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4           **Climate Change Impacts on Net Primary Production (NPP) and Export**  
5           **Production (EP) Regulated by Increasing Stratification and Phytoplankton**  
6           **Community Structure in the CMIP5 Models**

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17

18 **Abstract**

19 We examine climate change impacts on net primary production (NPP) and  
20 export production (sinking particulate flux; EP) with simulations from nine Earth  
21 System Models (ESMs) performed in the framework of the fifth Coupled Model  
22 Inter-comparison Project (CMIP5). Global NPP and EP are reduced by the end of  
23 the century for the intense warming scenario of Representative Concentration  
24 Pathway (RCP) 8.5. Relative to the 1990s, NPP in the 2090s is reduced by 2-16%  
25 and EP by 7-18%. The models with the largest increases in stratification (and  
26 largest relative declines in NPP and EP) also show the largest positive biases in  
27 stratification for the contemporary period, suggesting overestimation of climate  
28 change impacts on NPP and EP. All of the CMIP5 models show an increase in  
29 stratification in response to surface-ocean warming and freshening, which is  
30 accompanied by decreases in surface nutrients, NPP, and EP.

31 There is considerable variability across the models in the magnitudes of NPP,  
32 EP, surface nutrient concentrations, and their perturbations by climate change.  
33 The negative response of NPP and EP to increasing stratification reflects  
34 primarily a bottom-up control, as upward nutrient flux declines at the global  
35 scale. Models with dynamic phytoplankton community structure show larger  
36 declines in EP than in NPP. This pattern is driven by phytoplankton community  
37 composition shifts, with reductions in productivity by large phytoplankton as  
38 smaller phytoplankton (which export less efficiently) are favored under the  
39 increasing nutrient stress. Thus, the projections of the NPP response to climate  
40 change are critically dependent on the simulated phytoplankton community  
41 structure, the efficiency of the biological pump, and the resulting levels of  
42 regenerated production, which vary widely across the models. Community  
43 structure is represented simply in the CMIP5 models, and should be expanded to

44 better capture the spatial patterns and climate-driven changes in export  
45 efficiency.

46

47 **1 Introduction**

48 Ocean net primary production (NPP) and particulate organic carbon export  
49 (EP) are key elements of marine biogeochemistry that are vulnerable to ongoing  
50 climate change from rising concentrations of atmospheric CO<sub>2</sub> and other  
51 greenhouse gases. Ocean warming has increasing impacts on ocean ecosystems  
52 by modifying the ecophysiology and distribution of marine organisms, and by  
53 altering ocean circulation and stratification. Ocean ecosystems also are important  
54 components of the climate system, influencing the atmospheric abundance of  
55 radiative agents such as CO<sub>2</sub>, N<sub>2</sub>O, aerosols, and the bio-optical properties of  
56 seawater (Siegenthaler and Wenk, 1984; Goldstein et al., 2003; Manizza et al.,  
57 2008; Schmittner et al., 2008; Bopp et al., 2013). Therefore, understanding the  
58 mechanisms controlling NPP and EP is essential for understanding the global  
59 cycles of carbon and other bioactive elements, and their links to climate. (Passow  
60 and Carlson, 2012).

61 Upper ocean stratification plays a key role in ocean biogeochemical processes.  
62 In particular, mixed layer depth (MLD) regulates the interplay between light  
63 availability for photosynthesis (Hannon et al., 2001) and nutrient supply to the  
64 upper ocean (Pollard et al., 2009). Upper ocean stratification is defined here as  
65 the density difference between the surface and 200 m depth (Capotondi et al.,  
66 2012), which is indicative of the degree of coupling and nutrient fluxes between  
67 the euphotic zone and the ocean interior. The density gradient at the base of the  
68 mixed layer affects entrainment processes, which play a crucial role in mixed  
69 layer deepening and in particle sinking/export from the euphotic zone.  
70 Stratification can also influence ocean ventilation (Luo et al., 2009), which has  
71 important consequences for oceanic uptake of carbon and oxygen. Thus, changes

72 in stratification over the remainder of the 21<sup>st</sup> century have the potential to  
73 influence NPP and EP across marine ecosystems.

74 Stratification tends to increase in response to ocean surface warming and  
75 freshening in 21st century climate change simulations. Increased stratification  
76 reduces the input of sub-surface nutrients to the euphotic zone and can lead to  
77 decreasing NPP and EP through increasing nutrient limitation. Many studies  
78 have suggested decreases in global NPP and EP over the 21 century using  
79 models with varying degrees of complexity (Bopp et al., 2001; Plattner et al., 2001;  
80 Fung et al., 2005; Schmittner et al., 2008; Steinacher et al., 2009; Dutkiewicz et al.,  
81 2013; Cabré et al., 2015). For the RCP8.5 scenario, CMIP5 ESM estimates of  
82 changes in export production range from -7 to -18% relative to the 1990s, and for  
83 NPP these changes range from -2 to -16% (Bopp et al., 2013).

84 The relative importance of different ecological controls on NPP and EP  
85 depends, in part, on an individual model's capacity to represent plankton  
86 functional types (PFT) (Le Quere et al., 2005; Jin et al., 2006) and their unique  
87 physiological and ecological characteristics, which determine how efficiently  
88 they are exported from surface waters. Increasing nutrient stress can shift  
89 phytoplankton community composition, favoring smaller phytoplankton, which  
90 are more efficient at nutrient uptake, over larger phytoplankton (Bopp et al., 2001;  
91 Steinacher et al., 2010; Vichi et al., 2011; Moore et al., 2013). These community  
92 shifts can modify the efficiency of carbon export. However, treatment of  
93 plankton communities is relatively simple in the CMIP5 models, with 1-3  
94 phytoplankton functional types and typically one zooplankton group (Bopp et al.,  
95 2013).

96 Several previous studies examined the biogeochemical response to climate  
97 change in the CMIP5 models. Bopp et al. (2013) examined output from 10 CMIP5

98 models emphasizing model mean biogeochemical responses to multiple stressors  
99 and trends over the 21<sup>st</sup> century relative to 1990s means for each model. Cabré et  
100 al. (2015) analyzed the CMIP5 models examining changes between model output  
101 averaged over the period 1980-1999 with years 2080-2099. This study broke down  
102 the global output into different ocean biomes for analysis. Laufkötter et al. (2015)  
103 also analyzed output from nine coupled climate-carbon ESMs, including many of  
104 the CMIP5 models to study how climate change processes impact NPP,  
105 comparing two twenty-year periods (2012-2031 and 2081-2100). They suggested  
106 strong roles for temperature and top-down grazing control in driving the NPP  
107 response, particularly at lower latitudes. Both Cabré et al. (2015) and Laufkötter  
108 et al. (2015) conclude that changing light levels were not a primary driver of  
109 changes in NPP except at the highest latitudes where there were large decreases  
110 in sea ice cover. Thus, we do not consider light effects in this work, where our  
111 focus is on global-scale trends. More detailed regional studies of the CMIP5  
112 model output have been carried out for the Arctic Ocean (Vancoppenolle et al.,  
113 2013) and the Southern Ocean (Hauck and Volker, 2015; Ito et al., 2015; Leung et  
114 al., 2015).

115 We analyzed centennial-scale changes in NPP and EP in response to  
116 increasing surface stratification and other physical factors. We use historical  
117 (1850-2005) and Representative Concentration Pathway (RCP) 8.5 (2006-2100)  
118 ESM simulations from the fifth phase of the Coupled Model Inter-comparison  
119 Project (CMIP5) to study long-term trends in NPP and EP and to identify the  
120 mechanisms behind these changes, including the physical factors that regulate  
121 nutrient availability. We also examined variability in NPP, EP, and surface  
122 nutrient concentrations across the models, to highlight some of the large

123 differences and uncertainties in projections of climate change impacts on marine  
124 biogeochemistry in current-generation ESMs.

125

## 126 **2 Methods**

127 We analyzed simulations from a set of 9 ESMs that contributed output to the  
128 Earth System Grid Federation as a part of CMIP5 (Taylor et al., 2012). Required  
129 physical ocean variables were temperature, salinity, and potential density;  
130 required biogeochemistry variables were macronutrients (nitrate, phosphate, and  
131 silicic acid), iron, chlorophyll, NPP and EP. The selection of the 9 models  
132 investigated here (Table 1) was based on the availability of these variables.

133 The historical and RCP8.5 simulations we analyzed had prescribed  
134 atmospheric CO<sub>2</sub> mole fractions and forcing from other greenhouse gases and  
135 aerosols, anthropogenic land use, and solar variability. Volcanic forcing also was  
136 included during the historical period. The RCP 8.5 is a strong warming scenario  
137 with an increase in radiative forcing of 8.5 W/m<sup>2</sup> by 2100 as atmospheric CO<sub>2</sub>  
138 mole fractions reach 936 ppm (Moss et al., 2010; van Vuuren et al., 2011). In the  
139 case where several ensemble members were available from an individual ESM,  
140 we analyzed only the first member.

141 A simple description of the 9 ESMs is presented in Tables 1 and 2.  
142 Atmospheric and ocean resolutions vary across the models (Table 1). Typical  
143 atmospheric horizontal grid resolution is ~2°, but it ranges from 0.94° to 3.8°.  
144 Typical ocean horizontal resolution is ~1°, ranging from 0.3° to 2°. In the vertical,  
145 there are 24 to 95 levels in the atmosphere and 31 to 63 levels in the ocean. All  
146 marine biogeochemical components are nutrient–phytoplankton–zooplankton–  
147 detritus (NPZD) models, but with varying degrees of complexity, illustrated for

148 instance by the number of phytoplankton functional groups (from 1 to 3) or  
149 limiting nutrients (from 3 to 5) that are explicitly represented (Table 2).

150 In our analysis, we used the CMIP5 variable denoting the vertical integration  
151 of NPP (intpp) and sinking export of organic particles at 100 m (EP; epc100). We  
152 present global mean estimates as the area-weighted or volume-weighted mean  
153 by the grid-cell area/volume from an individual model. Monthly mean data are  
154 averaged to obtain annual means and the annual mean data are interpolated  
155 onto a common  $1^\circ \times 1^\circ$  regular grid for the comparison of the 2-D fields.

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## 157 **3 Results**

### 158 **3.1 Stratification changes**

159 Stratification, defined here as the density difference between the depth of 200  
160 m and the surface, is a useful indicator of change in the upper ocean, as it  
161 integrates changes in both temperature and salinity. In Figure 1a, we present the  
162 time series of global mean stratification changes for the historical period and the  
163 RCP8.5 projection. All the models project an increase in stratification (ranging  
164 from 6% to 30% by the 2090s). However, the amplitude of stratification differs  
165 considerably across the models. The GFDL-ESM2M and MPI models are  
166 relatively close to the observed mean stratification in the WOA09 dataset (red  
167 square,  $1.81 \text{ kg/m}^3$ ) for the present era. NorESM1-ME shows the weakest  
168 stratification ( $1.74 \text{ kg/m}^3$ ) while the stratification in HadGEM2-ES is strongest  
169 ( $2.45 \text{ kg/m}^3$ ). Long-term trends are in general agreement across models, but the  
170 rate of stratification increase varies, with IPSL-CM5A-MR showing the most  
171 rapid increase and NorESM1-ME the slowest increase.

172 Surface processes that decrease density can largely explain the stratification  
173 increase in the RCP8.5 projections. Global mean sea surface temperature (SST)



174 warms by 2.6-3.5°C, accompanied by sea surface salinity (SSS) decreases of 0.05-  
175 0.25 psu over the 21<sup>st</sup> century (Figure 1). By 2100, the global mean SST ranges  
176 from 20.4°C (HadGEM2-ES) to 21.8°C (NorESM1-ME). Model spread decreases  
177 in the RCP8.5 projections in response to strong anthropogenic forcing (Figure 1b).  
178 SSS shows a clear declining tendency from 1850 to 2100 (Figure 1). Compared to  
179 the WOA09 observational data, most of the models are too fresh at the surface in  
180 the 1990s, especially the HadGEM2-ES, which has the lowest global mean SSS.  
181 The model spread is partly due to internal variability simulated by the climate  
182 models. Model differences in physics, but also in spin up procedures, the way  
183 RCP scenarios are set up, and model climate sensitivities all likely contribute to  
184 the model spread (Knutti and Hegerl, 2008; Szopa et al., 2013).

185 Vertical density profiles help to further explain the changes in stratification.  
186 Mean vertical profiles of density in the 1990s and the density change between the  
187 1990s and the 2090s show that all the models become more buoyant at the surface  
188 as a consequence of heating and/or freshening of the upper ocean (Figure S1).  
189 The density changes at the surface vary by almost a factor of two among models,  
190 from -1.1 kg/m<sup>3</sup> (HadGEM-ES2) to -0.6 kg/m<sup>3</sup> (GFDL-ES2M), but converge to a  
191 relatively narrow range (approximately -0.2 kg/m<sup>3</sup>) at 500 m depth. Most of the  
192 density change occurs between the surface and 200 m. Below 200 m, the density  
193 change in most of the models varies linearly with depth. Thus, our definition of  
194 the stratification index, as the density difference between the surface and 200 m,  
195 is reasonable. The converging reductions in density among models at about 500  
196 m agrees with some previous studies based on observations and CMIP3 models  
197 (Bindoff et al., 2007; Lyman et al., 2010; Capotondi et al., 2012). Compared to  
198 WOA09 data, the models generally underestimate the density of the upper ocean

199 in the top 150 m and most models overestimate the density below 350 m  
200 (resulting in a positive stratification bias) (Figure S1a).

201 Vertical profiles of temperature and salinity from each model are also shown  
202 in Figure S1. The surface ocean exhibits strong warming of 1.7-3.5°C by the 2090s  
203 and the warming magnitude declines quickly with depth, which is associated  
204 with the heat uptake capacity of individual models. For instance, the GFDL  
205 models seem to be more efficient in transporting heat downward than the IPSL  
206 models. Above 300 m, the temperature changes vary widely among the models.  
207 Temperature changes as a function of depth are complex, and model-to-model  
208 differences may be related to a number of factors including rates of vertical  
209 mixing and the seasonal thermocline dynamics. At the depth of 500 m, the mean  
210 temperature change converges at about 1.2°C. The ocean heat uptake capacity is  
211 linked to ocean diapycnal mixing and other processes, such as mixing by  
212 mesoscale eddies. The weak temperature gradients in the GFDL models suggest  
213 high rate of heat uptake, and are consistent with ocean heat uptake estimates by  
214 (Kuhlbrodt and Gregory, 2012). The large model spread in temperature profiles  
215 suggests considerable differences and uncertainties in the parameterizations of  
216 these physical processes across the models. Vertical profiles of salinity are more  
217 scattered than for temperature (Figure S1c). In the 1990s, most of the models  
218 underestimate salinity from the surface down to 550 m. Surface salinity is  
219 generally biased low by 0.05-0.25 psu. Most of the freshening with climate  
220 change takes place above 100 m, which also acts to increase stratification. Note  
221 that the salinity increases at 100-300m in some models (IPSL, GFDL-ESM2M,  
222 HadGEM2-ES) partially compensates the impact of rising temperatures on  
223 density.

224 The percentage contribution of temperature change to the stratification  
225 change from the 1990s to the 2090s is shown in Figure S2. Previous studies have  
226 shown that salinity contributes significantly to the stratification changes at high  
227 latitudes ( $>40^\circ$ ) in both hemispheres and in the North Pacific as a consequence of  
228 increases in precipitation (Bindoff et al., 2007). From our comparisons,  
229 temperature dominates the stratification changes in the tropical and sub-tropical  
230 regions (Figure S2). Salinity dominates the stratification changes in the much of  
231 the Arctic Ocean and in the high-latitude North Atlantic. While stratification is a  
232 function of SSS and SST to a good approximation (Cabr e et al., 2015),  
233 stratification change at high latitudes is also dependent on temperature and  
234 salinity at depth as vertical mixing and exchange are stronger.

235 In some regions the spatial distributions and the driving process differs  
236 substantially across models. Generally, the models agree well in the tropics and  
237 subtropical gyres that surface warming drives increasing stratification. In the  
238 high latitude North Atlantic, the subpolar Pacific and the western Pacific Ocean,  
239 there is weaker agreement across the models. In the subtropical gyre of the south  
240 Pacific, stratification changes in the IPSL and CESM1(BGC) models have a  
241 stronger influence from temperature change, while the other models exhibit  
242 more complicated spatial patterns. In the North Atlantic, salinity contributes  
243 more in the IPSL and HadGEM2 models than in the other models. The  
244 southeastern Pacific is more dominated by salinity in the two GFDL models. In  
245 the Southern Ocean, the models show relatively large contributions from both  
246 salinity and temperature but with complicated spatial patterns that differ  
247 considerably across models. Projections for the regions where the models do not  
248 agree even on the driving factor should be viewed with more caution. Climate-

249 change and biogeochemical impacts in these regions tend to be projected with  
250 less consistency across models (Bopp et al., 2013; Cabré et al., 2015).

251 Stratification increases globally in all the models with climate change  
252 (Figures 1 and 2). Nearly all the models predict large increases in stratification in  
253 the western tropical Pacific, the tropical Indian Ocean, the Arctic Ocean, and in  
254 the high latitude North Atlantic (particularly in the Labrador Sea). The Southern  
255 Ocean has weaker increases in stratification, partly because the surface layer  
256 mixing and upwelling are intensified due to the poleward shift of strengthened  
257 westerly jets (Swart and Fyfe, 2012). Our stratification index may underestimate  
258 the changes in the high latitude North Atlantic, as the relatively deep mixing  
259 means that temperature and salinity at 200m depth are changing much more  
260 rapidly than in other regions. Reductions in the deep winter mixing and NADW  
261 formation in this region are a common pattern seen in strong warming climate  
262 simulations (e.g., Cheng et al., 2013; Schwinger et al., 2014). Less drastic increases  
263 in stratification are seen over much of the rest of the oceans, with only a few  
264 small regions showing decreases in some models. An exception is the HadGEM2-  
265 ES model, which has large stratification reductions in the Arctic (Figure 2).

### 266 **3.2 Surface Nutrient Trends with Climate Change**

267 One of the key factors determining global NPP is nutrient availability in the  
268 euphotic zone. Time series of global mean nutrient (0-100m) concentrations for  
269 nitrate ( $\text{NO}_3$ ), phosphate ( $\text{PO}_4$ ), silicic acid ( $\text{SiO}_4$ ) and dissolved iron (dFe) are  
270 presented in Figure 3. The magnitude of surface nutrient concentrations differs  
271 substantially across the models (varying by a factor of ~1.5-2, and by a factor of 5  
272 for dissolved iron). The IPSL models have relatively low surface nutrient  
273 concentrations. Compared to the WOA09, 2 models overestimate phosphate  
274 (CESM1(BGC) and GFDL-ESM2G) and 5 models overestimate nitrate. All of the

275 models overestimate the silicic acid observations, with the exception of  
276 CESM1(BGC). The CESM1(BGC) model overestimates surface phosphate  
277 concentrations initially, due to excessive nitrogen limitation, but then shows the  
278 strongest surface phosphate declines over the 21<sup>st</sup> century (Figure 4).

279 Over the entire period from 1850-2100, the models all display decreasing  
280 trends for surface nitrate, phosphate and silicic acid. Interestingly, surface iron  
281 concentrations increase modestly in all but one of the models, by 4-10%. Changes  
282 in iron concentrations impact marine productivity, nitrogen fixation rates, and  
283 oceanic net CO<sub>2</sub> uptake. In the CMIP5 simulations, iron inputs to the oceans from  
284 deposition and rivers are held constant over time, so the increasing surface iron  
285 concentrations may reflect increasing macronutrient limitation of phytoplankton  
286 growth, leading to reduced biological uptake of iron. The reductions in the  
287 sinking export flux also reduce the particle scavenging loss term for dissolved  
288 iron. In the CESM1(BGC) model, increased production in the High Nutrient, Low  
289 Chlorophyll (HNLC) regions offset about 25% of the reduction observed in the  
290 macronutrient-limited areas with climate change, and changing circulation  
291 patterns also altered the lateral transport of iron within the oceans (Moore et al.,  
292 2013; Misumi et al., 2014).

293 The relative changes in nutrient concentrations (0-100m) (normalized to  
294 1990s means) are presented in Figure 4. The relative changes in the historical run  
295 show a consistent pattern across the models for nitrate, phosphate and dissolved  
296 iron (except for HadGEM2-ES). In the RCP8.5 projection, the models show  
297 diverging estimates of the magnitude of the relative changes. For nitrate, the  
298 reductions range between 3% and 14%, whereas for phosphate the reductions  
299 range between 3% and 20%. Silicic acid and iron trends are even more variable  
300 than for nitrate and phosphate. For silicic acid, 3 models exhibit slight increases,

301 while the others exhibit decreases ranging from 5-17%. The variability in relative  
302 change in silicic acid concentration in the RCP8.5 is likely associated with  
303 changes in plankton community and variable diatom production (Bopp et al.,  
304 2005). All of the models include some representation of diatoms (Table 2) but the  
305 match to observed silicic acid concentrations for the current era is generally poor  
306 (Figure 3).

307 The spatial distributions of mean nitrate concentration for 0-100 m in the  
308 1990s are shown in Figure S3. The CMIP5 models reproduce key observed  
309 features of the basin scale distributions of surface nitrate. For example, all of the  
310 models exhibit elevated nitrate concentrations in the eastern equatorial Pacific,  
311 Southern Ocean, subarctic North Atlantic and subarctic Pacific. In the subtropical  
312 gyres of the Atlantic and Pacific basins, mean nitrate concentrations are low.  
313 These general patterns are consistent with the WOA09 observations. However,  
314 there are clear disagreements in some regions. For example, the details of the  
315 high-nitrate surface water distributions vary considerably in the eastern  
316 equatorial Pacific. The HNLC condition extends too far north and south of the  
317 equator in some models, and too far to the west in others (Figure S3). The models  
318 also differ in the intensity and extent of high nitrate concentration waters in the  
319 subarctic North Pacific, where 6 of 9 models show lower nitrate concentrations  
320 than the WOA09 data (MPI-ESM-LR, MPI-ESM-MR and HadGEM2-ES are  
321 closest to the observations). There are also differences in the Arabian Sea and Bay  
322 of Bengal, where most models underestimate nitrate concentrations except the  
323 GFDL-ESM2M and MPI-ESM-LR models.

324 Inter-model spread in NPP during the 1990s is pronounced, with NPP as low  
325 as 29 PgC yr<sup>-1</sup> (IPSL-CM5A-LR and IPSL-CM5A-MR), while NPP in one model  
326 exceeds 75 PgC yr<sup>-1</sup> (GFDL-ESM2M) (Table 3, Figure 5). In addition, the spatial

327 pattern of NPP is not well represented by the multi-model mean (Bopp et al.,  
328 2013). Satellite based estimation of NPP is approximately 50 PgC yr<sup>-1</sup> (Behrenfeld  
329 et al., 2006; Carr et al., 2006). The MPI-ES-MR and CESM1(BGC) models had NPP  
330 of 49.8 PgC yr<sup>-1</sup> and 54.2 PgC yr<sup>-1</sup>, closer to the satellite-based estimates, and the  
331 observationally constrained, model estimate of 56 PgC yr<sup>-1</sup> by Buitenhuis et al.  
332 (2013). The magnitude of EP also varies substantially across models in the 1990s,  
333 ranging from 4.4 to 7.2 PgC yr<sup>-1</sup> (Table 3).

### 334 **3.3 Climate Change Impacts on NPP and EP**

335 All of the models exhibit decreasing trends in global NPP and EP with  
336 climate change, as shown in previous studies (Bopp et al., 2013; Dutkiewicz et al.,  
337 2013) and most models show more rapid decreases during the middle to latter  
338 part of the 21<sup>st</sup> century (Figures 5-6, Table 3). All nine models project decreases in  
339 export production by the 2090s, exceeding 5% relative to levels in the 1990s,  
340 whereas the response for NPP is divided into 2 groups after 2020. The  
341 CESM1(BGC) and GFDL models experience smaller changes in NPP (< 5%  
342 relative to 1990s) while other models have larger decreases (8-16%). The largest  
343 relative change for NPP is about -16% (MPI-ESM-LR). The EP decreases range  
344 from 7% for GFDL-ESM2G to 28% for IPSL-CM5A-LR. Cabré et al. (2015) found  
345 reductions in NPP and EP for all biomes, except at the highest latitudes. The  
346 reductions in global NPP and EP co-vary with the increases in stratification  
347 (Figure 6). By the 2090s, stratification increases by about 16% in GFDL-ESM2M  
348 and up to 33% in HadGEM1-ES. The rate of stratification increase is slower in the  
349 two GFDL models and CESM1(BGC), which also agrees with the slower rates of  
350 relative NPP and EP change.

351 The variability across models in NPP is substantially larger than that seen in  
352 EP (Table 3). The normalized standard deviation was  $\pm 27\%$  for NPP, but only  $\pm$

353 12 % for EP in the 1990s. The large spread in simulated NPP and its response to  
354 climate change was also noted by (Laufkötter et al., 2015). Seven of the nine  
355 models have an EP between 6.0 and 7.2 PgC yr<sup>-1</sup> in the 1990s, with the remaining  
356 two (HadGEM2-ES and GFDL-ESM2G) having considerably lower EP (< 5 PgC  
357 yr<sup>-1</sup>). EP is tightly coupled to new nutrient inputs to the euphotic zone in these  
358 models. NPP is less tightly coupled as the fraction of regenerated production  
359 varies across the models, and can vary spatially and temporally within some  
360 models. Thus, the large spread in NPP is not just a function of the different  
361 physical models and their transport of nutrients to the euphotic zone, but rather  
362 it is also strongly impacted by the phytoplankton community structure and  
363 export efficiency inherent in the models, and the resulting varying levels of  
364 regenerated production.

365       The sinking carbon flux out of the euphotic zone to net primary production  
366 ratio (particle export ratio or pe-ratio) is a measure of the export efficiency and  
367 also reflects the variable contribution of regenerated production to total NPP  
368 (Dugdale and Goering, 1967; Eppley and Peterson, 1979; Dunne et al., 2007).  
369 High pe-ratio values are typically associated with productive ecosystems  
370 dominated by larger phytoplankton (often diatoms; Buesseler, 1998; Boyd and  
371 Newton, 1999), while low pe-ratios are associated with oligotrophic food webs  
372 with most carbon flow through the microbial loop (Pomeroy, 1974; Azam et al.,  
373 1983). The CMIP5 models that include both large and small phytoplankton,  
374 assume a higher export efficiency for the large phytoplankton (Aumont and  
375 Bopp, 2006; Seferian et al., 2013; Tjiputra et al., 2013). The fraction of grazed  
376 material routed to sinking export is higher, often by a factor of 3 or more, than  
377 the fraction routed to sinking export for the small phytoplankton. Diatoms are  
378 also likely to dominate phytoplankton blooms in these models. This can drive



379 additional, very efficient, export through aggregation, further enhancing the  
380 differences in export efficiency between large and small phytoplankton. Relative  
381 to the 1990s, six of the nine models show decreasing trends in the pe-ratio (up to  
382 a 10% reduction) (Figures 5-6, Table 3; see also Cabré et al., 2015). Diatoms  
383 account for a smaller percentage of NPP in the 2090s than in the 1990s in all the  
384 models, except for the MPI model, where nearly all of the production is by  
385 diatoms and the smallest phytoplankton are not explicitly represented (Table 3).

### 386 **3.4 Increasing Stratification and Declining Nutrients, NPP, and EP**

387 We quantify the relations between stratification and key biogeochemical  
388 variables with annual model output over the entire time period of 1850-2100.  
389 This approach captures physical-biological interactions over interannual to  
390 centennial timescales, and is thus more robust than the earlier work comparing  
391 two end points from the beginning and end of the 21<sup>st</sup> century (Bopp et al., 2013;  
392 Cabré et al., 2015). Changes in global NPP relative to the 1990s are plotted  
393 against the relative change in stratification in Figure 7a. Across all the ESMs, a  
394 relatively good relationship exists with a correlation  $r^2=0.72$ . Larger relative  
395 increases in stratification correspond to larger relative declines in NPP. In  
396 addition, the globally-fitted line with a slope of 0.38 separates the models into  
397 two groups. In one group (GFDL, IPSL and CESM1(BGC)), the NPP reductions  
398 are more modest as stratification increases; the other group is composed of the  
399 two MPI models, HadGEM1-ES and the NorESM model, which show more  
400 intense and linear reductions in NPP with increasing stratification. The reduction  
401 of NPP can be partly explained by nutrient changes responding to stratification  
402 increases. Across the models, surface nitrate and phosphate concentrations  
403 clearly decline as the stratification is enhanced (Figure 7c and 7d, with  $r^2$  of 0.80  
404 and 0.82, respectively). Note that all of these trends are robust across the full time

405 series. Compared to the 1990s, the preindustrial stratification is weaker, surface  
406 nutrient concentrations are higher, and NPP and EP are elevated (Figures 3-7).  
407 This indicates significant climate change impacts on ocean biogeochemistry,  
408 prior to the 1990s and the modern era when most ocean observations have been  
409 made. The response of surface silicic acid to increasing stratification is more  
410 variable. The projected changes are more divided, as three models (MPI-ESM-LR,  
411 MPI-ESM-MR and HadGEM1-ES) show slight increases and the others show  
412 reductions in surface silicic acid concentrations (Figure 7b).

413 EP is even more closely related to the stratification changes ( $r^2=0.89$ ) than  
414 NPP ( $r^2=0.72$ ; Figure 7e). The EP change is also closely related to the NPP change  
415 ( $r^2=0.85$ ). EP decreases by up to 20% (Figure 7e) whereas NPP decreases by 10-  
416 18%. The models display two patterns in terms of the response of NPP and EP to  
417 climate change. The first group includes five models (the two IPSL models,  
418 CESM1(BGC) and the two GFDL models) where the relative declines in NPP are  
419 smaller than the relative declines in EP by a factor of 2 or more (Figure 6 and  
420 Table 3). In this group, the EP drops by about 10% and the NPP decreases by 5% .  
421 In the remaining models the relative declines in EP and NPP are larger and more  
422 similar in magnitude. For example, both EP and NPP decrease by about 14% in  
423 the HadGEM2-ES model. The differential declines in NPP and EP in the first  
424 group of models documents declining export efficiency for the ocean biological  
425 pump, driven by phytoplankton community shifts and a decreased contribution  
426 to NPP by large phytoplankton (diatoms) (see below and Figures 6-10; also Cabré  
427 et al., 2015).

428 Reduced nutrient availability is a major contributor to the declines in NPP  
429 and EP. However, the relationship varies from one model to another because  
430 growth and export are complicated functions of macronutrient limitation,

431 temperature, irradiance and iron limitation, as well as the routing of organic  
432 matter within the ecosystem that drives export efficiency. Higher metabolic rates  
433 with warming can be compensated to a large degree by changes in the supply of  
434 nutrients in terms of globally integrated productivity (Dutkiewicz et al., 2013).  
435 The NPP response is also strongly impacted by phytoplankton community  
436 structure, which modifies export efficiency, and the corresponding magnitude of  
437 regenerated primary production. For the IPSL, CESM1(BGC), and GFDL models  
438 that show larger declines in EP than in NPP, this pattern is driven by a  
439 decreasing contribution to total NPP by large phytoplankton (Table 3, Figures 8-  
440 9). Most of the primary production in these models is by smaller phytoplankton.  
441 The GFDL models express this pattern most strongly, with minimal declines in  
442 NPP, despite declines in EP approaching 10% (Figure 6 and Table 3). The other  
443 models tend to have production that is dominated by diatoms, and do not  
444 capture the community shifts towards increasing small phytoplankton  
445 dominance (and reduced export efficiency) under increasing nutrient stress. The  
446 declines in NPP with increasing stratification are more linear and more similar in  
447 magnitude to the declines in EP (Figure 7, panels a, b, and h). Thus, there are also  
448 very strong correlations between the climate-driven changes in the fractional  
449 contribution of diatoms to NPP and both the changes in stratification and the  
450 changes in EP (Figure 7, panels f and g, correlations of  $r^2=0.85$  and  $r^2=0.95$ , both  
451 much higher than the correlation between changing stratification and NPP,  
452  $r^2=0.71$ ). Cabré et al. (2015) found similar patterns relating community  
453 composition, NPP, and EP comparing the period from 1980-1999 with 2080-2099,  
454 across low to mid-latitude biomes.

455 Some of these patterns are illustrated in Figure 8, which shows the  
456 contribution of diatoms (large phytoplankton) to NPP for the 1990s. Most of the

457 models show elevated diatom production at high latitudes and low diatom  
458 contributions in the subtropical gyres. However, there are large discrepancies in  
459 the magnitude of the diatom contribution, ranging from about 30% to more than  
460 90% in the Arctic Ocean, for example. At the global-scale diatoms account for  
461 only 9.4% of NPP in the GFDL-ESM2M model and reach a maximum of 91% in  
462 the MPI-ESM-MR model (Table 3). The large variability across the models  
463 reflects, in part, the lack of observational datasets to constrain phytoplankton  
464 community composition, at the time these models were being developed. The  
465 new globally-gridded ocean atlas of plankton functional types, Marine  
466 Ecosystem Data (MAREDAT) (Buitenhuis et al., 2013) has started to fill this gap,  
467 and should lead to improved representations of plankton community structure  
468 in the future as the dataset becomes increasingly populated and is entrained into  
469 model development and validation. Remote sensing estimates of phytoplankton  
470 community composition and size class structures are also providing useful  
471 constraints for global-scale modeling efforts (e.g., Alvain et al., 2005; Hirata et al.,  
472 2008; Kostadinov et al., 2009; Siegel et al., 2014).

473 The spatial patterns of the shifts in phytoplankton community composition  
474 with climate change are illustrated in Figure 9; which shows the change in the  
475 percentage of NPP by diatoms (2090s – 1990s). There are some robust trends  
476 across the models. One of the areas with the biggest declines in diatom  
477 production is the high-latitude North Atlantic. This region typically has some of  
478 the biggest stratification increases with climate change, greatly reducing the deep  
479 winter mixing that entrains nutrients to the surface (Moore et al., 2013;  
480 Randerson et al., 2015). Nearly all the models also show large declines in diatom  
481 contributions to production in the Arctic Ocean. The CMIP5 models show  
482 consistent trends of increasing stratification, declining surface nutrient

483 concentrations, and a longer growing season with climate change in the Arctic  
484 (Vancoppenolle et al., 2013). Increasing surface temperatures and substantial  
485 declines in sea ice cover allow for a longer growing season with climate change.  
486 Thus, nutrients in surface waters are more completely used up by summer's end,  
487 leading to community shifts with decreased diatom production and an increased  
488 fraction of production by smaller phytoplankton. In the CESM1(BGC) model, this  
489 community shift allows for a small increase in central Arctic NPP, even as export  
490 production and surface nutrient concentrations decline, due to the increased  
491 fraction of NPP from small phytoplankton and the resulting increases in  
492 regenerated production (Moore et al., 2013).

493 All of the models show some increase in the fraction of NPP by diatoms in  
494 the Southern Ocean (Figure 9). The increase is particularly strong in the  
495 CESM1(BGC), IPSL, and GFDL models. Most of the models also show some  
496 increased diatom production in the tropical Pacific. Bopp et al. (2005) reported  
497 decreasing diatom production in the Arctic and high-latitude North Atlantic,  
498 with some increases in the Southern Ocean under a strong warming climate  
499 scenario. (Steinacher et al., 2010) also found declining productivity in the North  
500 Atlantic, and shifts in the export ratio due to phytoplankton community shifts  
501 with decreasing diatom production. An earlier version of the CESM used in that  
502 study (CCSM3) showed only small shifts in export ratios with climate change, as  
503 the range in export ratios and the differences in export efficiencies between large  
504 and small phytoplankton were smaller than in CESM1(BGC) (Steinacher et al.,  
505 2010; Moore et al., 2013). Three models in this study (HadGEM2-ES and the two  
506 MPI models) show increases in diatom production in the low latitudes (Figure 9).  
507 However, the diatoms dominate production nearly everywhere in these three  
508 models (Figure 8).

509        There are also large inter-model differences in the spatial patterns of the pe-  
510 ratio (Figure 10). Some of the models (GFDL, IPSL, CESM1(BGC)) show a close  
511 correlation between the pe-ratio and diatom production (compare Figures 8 and  
512 10), due to the enhanced export efficiency for diatoms (large phytoplankton)  
513 built into the models. Thus, there is a very high correlation between the changing  
514 contribution of diatoms to NPP and the changes in EP (Figure 7, panel g, Table 3).  
515 The MPI model includes one phytoplankton group and has an essentially  
516 constant pe-ratio of 0.15, explaining the linearity of the changes in NPP and EP  
517 with warming (Figures 8 and 10). Production in the HadGEM1-ES model is  
518 dominated nearly everywhere by the diatoms (Figure 8). Therefore, the MPI and  
519 HadGEM models cannot capture a shift towards increasing small phytoplankton  
520 dominance under declining surface nutrient concentrations. This leads to export  
521 production being closely correlated with diatom production in these models as  
522 most production is by diatoms, as well as in the other models where diatoms are  
523 assumed to export more efficiently but account for a smaller fraction of total NPP  
524 (Table 3).

525        There is also a strong correlation between the declines in the fraction of NPP  
526 by diatoms and declines in the pe-ratio (compare Figures 7, 9 and 11). The largest  
527 declines in the pe-ratio are seen in the Arctic and the high-latitude North Atlantic,  
528 regions where diatom production also decreases the most. The GFDL, IPSL, and  
529 CESM1(BGC) models show some reductions in pe-ratio in the subarctic North  
530 Pacific, but the spatial patterns are inconsistent (Figure 11). The models display  
531 considerable variability in the degree of stratification increase and in the  
532 dominant factor driving these changes in the subarctic North Pacific (Figures S2  
533 and 2).

534 The correlation for the relationship between the changing percentage of NPP  
535 by diatoms versus the changes in EP across all the models has an  $r^2$  of 0.96 and a  
536 slope close to 1 (0.94, Figure 7g), indicating that phytoplankton community  
537 structure plays a dominant role in determining the responses of NPP, EP, and the  
538 pe-ratio to climate change. The biggest declines in the fraction of production by  
539 diatoms and pe-ratios are in precisely the areas where some of the largest  
540 increases in upper ocean stratification are seen, along with declining surface  
541 nutrient concentrations, as in the Arctic Ocean and in the high latitude North  
542 Atlantic (Figures 6-8; see also Steinacher et al. (2010); Moore et al. (2013)).

### 543 **3.5 Projected Changes in NPP, EP and Stratification Biases**

544 At the global scale, the CMIP5 models show considerable stratification biases  
545 for the 1990s when compared to the WOA09 data (Figure 1, Table 3). Only the  
546 GFDL-ESM2M model is within 10% of the observed value. From the density  
547 profiles as well (Figure S1) it is apparent that most of the models have stronger  
548 stratification in the 1990s than seen in the observations. Liu et al. (2014) argue  
549 that climate bias is important when projecting the impact of climate change on  
550 land surface processes and Hoffman et al. (2014) document this for atmospheric  
551 CO<sub>2</sub> mole fractions. Here, we examine how stratification biases in the 1990s may  
552 affect model projections of NPP and EP in the 2090s.

553 Models with stronger bias in the 1990s for surface stratification tend to  
554 predict larger climate-induced declines in both NPP and EP (Figure 12,  $r^2=0.47$   
555 and  $r^2=0.54$ , respectively). The slopes are plotted when the correlation is  
556 significant at >95% level. Five of the models have positive biases in stratification  
557 for the current era that exceed 20%. These models also show the largest relative  
558 increases in stratification with climate change of 26-30% (Figure 12, Table 3). The  
559 remaining four models (GFDL models, CESM1(BGC), and NorESM1-ME) do a

560 better job of simulating observed stratification for the current era, and predict  
561 relative increases in stratification over the 21<sup>st</sup> century that are roughly half as  
562 large, ranging from ~15-18%. This suggests that the more biased models (for the  
563 1990s) may be overestimating the projected reductions in NPP and EP for the end  
564 of the century.

565

#### 566 **4 Discussion and Conclusions**

567 The ESMs analyzed here have different resolutions and incorporate marine  
568 biogeochemical-ecosystem models with different mechanisms and degrees of  
569 complexity. We find this set of models has consistent trends of increasing  
570 stratification and decreasing NPP and EP. However, a large model spread is  
571 apparent for the 1990s, particularly for NPP, and in the relative changes to NPP  
572 and EP over the 21<sup>st</sup> century due to climate change. NPP is reduced by 2-18% in  
573 the 2090s and EP is reduced by 7-20%. Mean stratification increased by 16%  
574 (GFDL-ESM2M) up to 33% (HadGEM1-ES) from the 1990s to the 2090s. Under  
575 strong warming scenarios like RCP8.5, ocean stratification will continue to  
576 rapidly increase after the year 2100 in all of these models (Randerson et al., 2015).

577 The strongly linear relationship between stratification increases and EP  
578 decreases seen within each model and across all the models (Figures 7 and 12)  
579 indicates a strong bottom up control on EP, through declining upward nutrient  
580 flux to the euphotic zone. Declining surface nutrient concentrations are seen in  
581 all the models with climate change under the RCP 8.5 scenario (Figures 5-6).  
582 Nitrate is reduced by 3 to 14% and phosphate is reduced by 3 to 20%. Changes in  
583 surface silicic acid and iron concentrations are more variable across the models.  
584 For silicic acid, there are 3 models showing slight increases, while the others  
585 exhibit decreases of 5-17%. With respect to iron, 8 models indicate an increase of



586 4-10% relative to the 1990s; with the exception being the NorESM-ME model,  
587 which is reduced by 3%. Changes in the temperature and light fields also have  
588 impacts on EP in some regions, but increasing stratification and nutrient stress,  
589 and the resulting impacts on phytoplankton community composition and EP is  
590 the dominate process at the global scale. On a global scale, over the full 1850-2100  
591 time period, the changes in NPP and EP are more highly correlated with the  
592 changes in stratification, than with the changes in SST ( $r^2$  0.72 for stratification-  
593 NPP and 0.66 for SST-NPP, Figure 7). This is because that the stratification metric  
594 captures both the temperature-driven changes that dominate at low to mid-  
595 latitudes, and the salinity-driven changes at higher latitudes. The temperature-  
596 driven increases in growth rates are partially offset by reduced nutrient supply  
597 in many regions as stratification increases (Bopp et al., 2005; Cabré et al., 2015).

598 Simulated NPP and its response to climate change are both more variable  
599 across the models than EP, and are less strongly correlated with changes in  
600 stratification (Figure 7). This is driven by model differences in the export  
601 efficiency of the biological pump and its relation to phytoplankton community  
602 structure. The models that allow for shifts in phytoplankton community  
603 structure show strongly non-linear NPP response to climate change. NPP  
604 declines less rapidly than EP with increasing nutrient stress, as the percentage of  
605 NPP by large cells declines and export efficiency decreases (and the regenerated  
606 production fraction increases). Models without this dynamic community  
607 composition and export efficiency show a much more linear NPP response to  
608 climate change (Figure 7). Thus, projections of the response of NPP to climate  
609 change in the CMIP5 models are critically dependent on the simulated  
610 phytoplankton community structure, the efficiency of the biological pump, and  
611 the resulting (highly variable) levels of regenerated production.

612 Spatial patterns of diatom productivity are influenced by changes in surface  
613 nutrients and the resulting shifts in plankton community composition. The  
614 response of the %NPP by diatoms depends on several factors, including whether  
615 they were a small or large component of the community initially. Therefore, the  
616 spatial patterns of changes in stratification and %NPP by diatoms can differ  
617 (Figure 2 and Figure 9). The largest decreases are seen in areas with high diatom  
618 production initially and large increases in stratification, particularly in the  
619 Northern Hemisphere, leading to a North-South hemispheric asymmetry  
620 (Marinov et al., 2013; Cabré et al., 2015). In the Southern Ocean, the winds that  
621 drive upwelling, strengthen and shift poleward with climate change, influencing  
622 iron supply and productivity patterns (Moore et al., 2013; Misumi et al., 2014).

623 The large spread in the simulated NPP rates for the 1990s and the variability  
624 seen across models in the response of NPP to climate change introduce  
625 challenges for climate impact and risk assessment, as NPP is a key product of  
626 both terrestrial and marine ecosystem models, and changes to NPP are perhaps  
627 the most cited result from this class of models. We have demonstrated that the  
628 wide spread seen in simulated NPP is not due to the different physical  
629 circulation models and the flux of nutrients they deliver to surface waters, but  
630 rather to the efficiency of the biological pump (tied to community structure in  
631 most models) and the resulting levels of regenerated primary production.  
632 Changes in EP are an additional useful metric of climate impacts on marine  
633 ecosystems. EP is more strongly tied to the climate feedback, as it is mainly the  
634 fixed carbon sequestered to the deeper ocean by the biological pump that will  
635 impact air-sea CO<sub>2</sub> exchange. In addition, in terms of impacts up the food chain,  
636 EP may be a better metric than NPP. Friedland et al. (2012) demonstrated that  
637 there is no correlation between fishery yield and NPP at the global scale, but that

638 there are strong correlations between fishery yield and several other variables  
639 including chlorophyll concentration, the pe-ratio, and EP. These three proxies all  
640 correlate with the fraction of primary production by large phytoplankton. In this  
641 context, the results presented here suggest large future declines in fishery yield  
642 across the high-latitude North Atlantic.

643 Laufkötter et al. (2015) suggest a strong impact of temperature modification  
644 of phytoplankton growth rates and other ecosystem processes (including  
645 zooplankton growth and grazing rates) to infer a strong top-down grazing  
646 influence on the NPP response to climate change, noting that phytoplankton  
647 growth rates appear to increase at low latitudes in some models, even as  
648 available nutrient concentrations decline. However, many of the key fluxes and  
649 fields needed to support this hypothesis were not available in the archived  
650 output from the CMIP5 models. This study relied on estimated nutrient  
651 limitation factors and growth rates for only the surface ocean in their analysis.  
652 Temperature warming is strongest at the surface (Figure 1S). Thus, the analysis  
653 may overestimate the temperature impacts for the whole euphotic zone. Their  
654 conclusions were based on diatom-specific nutrient limitation patterns, on the  
655 phytoplankton group with the largest changes in limitation factors, and on  
656 comparing total grazing with total NPP for some models (Figures 6-8, Laufkötter  
657 et al. (2015)). These may not be representative of the community nutrient  
658 limitation patterns and growth response. At low latitudes the diatoms might  
659 show the biggest declines in growth due to nutrient limitation, but they are only  
660 a small component of the community in many of the models (Figure 8). Under  
661 increasing nutrient stress, phytoplankton community growth rates may increase  
662 simply due to a declining contribution from diatoms, as the smaller  
663 phytoplankton will typically grow faster at low nutrient concentrations. Grazing

664 rates are also higher on the small phytoplankton in models with multiple groups.  
665 Thus, comparing total grazing rates to NPP cannot account for these influences  
666 of phytoplankton community shifts. A community shift towards smaller  
667 phytoplankton will likely increase mean community growth rate and total  
668 grazing, even with no change in temperature.

669 We agree that temperature effects may be important in the NPP climate-  
670 change response, and that the temperature influence on phytoplankton growth  
671 rates, and on the ecosystem processing of NPP that leads to export, are highly  
672 uncertain (Laufkötter et al., 2015). Sherman et al. (2016) compiled in situ  
673 estimates of phytoplankton community growth rates at the global-scale, and  
674 found a relatively weak apparent temperature effect (apparent  $Q_{10} \sim 1.5$ ). These  
675 observational estimates of phytoplankton community growth rates were  
676 compared with the CESM and GFDL CMIP5 simulations analyzed here. ESMs  
677 used in climate change studies need to ensure that the emergent, community  
678 temperature-growth relation matches this observed value (even though higher  
679 explicit  $Q_{10}$  values may be prescribed for individual plankton functional types)  
680 to avoid biases in the response to temperature change (Sherman et al., 2016).

681 Some of the CMIP5 models have an assumed higher export efficiency for  
682 diatoms relative to small phytoplankton, building on a long-standing paradigm,  
683 strengthened by results from the detailed ecosystem studies of the Joint Global  
684 Flux Study (JGOFS) program (Buesseler, 1998; Boyd and Newton, 1999). In the  
685 current models, the spectrum of phytoplankton size structure is often  
686 represented very simply with only the end members of one large and one small  
687 phytoplankton group. Thus, the “diatom” group is a proxy for larger, efficiently-  
688 exporting, blooming phytoplankton functional types. DOM cycling,

689 heterotrophic bacteria, microzooplankton, and the microbial loop are typically  
690 treated in an idealized, implicit manner in the current models as well.

691 To accurately predict the response of NPP and EP to climate change, it may  
692 be necessary to develop more robust ecosystem models with additional explicit  
693 phytoplankton, heterotrophic microbial, and zooplankton groups, including their  
694 impacts on nutrient cycling, export efficiency and the downward transport of  
695 organic matter. Models that include much greater diversity in the phytoplankton,  
696 show large community composition shifts with climate change (Dutkiewicz et al.,  
697 2013). Quantifying the links between NPP, EP and community composition in  
698 observational datasets is a high priority. There are only limited field observations  
699 of the pe-ratio, some of which rely on nutrient drawdown and other indirect  
700 estimates of the sinking particle flux (Dunne et al., 2007). Further progress to  
701 improve model performance requires combined efforts from satellite, field, and  
702 laboratory observations, empirical and inverse modeling approaches, as well as  
703 process-based, forward models.

704 The large model spread in EP and NPP, and significant biases seen in key  
705 nutrient fields for the 1990s suggest that the current ocean biogeochemical  
706 models are far from perfect and their results must be interpreted with caution.  
707 However, the relationships between stratification and EP, NPP and nutrients do  
708 reveal some common mechanisms driving the climate change response. The  
709 large inter-model differences for the current era in NPP, EP and nutrient  
710 concentrations are partially associated with how these biogeochemical models  
711 are initialized and spun up for these experiments. The ocean biogeochemical  
712 models are often integrated in an offline mode for a thousand years or more  
713 before coupling to other components of the ESM (Séférián et al., 2016). The  
714 achieved preindustrial, near-steady state of biogeochemical fields may deviate

715 substantially from the observed climatology, driven by biases in the physics and  
716 biogeochemistry. These differences typically persist in the present-day  
717 simulations and future projections. The advantage of the initialization and spin  
718 up process is that the biogeochemical fields are consistent with the simulated  
719 ocean circulation, and will respond to climate-driven changes appropriately. The  
720 strong intrinsic variability helps to reduce model drift and generate reasonable  
721 longer-term variability. As a result, these long-term simulations are suitable for  
722 analyzing climate trends, variability and sensitivities. RCP 8.5 is a strong  
723 warming scenario and the relationship between stratification changes and  
724 NPP/EP changes may be somewhat different under other RCP scenarios.  
725 Although the relations between the degree of surface warming and the ocean  
726 biogeochemical responses were largely linear across RCP 4.5 and 8.5 for the  
727 CESM(BGC) (Moore et al., 2013).

728       Some potentially important marine biogeochemical feedbacks on the climate  
729 system were missing completely or not well represented in the CMIP5 models,  
730 including important feedbacks through aerosol transport and deposition on the  
731 marine iron cycle, feedbacks involving the oxygen minimum zones and the  
732 marine nitrogen cycle, and the impacts on ocean biology by ongoing ocean  
733 acidification. Each of these feedbacks could impact phytoplankton and  
734 zooplankton community structures, NPP, EP, and pe-ratios in the future.

735       It is also important to consider the longer-term climate change responses of  
736 both ocean physics and marine biogeochemistry. Moore et al. (2013) noted that  
737 climate impacts on the oceans were still accelerating at year 2100 under the RCP  
738 8.5 scenario (but not under the more moderate RCP 4.5 scenario). Randerson et al.  
739 (2015) extended the CESM1(BGC) RCP 8.5 scenario simulation examined here, to  
740 the year 2300. In these longer simulations, the climate impacts on ocean physical

741 fields and biogeochemistry lead to even stronger perturbations after 2100 than  
742 those presented here. In addition, the ocean contribution to the climate-carbon  
743 feedback exceeded the land contribution after the year 2100 (Randerson et al.,  
744 2015).

745

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756

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1001

1002 Figure Captions

1003

1004 **Figure 1.** Time series of global mean surface stratification, SST and SSS for  
1005 historical run and RCP8.5 over 1850-2100. Surface stratification is defined as the  
1006 density difference between 200m and the surface. Red square indicates  
1007 observations from the WOA2009 data.

1008

1009 **Figure 2.** The spatial pattern is shown for changes in stratification between the  
1010 1990s and the 2090s.

1011

1012 **Figure 3.** Time series of nitrate ( $\text{NO}_3$ ), phosphate ( $\text{PO}_4$ ), silicate ( $\text{SiO}_4$ ) and  
1013 dissolved iron (dFe) concentrations (0-100 m) are shown for 1850-2100. Red  
1014 square indicates WOA2009 global mean values.

1015

1016 **Figure 4.** Time series are displayed of mean changes (in percent) relative to the  
1017 1990s for (a)  $\text{NO}_3$ , (b)  $\text{PO}_4$ , (c)  $\text{SiO}_4$  and (d) dFe (0-100m) during 1850-2100.

1018

1019 **Figure 5.** Time series of global mean net primary production, export production,  
1020 and the particle export ratio over 1850-2100 are shown for each model.

1021

1022 **Figure 6.** Time series are displayed of the percent changes in net primary  
1023 production, export production, and the particle export ratio, and stratification  
1024 over the period 1850-2100 (each relative to their 1990s means).

1025

1026 **Figure 7.** Relationships are shown between the relative percent change in surface  
1027 stratification with climate and the relative change in several biogeochemical  
1028 variables including net primary production (NPP) (a), silicate (b), nitrate (c),  
1029 phosphate (d), export production (EP) (e), the fraction of NPP by diatoms (g). EP  
1030 is plotted against the change in the fraction of NPP by diatoms (g) and against  
1031 the change in NPP (h). All changes are relative to the 1990s and plotted over  
1032 1850-2100. These time series are derived from global annual mean data.

1033

1034 **Figure 8.** The fraction of total NPP by diatoms for the 1990s is shown for each  
1035 model (data for NorESM not available).

1036

1037 **Figure 9.** The percent change in NPP by diatoms between the 2090s and the  
1038 1990s.

1039

1040 **Figure 10.** The mean particle export ratio (pe-ratio) for the 1990s is shown for  
1041 each model.

1042

1043 **Figure 11.** The percent change in the particle export ratio (pe-ratio) between the  
1044 2090s and the 1990s).

1045

1046 **Figure 12.** The stratification bias for the 1990s is plotted for each model versus  
1047 the relative changes in NPP (a), EP (b), and stratification (c) with climate change  
1048 (2090s – 1990s). All three linear regressions are statistically significant at a level  
1049 >95%.

1050

1051

### 1052 **Supplementary Figure Captions**

1053

1054 **Figure S1.** Mean vertical profiles are shown for density (a), temperature (c) and  
1055 salinity (e) for the 1990s. Changes between the 2090s-1990s are shown in (b), (d)  
1056 and (f), for the same variables. Solid black line denotes WOA2009 data.

1057

1058 **Figure S2.** Fractional contribution of temperature to the stratification change  
1059 from the 1990s to the 2090s is shown for each model.

1060

1061 **Figure S3.** Mean NO<sub>3</sub> concentrations in the first 100 m for the 1990s, the modeling  
1062 efficiency (MEF) (Stow et al., 2009) and logarithmic transformed root mean  
1063 square error (RMSE) are indicated relative to observations from the WOA2009.

1064

1065 **Supplementary References:**

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