 m water depth. Unit = ‰.

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