

Biogeochemical variability in the California Current System

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Abstract

The biological productivity and diversity of the California Current System (CCS) is at the leading edge of major emerging climate trends, including hypoxia and acidification. We present results from a hindcast simulation (reanalysis) of an eddy-resolving oceanic physical-biogeochemical model of the CCS, to characterize its mean state and its patterns and drivers of variability in marine biogeochemical and ecosystem processes from 1995-2010. This is a companion paper to a physical analysis in Renault et al. (2021). The model reproduces long-term mean distributions of key ecosystem metrics, including surface nutrients and productivity and subsurface O_2 and carbonate undersaturation. The spatial patterns of Net Primary Productivity (NPP) are broadly consistent with measured and remotely sensed rates, and they reflect a predominant limitation by nitrogen, with seasonal and episodic limitation by Fe nearshore in the central CCS, and in the open ocean northern CCS. The vertical distribution of NPP is governed by the trade-off between nutrient and light limitation, a balance that reproduces and explains the observed spatial variations in the depth of the deep Chl maximum. The seasonal to interannual variability of biogeochemical properties and rates is also well captured by model simulations. Because of the prevailing nutrient limitation, fluctuations in the depth of the pycnocline and associated nutricline are the leading single factor explaining interannual variability in the interior biogeochemical state, and the relationships between density and biogeochemical rates and tracers are consistent between model and observations. The magnitude and relationship between density structure and biogeochemical processes is illustrated by the 1997-98 El Niño event, which faithfully reproduces the single largest deviation from the mean state in the simulated period. A slower decadal shoaling of the pycnocline also accounts for the concomitant trends in hypoxic and corrosive conditions on the shelf. The resulting variability is key to understanding the vulnerability of marine species to oceanic change, and to the detection of such changes, soon projected to exceed the range of conditions in the past century.

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1 Introduction

Coastal upwelling systems along the eastern boundaries of the subtropical oceanic basins are some of the most climatically and biologically dynamic regions of the world’s oceans (Carr and Kearns, 2003; Kudela et al., 2008). In the California Current System (CCS) of the Eastern North Pacific, the seasonal cycle of alongshore winds and offshore surface currents yields a direct conduit for nutrient-rich water to rise from the deep ocean to the sunlit surface (Chavez and Messie, 2009). The upwelling of cold water near the coast is evidenced in satellite remote sensing of sea surface temperature, expansive offshore marine stratocumulus clouds, and a nearshore ribbon of high chlorophyll from phytoplankton. The resulting cascade of phytoplankton biomass up the food chain supports high biodiversity and productive fisheries (FAO, 2009).

The physical state of the Northeastern Pacific varies on time scales of days to decades. This variability includes mesoscale eddies, seasonal cycle, interannual El Niño Southern Oscillation (ENSO) variability, and lower frequency climate fluctuations characterized by the Pacific Decadal Oscillation (PDO, Mantua et al. (1997)) and the North Pacific Gyre Oscillation (NPGO, Di Lorenzo et al. (2008)). Ecosystems throughout the Eastern Pacific respond strongly to physical forcing at each of these timescales, through the physical influence of winds, light, and heat, and their effects on the supply of nutrients to phytoplankton and oxygen to marine animals. Interannual climate cycles associated with ENSO in particular are major perturbations to these parameters and thus to plankton productivity (Chavez et al., 2002; Bograd and Lynn, 2001; Turi et al., 2018). Because thermocline waters entering the CCS are, like other subtropical Eastern Boundary Upwelling Systems, far from the sites of atmospheric ventilation in the west, the rising waters are “old” and bear the signature of decades of biogeochemical process that yield low O_2 , high nutrients, and low pH. The upwelling in eastern boundary systems also generates energetic mesoscale eddy fields (Capet et al., 2008). These eddies can transport the extreme properties of CCS thermocline waters far offshore and down into the subtropical interior, connecting the biogeochemistry of the coastal CCS with the adjacent oceanic gyres (Nagai et al., 2015; Gruber et al., 2011; Renault et al., 2016a; Frenger et al., 2018).

Oceanic acidification and deoxygenation are also emerging trends in the CCS ecosystem (Gruber et al., 2012; Chan et al., 2008). Anthropogenic CO_2 has been detected in coastal subsurface waters off Northern California (Feely et al., 2008). Decadal trends in oxygen have also been observed in the California Current (McClatchie et al., 2010; Pierce et al., 2012; Dussin et al., 2019) and have altered the proportions of biologically important nutrients (Deutsch et al., 2011). In the Northern California Current, massive benthic die-offs have been attributed to episodes of extreme hypoxia along the Oregon shelf (Chan et al., 2008). Fluctuating abundance of species in upper trophic levels observed over decadal and longer time scales arise from climate variability, but the specific mechanisms remain obscure (Chavez et al., 2003; Rykaczewski and Checkley, 2008). In addition to climate forcing, the coastal ocean is subjected to anthropogenic pollution that could locally exacerbate hypoxia and acidification (Doney et al., 2007), and may contribute to the increasing frequency and toxicity of harmful algal blooms along California’s coast (Andersson et al., 2008).

The continental margin of western North America has a rich endowment of historical and ongoing observational programs aiding evaluation of climate-ecosystem interactions, including one of the most extensive time series programs (the California Cooperative Ocean Fisheries Investigations, CalCOFI) anywhere in the world’s oceans (Ducklow et al., 2009). The age of shallow

82 thermocline waters also make them prone to large amplitudes of low-frequency variability (Ito and
83 Deutsch, 2010) that confound the detection of climate and other anthropogenic changes imping-
84 ing through the oceanic surface. Thus, even in this relatively well-studied region of the ocean,
85 it is difficult to distinguish long-term trends associated with anthropogenic climate change from
86 the low-frequency variability that pervades oceanic properties in regions of close contact with old,
87 deep waters. For example, it remains unclear whether the changes in hypoxia in the California Cur-
88 rent are driven by internal climate variability, or are an early sign of long-term climate warming
89 (McClatchie et al., 2010; Pierce et al., 2012; Long et al., 2016).

90 Models of this dynamic region often exhibit substantial biases in the mean state and unknown
91 fidelity in representing historical variability and its causal mechanisms. The purpose of this pa-
92 per is to report a system-wide validation of eddy-resolving, regional model fields and property
93 relationships through comparison to a variety of hydrographic, experimental, and remote sensing
94 observations. Here we focus on biogeochemical aspects of the model solution; the physical dy-
95 namics and its model validation are discussed in a companion paper (Renault et al., 2021). In this
96 work, we demonstrate the fidelity of modeled spatial patterns and seasonal to interannual variabil-
97 ity to observational datasets. We further probe the mechanisms underlying this variability—better
98 understanding these drivers is critical to attributing and projecting the biogeochemical responses of
99 the CCS to natural climate fluctuations and anthropogenic change. Section 2 provides a description
100 of the model, its boundary conditions, the simulations performed, and the datasets used for model
101 evaluation. These constitute the core results described in Section 3 and summarized in the final
102 Section 4.

103 **2 Methods**

104 **2.1 Model description**

105 The ecosystem and biogeochemical cycles are simulated in the Regional Ocean Modeling Sys-
106 tem (ROMS, Shchepetkin and McWilliams (2005)). As in Renault et al. (2016b), the primary
107 domain extends from 144.7°W to 112.5°W and from 22.7°N to 51.1°N. Its grid is 437 x 662
108 points with a horizontal resolution of $dx = 4$ km and 60 vertical levels. Initial and horizontal
109 boundary data for temperature, salinity, surface elevation, and horizontal velocity are taken from
110 the quarter-degree, daily-averaged Mercator Glorys2V3 data-assimilating ocean reanalysis (acces-
111 sible via <http://www.myocean.eu>; described further at [https://www.mercator-ocean.fr/en/science-](https://www.mercator-ocean.fr/en/science-publications/glorys/)
112 [publications/glorys/](https://www.mercator-ocean.fr/en/science-publications/glorys/)). In order to maintain a realistic representation of the variability in water mass
113 properties and transport into the model domain over time, monthly anomalies from the Mercator
114 data are added to the mean monthly climatology from the World Ocean Atlas (WOA, Locarnini
115 et al. (2013); Zweng et al. (2013)) over the period 1995-2004. The freshwater, turbulent heat,
116 and momentum fluxes are estimated using bulk formulae (Large, 2006) and the atmospheric fields
117 derived from an uncoupled simulation with the Weather Research and Forecasting model (WRF).
118 Heat and momentum fluxes are computed from bulk formulae, as detailed in Renault et al. (2021).
119 The freshwater flux from river runoff is included as surface precipitation and is spread using a
120 Gaussian distribution over the grid cells that fall within the range from the coast to 150 km off-
121 shore; this excludes a detailed representation of river plumes. The river-runoff forcing dataset we
122 use is a monthly climatology from Dai et al. (2009). The river inputs are assumed to carry negli-

123 gible chemical concentrations, except the outflow from the Juan de Fuca Strait (see below). This
 124 statistically equilibrated solution, named USW4, is integrated over the period 1995-2010 after a
 125 spin up of 1 year (with initial conditions derived from the World Ocean Atlas).

126 Further details are described in the companion paper (Renault et al., 2021).

127 The coastal biogeochemical dynamics are simulated using an ecosystem model (the Biogeo-
 128 chemical Elemental Cycling (BEC) model, Fig. 1 and Appendix). This model includes both phy-
 129 toplankton and zooplankton, and dissolved, suspended, and sinking particulates (Moore et al.,
 130 2004). The model includes four phytoplankton functional groups (picoplankton, diatoms, coc-
 131 colithophores, and diazotrophs) characterized by distinct biogeochemical functions (nutrient re-
 132 cycling, silicification, calcification, and N_2 fixation, respectively). Four nutrient cycles (nitrogen,
 133 silicic acid, phosphate, and iron) are simulated and are coupled through a fixed phytoplankton sto-
 134 ichiometry, except for iron (Fe), which varies in proportion to the other nutrients (see Equations
 135 A100-109; Moore et al. (2002, 2004)). The ecosystem is linked to an oceanic biogeochemistry
 136 module that includes total inorganic carbon (DIC), alkalinity, iron, and dissolved O_2 . Reminer-
 137 alization of sinking organic matter is parameterized according to the mineral ballast model of
 138 Armstrong et al. (2001). Gas exchange fluxes for O_2 and CO_2 are based on Wanninkhof (1992).
 139 The BEC equations are listed in the Appendix, and the model code, including parameter settings,
 140 are available through the GitHub repository (see the remark at the end of the paper).

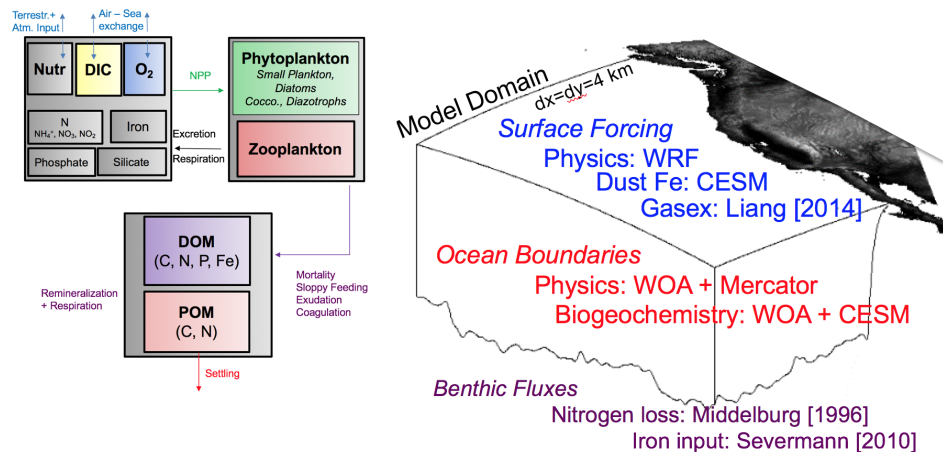


Figure 1: Schematic structure and physical configuration of ROMS-BEC biogeochemical model. (a) The main ecosystem state variables and fluxes. (b) Geographic scale of simulation, and sources of surface forcing, open boundary condition data and representations of benthic nutrient fluxes.

141 The iron Fe cycle includes dissolved iron, scavenged iron, and iron associated with organic
 142 matter pools and dust particles, but only dissolved iron and organically bound iron are explic-
 143 itly modeled as state variables. For dissolved iron, four processes are considered: atmospheric
 144 deposition, biological uptake and remineralization, scavenging by sinking particles, and release
 145 by sediments. Atmospheric iron deposition is based on the dust climatology of Mahowald et al.
 146 (2006). We implemented a sedimentary iron source based on benthic flux chamber measurements
 147 in the California margin. An equation relating sediment Fe release as a function of bottom water
 148 O_2 ($\log_{10}[F_{fe}] = 2.5 - 0.0165 \cdot O_2$, where O_2 is in mmol m^{-3} and the efflux units are $\mu \text{mol m}^{-2}$

149 d^{-1}) is derived from data compiled by Severmann et al. (2010). The resulting rates of Fe sup-
 150 ply from sediments (Fig. 2) exceed those from atmospheric dust deposition throughout the model
 151 domain.

152 We also added an anoxic nitrogen cycle, with losses to the sediments and water column. Bot-
 153 tom water nitrate is removed using a statistical description of sediment denitrification proposed by
 154 Middelburg et al. (1996), based on a vertically resolved diagenetic model that predicts the primary
 155 dependence of benthic denitrification to be on organic carbon sedimentation rate, with a secondary
 156 sensitivity to bottom water oxygen concentration. This statistical description of the complete di-
 157 agenetic model reproduces basic controls on observed sediment fluxes, without the considerable
 158 computational cost of a sedimentary submodel. The predicted rates of NO_3^- loss from this sedi-
 159 mentary sink (Fig. 2) amount to a small loss of $\approx 3 \times 10^{12}$ gN yr^{-1} . Denitrification in the water
 160 column is also modeled, but its integrated removal rate is an order of magnitude smaller than sedi-
 161 mentary losses, and has negligible impact on the results because O_2 in the model domain rarely
 162 falls below the threshold (5 mmol m^{-3}) assumed for this process. The higher O_2 thresholds as-
 163 sociated with anaerobic particle micro-environments could increase the importance of anaerobic
 164 processes in the CCS, but they are not represented in this model (Bianchi et al., 2018). The removal
 165 of NO_3^- by denitrification also acts as a sink for alkalinity.

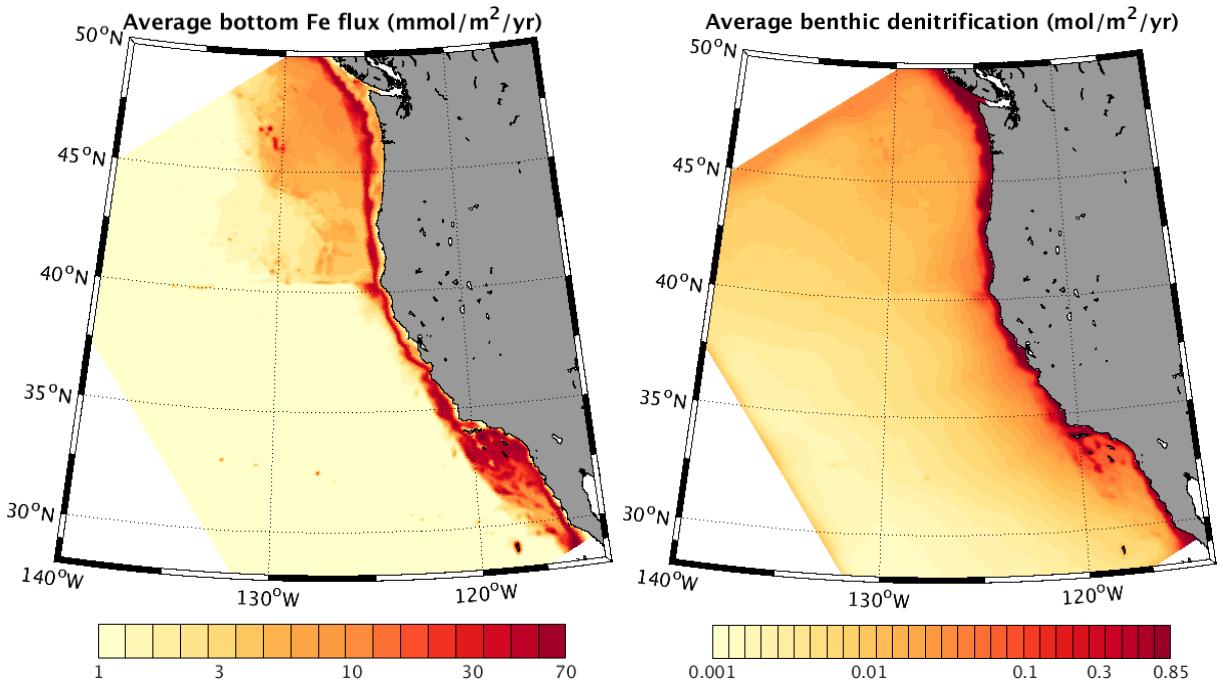


Figure 2: Parameterized fluxes of iron and nitrate between the water column and sediments. (a) The Fe efflux (in $\text{mmol m}^{-2} \text{ yr}^{-1}$) to the water column from the sediments. (b) The NO_3^- flux from the water column to the sediments due to net denitrification in sedimentary pore waters. Both fluxes are parameterized as a function of bottom water O_2 , and denitrification is additionally parameterized as a function of organic matter flux to the seafloor. These maps are therefore part of the model solutions, and not prescribed forcings.

2.2 Biogeochemical forcing and validation data

The biogeochemical model components and the physical (Renault et al., 2021) and biogeochemical forcings are schematically represented in Fig. 1. The biogeochemical boundary conditions for nutrients (NO_3^- , PO_4^{3-} , $Si(OH)_4$) and O_2 are taken from monthly climatological observations in the 2013 World Ocean Atlas (WOA) (Garcia-Reyes et al., 2014). Additionally, lateral NO_3^- fluxes derived from prior model simulations (Davis et al., 2014) of nutrient exchange between the Strait of Juan de Fuca and the coastal ocean were imposed as landward boundary conditions in the Northern CCS. Non-nitrogenous nutrients were not available for inclusion in fluxes from the Strait of Juan de Fuca. However, we imposed an Fe concentration at this boundary by scaling it to nitrate. The scaling factor ($Fe=NO_3^-*3e-5$) was chosen to be similar to that of the surrounding coastal water. This ensures that Juan de Fuca nutrient inputs will not alter the locally limiting nutrient, absent data to support such an alteration. Boundary condition data for Fe is taken from global simulations with the Community Earth System Model (CESM) that used an earlier version of the same BEC ecosystem model. The NH_4^+ boundary concentrations, being small in nature, are set to zero, but adjust rapidly to the ecosystem processes in the interior of the domain. Time-dependent carbon cycle parameters, DIC and Alk, are taken from GLODAP (Key et al., 2004), with a reference year of 1995. An imposed trend at the boundary scales the anthropogenic component of DIC in proportion to the rise of atmospheric CO_2 since 1995. Time-dependent atmospheric pCO_2 is also used as a surface boundary condition for air-sea gas exchange. Aside from the boundary carbonate system parameters, the only non-stationary forcing of the model solution comes from the physical boundary conditions and surface forcing (Renault et al., 2021).

In order to ensure the integrity of tracer relationships along isopycnal surfaces, we map the biogeochemical boundary conditions from source data to the model grid using density rather than depth as the vertical coordinate, while retaining the mean-seasonal values of T and S (hence density) along the boundary as specified in the physical conditions (Renault et al., 2021). This prevents any errors in the depth of isopycnal surfaces inherited from the physical boundary data (Mercator) from biasing the biogeochemical properties along that surface. Concurrently, this results in biogeochemical boundary conditions responding to interannual variability in isopycnal depth, despite being climatologically fixed along isopycnal surfaces at the boundary. Thus interannual biogeochemical variability is propagated into the model domain because of the covariance between property isopleths and isopycnal displacements rather than changing biogeochemical water mass properties on density surfaces. However, biogeochemical variability along isopycnal surfaces in the interior domain can still arise from varying the proportions (mixture) of water masses entering from different boundaries, or from time variable rates of biogeochemical transformation on isopycnals in the model interior.

The CCS is among the best-sampled regions of the world's oceans. Hydrographic sampling and biological rate measurements have been conducted repeatedly if not routinely along several sections off the West coast, most notably in the CalCOFI sampling area in the Southern California Bight, off Monterey Bay, California, off Newport, Oregon, and Line P off Victoria, British Columbia at the northern edge of the 4 km model domain. Despite the abundant datasets from this region, data density is still sparse for much of the central California coast and for many biogeochemical properties of interest (e.g., Fe). The total number of profiles in the 2013 World Ocean Database (WOD; downloaded from <https://www.nodc.noaa.gov/OC5/WOD13/> and including the hydrographic line data) are plotted for NO_3^- and O_2 over the entire historical data period (1955-

210 2013; Fig. 3).

211 To facilitate comparison of model outputs to data, and particularly seasonal cycles, we de-
212 fined 6 regions (Fig. 3), dividing the CCS by distance from the coast into nearshore (0-100 km)
213 and offshore (100-500 km) regions, and by latitude into the Northern, Central and Southern CCS.
214 These designations are somewhat arbitrary, but are based on a combination of topographic delin-
215 eations and to ensure adequate data coverage in each region. We focus our validation efforts on
216 broad-scale measures that can be evaluated from climatological databases, namely WOD and its
217 objectively-mapped climatological representation, the World Ocean Atlas (WOA; Garcia-Reyes
218 et al. (2014)). We further evaluate the vertical and cross-shelf structure of biogeochemical vari-
219 ables at greater resolution at the hydrographic line locations. Higher frequency biogeochemical
220 measurements from moorings are generally available only for more recent periods, and primarily
221 from nearshore environments. Model comparison to mooring data is left for planned downscaling
222 of these simulations better suited to examining high-frequency variability.

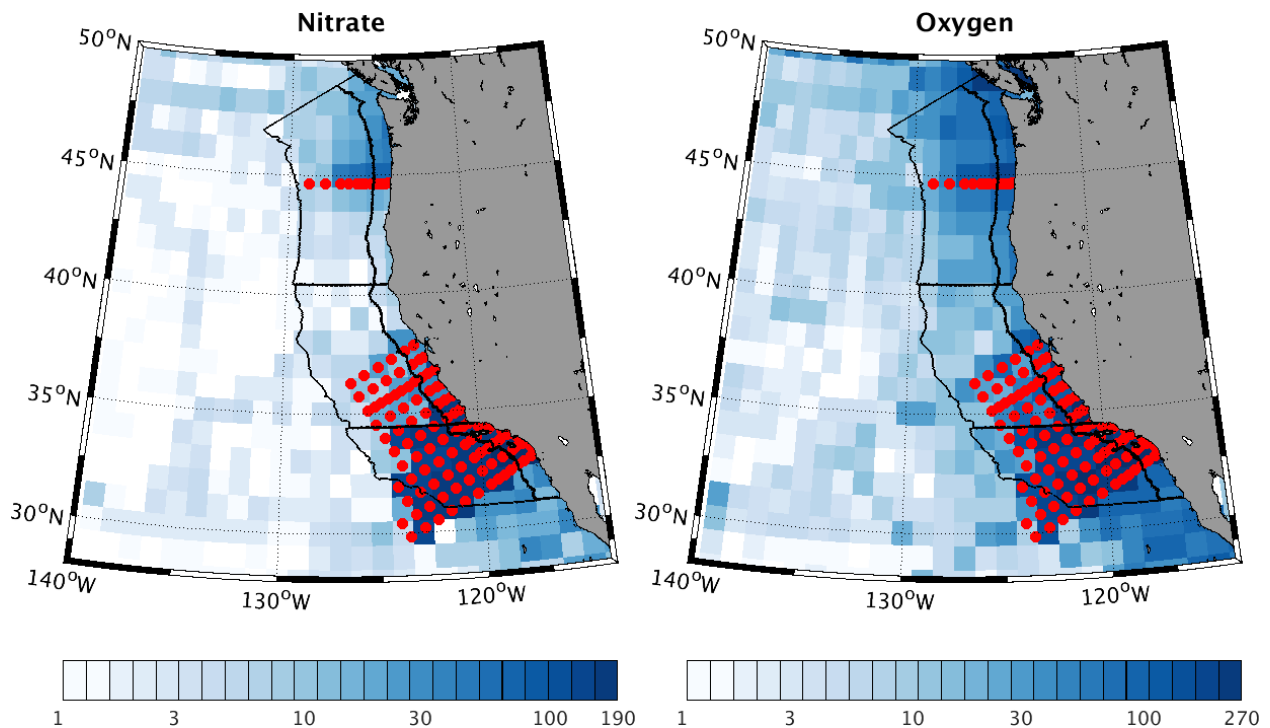


Figure 3: Hydrographic data density used for model validation. Observations in the World Ocean Database are binned in a regular 1° latitude/longitude grid for each month over the entire historical data period (1955-2013). Total number of months with a profile are shown for NO_3^- (left) and O_2 (right). Nominal station locations for major repeat hydrographic lines used for validation (see Fig. 16) are shown (red circles), along with the boundaries (black lines) used for regional time series comparisons in Figs. 6, 7, 14, and 21.

223 3 Results

224 We describe the spatial patterns and temporal variability of model biogeochemical solutions and
225 their fidelity to observational datasets. Of the numerous properties and rates of the biogeochemical
226 system that are predicted by the model, we focus on those that are most important to ecosystem
227 primary productivity and the overall elemental cycling of carbon, nitrogen, and oxygen, and are
228 best observed over scales captured by the model. This analysis begins with the photic zone, with
229 particular emphasis on the factors driving NPP variability at multiple scales of space and time.
230 We also evaluate the export of this productivity to depth. Second, we present results from the
231 ocean thermocline, where the respiration of exported surface productivity contributes to hypoxic
232 and corrosive conditions. Aspects of the model solution that are not presented include nutrients
233 that are not limiting, rates that lack large-scale and climatological datasets, and variability that is
234 poorly resolved by a $dx = 4$ km model (*e.g.*, submesoscale and nearshore phenomena).

235 3.1 Photic Zone

236 3.1.1 Chlorophyll and Net Primary Production

237 We begin with an evaluation of model distributions of Chlorophyll-a (hereafter *Chl*), and Net
238 Primary Productivity (NPP), both of which can be estimated from remote sensing of ocean color.
239 While NPP is of greater biological significance, its estimation is less direct than for *Chl*.

240 The model *Chl* concentrations are governed by the product of biomass and the $C : Chl$ ratios.
241 Biomass is subject to advection and to ecosystem transformations (see Appendix). The $C : Chl$
242 ratio is determined by photoacclimation, or the amounts of light-harvesting pigments and photo-
243 protective compounds produced by phytoplankton in response to their growth environment. This
244 process is included in the ROMS-BEC representation of phytoplankton physiology following the
245 model of Geider et al. (1998), which relates changes in chlorophyll synthesis and nutrient uptake
246 in response to changing PAR. The dominant patterns of *Chl* are also found in biomass (see be-
247 low), indicating that photoacclimation is not the leading factor in *Chl* variability. While biomass
248 may be a more ecologically meaningful comparison, we validated model solutions using *Chl* be-
249 cause it is more directly estimated from ocean color sensing. The frequency distribution of *Chl* in
250 both ROMS and SeaWiFS remote sensing data is approximately log-normal and is mapped after
251 logarithmic transformation.

252 The annual mean concentrations of *Chl* vary most strongly in the cross-shore direction, with
253 relatively weak alongshore gradients, a well-known pattern in observations (Banas and Hickey,
254 2008) that is well represented by the model (Fig. 4a,b). The offshore drop-off in *Chl* is somewhat
255 weaker in model simulations, resulting in a wider band of high coastal *Chl*, a tendency that is
256 not reflected in NPP (discussed immediately below). Kessouri et al. (2020) shows that there is
257 some sensitivity in these distributions to model resolution, with higher resolution increasing the
258 nearshore biomass and productivity. The leading pattern of variability in climatological *Chl* is
259 characterized by a seasonal cycle that is also largely synchronous along the coast.

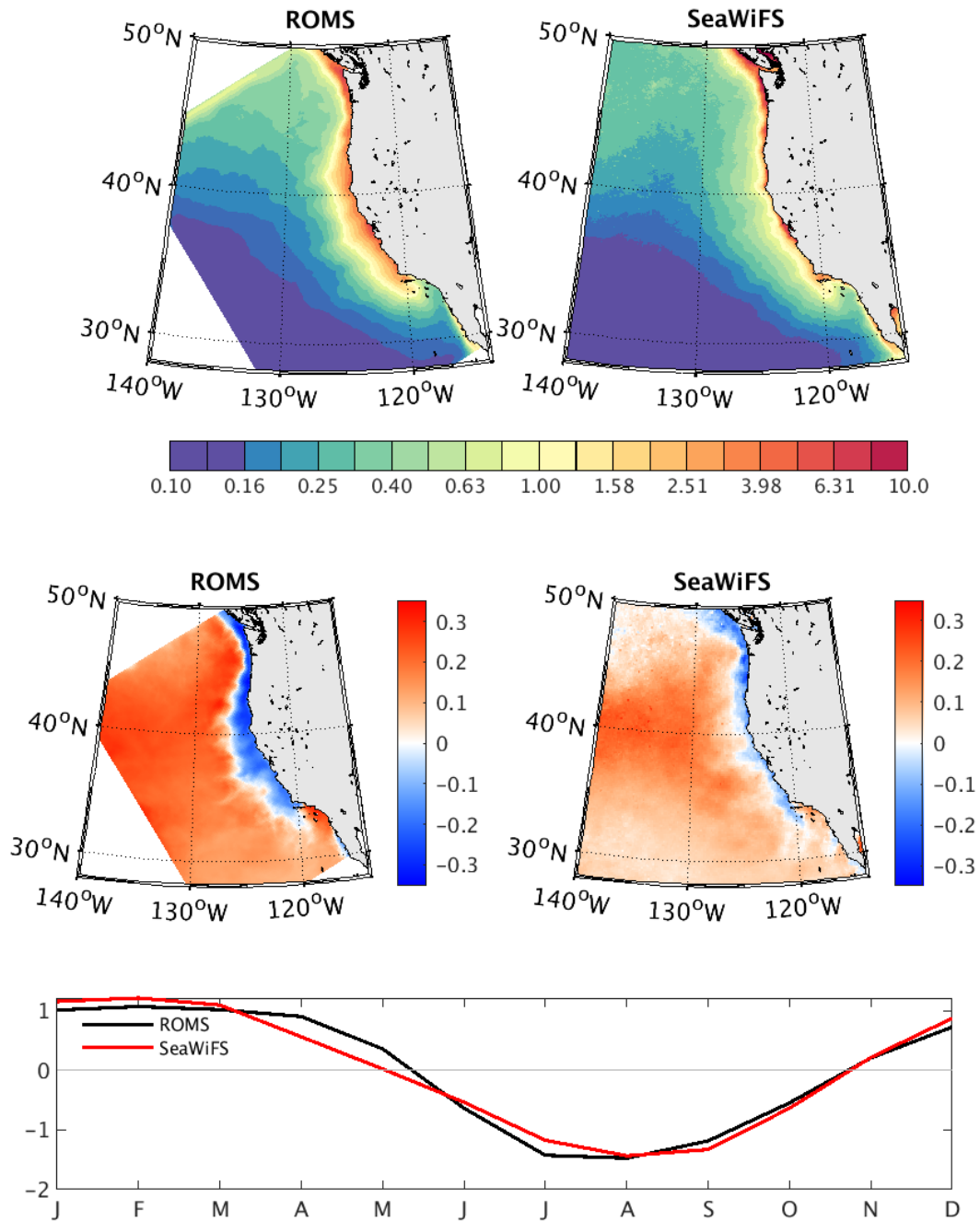


Figure 4: Mean annual Chlorophyll (Chl) and its seasonal cycle. Mean annual concentrations (upper panels) are shown in mg m^{-3} averaged over the simulation period in both model output and SeaWiFS remote sensing level 3 product. The seasonal cycle is shown as the spatial loading patterns (middle panels) and time series (bottom panel) of the first EOF of climatological values of $\log_{10}[Chl]$. In both the mean annual and seasonal variations of Chl , the dominant variations are cross-shore and at Point Conception, which separates the central CCS from the SCB. High coastal Chl extends further offshore in model solutions than in observations, a bias that is not found in productivity (Fig. 5).

260 The leading Empirical Orthogonal Function (EOF) (Fig. 4c,d,e), has *Chl* reaching peak values
 261 in late summer, and it accounts for the large majority of the climatological variance in both ob-
 262 servations (EOF1=64% variance) and model solutions (EOF1 = 69% variance). The second EOF
 263 (not shown) also has a similar loading pattern, but with more meridional structure offshore and a
 264 minimum in spring—this accounts for only $\approx 13\%$ and 19% of the variance in observations and the
 265 model, respectively. The seasonality of *Chl* reveals anti-phased cycles between nearshore high-
 266 *Chl* band and the lower *Chl* offshore. While near-shore *Chl* peaks in late summer, the offshore
 267 surface *Chl* has a minimum. This pattern is not found in the EOFs of depth-integrated NPP, indi-
 268 cating that the dipole structure of the *Chl* pattern results from a vertical redistribution of *Chl* to
 269 greater depths in offshore waters as they become more oligotrophic due to nutrient uptake during
 270 summer months. This interpretation is confirmed in the analysis of the vertical *Chl* maximum
 271 (Sec. 3.1.2).

272 In BEC, NPP depends on the sum of j model phytoplankton biomasses (B_j), their maximum
 273 growth rates ($\mu_{max}(T) = \mu_0 T^{1.06}$), and the limitation of those rates by light ($0 \leq \gamma_j(I) \leq 1$)
 274 and the minimum Michaelis-Menten function $\lambda_{i,j}$ ($0 \leq \lambda_{i,j}(N_i) \leq 1$) among the i nutrients with
 275 half-saturation $K_{i,j}$ ($0 \leq \lambda_{i,j}(N_i) \leq 1$), written as:

$$NPP = \sum_{i,j} \mu_{max j}(T) \gamma_j(I) \min[\lambda_{i,j}(N_i)] B_j. \quad (1)$$

276 The spatial patterns of modeled NPP fall within the range of satellite-derived estimates (Fig. 5).
 277 The two commonly used satellite-based algorithms are the vertically generalized production model
 278 (VGPM) (Behrenfeld and Falkowski, 1997) and the carbon-based productivity model (CbPM)
 279 (Westberry et al., 2008). The VGPM estimates productivity on the basis of light and chlorophyll
 280 concentrations, calibrated to a predominantly coastal radiocarbon incubation dataset. The CbPM
 281 additionally incorporates phytoplankton backscattering and growth rate relationships in order to es-
 282 timate productivity as a function of phytoplankton biomass, and it is calibrated to subtropical gyre
 283 radiocarbon incubations. The two algorithms exhibit a relatively wide range for the CCS region,
 284 reflecting the considerable uncertainty in “empirical” NPP estimates as well as differences in the
 285 measurements underlying each. The VGPM algorithm has a larger offshore gradient, with higher
 286 coastal values, and lower values in the open ocean, compared to the carbon-based CbPM. NPP
 287 rates from ROMS-BEC fall between the two remote sensing products, but are generally closer
 288 to the values of the VGPM algorithm, supporting higher near-shore rates, lower offshore rates,
 289 and increased seasonality relative to the CbPM. The VGPM has been explicitly calibrated against
 290 radiocarbon bottle incubations from the CalCOFI program, and it is therefore likely to be more ac-
 291 curate in this region (Kahru et al., 2009). Indeed, we find that ROMS-BEC rates and distributions
 292 of productivity are also consistent with direct estimates from ship-based data both from CalCOFI
 293 (Fig. 5; Munro et al. (2013)) and the broader subtropical Northeast Pacific (Palevsky et al., 2016).

294 Productivity in the northern CCS has been consistently biased in regional models (Banas and
 295 Hickey, 2008), including in our initial simulations. We conducted simulations with and without
 296 lateral nitrogen fluxes at the Strait of Juan de Fuca imposed as boundary condition from model
 297 simulations by Davis et al. (2014). Consistent with that study, without the nutrient inputs from
 298 the Salish Sea, NPP was biased low by $> 50\%$. The inclusion of the effects of nitrogen inputs at
 299 the Strait of Juan de Fuca brought the model much closer to satellite-based empirical models. The
 300 inclusion of these inputs is consistent with the study by Davis et al. (2014), and it is used in all
 301 results reported here.

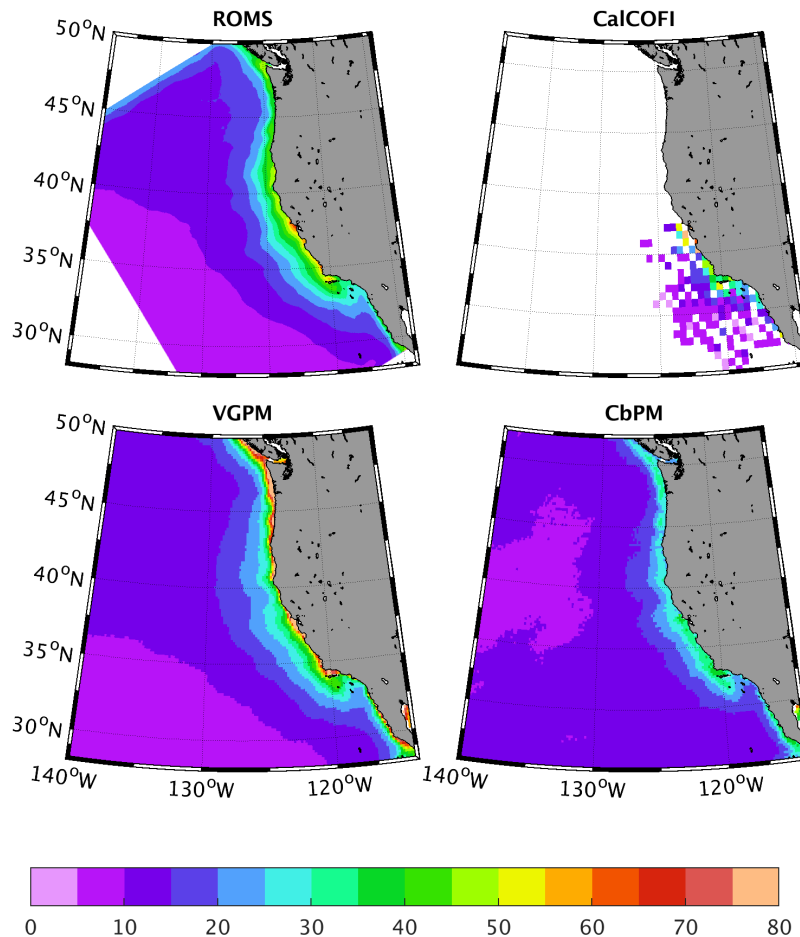


Figure 5: Spatial distribution of annual Net Primary Productivity ($\text{mol } C \text{ m}^{-2} \text{ yr}^{-1}$) integrated over the depth of the photic zone, from (top left) ROMS, (top right) CalCOFI, (bottom left) VGPM, and (bottom right) CbPM.

302 The seasonal cycle of NPP is also well captured by the model (Fig. 6). In all 6 regions of
 303 the CCS, the climatological NPP, integrated over the depth of the photic zone and averaged over
 304 the regional mask, exhibits an amplitude and phasing that is within the range of satellite-based
 305 empirical models. The most notable exception is the offshore Northern domain, where a spring
 306 bloom is predicted to be of stronger magnitude than estimated by either satellite product. This
 307 model result is consistent with measured geochemical tracers, which also indicate the spring bloom
 308 in the offshore Northeast Pacific is greater than estimated from the satellite algorithms (Palevsky
 309 et al., 2016). A smaller discrepancy occurs in the southern nearshore region, where ROMS-BEC
 310 generates greater summer production than either of the satellite algorithms. Model NPP in the
 311 oligotrophic part of the domain is lower than satellite estimates, but is more consistent with the
 312 most offshore values in the depth-integrated rates based on radiocarbon bottle incubations from
 313 CalCOFI. Overall, where and when ROMS-BEC and satellite algorithms for NPP disagree, ROMS-
 314 BEC output is generally closer to the available observational data.

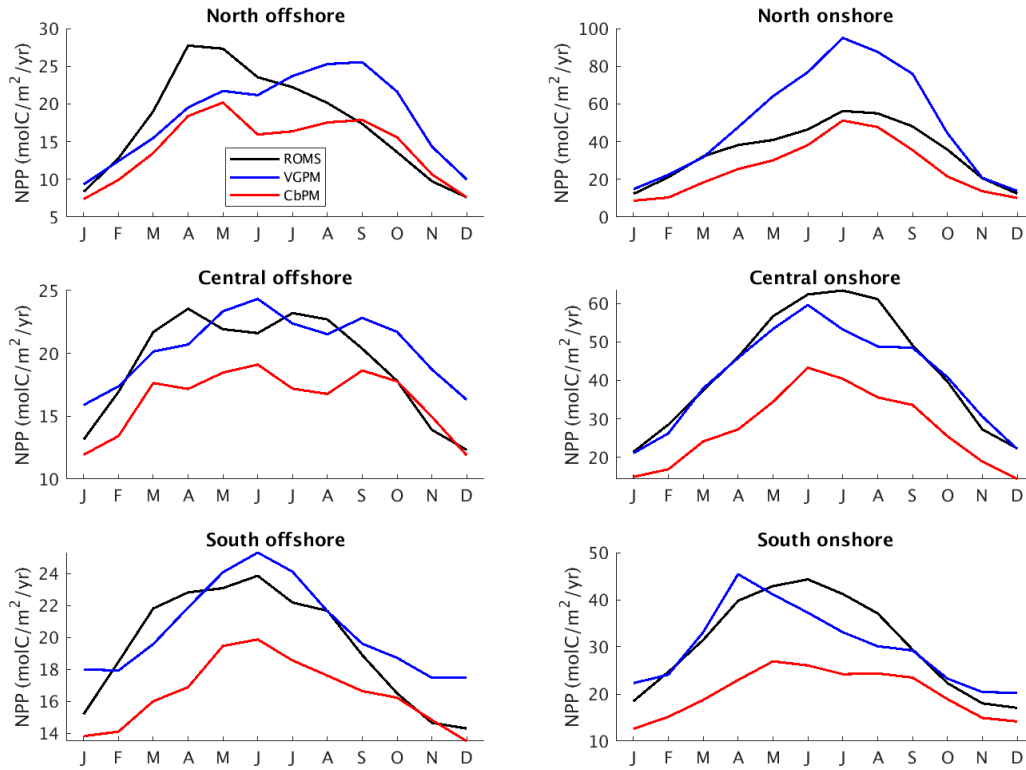


Figure 6: Seasonal cycle of annual Net Primary Productivity. The NPP rate ($\text{mol } C \text{ m}^{-2} \text{ yr}^{-1}$) from ROMS-BEC (black), and two satellite algorithms (VGPM, blue; CbPM, red) are integrated over the depth of the photic zone, and averaged over 6 regions (see Fig. 3) from northern (top row), central (middle row), and southern (bottom row) of the CCS, and divided by distance from the coast into a nearshore (0-100 km; right column) and offshore (100-500 km; left column) region.

315 3.1.2 Seasonal limitation of productivity by light and nutrients

316 To evaluate the role of environmental factors shaping the seasonal cycle and regional differences
 317 in rates of productivity, we computed monthly mean limitation factors for each of the environmen-
 318 tal variables that modulate the maximum growth rates, including macronutrients (NO_3^- , PO_4^{3-} ,
 319 $Si(OH)_4$), Fe , and light; see (1). By construction, growth rates are limited by only one nutri-
 320 ent at a time (Liebig’s Law of the Minimum), such that only the lowest value has an influence
 321 on rates. Light operates as a multiplicative factor on nutrient limitation, reducing growth relative
 322 to the light-saturated photosynthesis rate (Geider et al., 1998). Temperature influences maximum
 323 growth rate, but is not considered a limiting factor in the upper ocean, so is not analyzed here.

324 Over a climatological seasonal cycle, small plankton growth rates are almost always reduced
 325 by light more so than by nutrients, regardless of season or location (Fig. 7). The small plankton
 326 are assumed to have a lower half-saturation constant for nutrients, and the resulting higher affinity
 327 makes them less prone to nutrient limitation than large plankton are. Thus, in the inshore regions
 328 where nutrients are high, light is always limiting. Offshore, nutrients can limit small phytoplank-

329 ton to a similar degree as light in the summer. However, large phytoplankton make up $\approx 90\%$ of
 330 modeled NPP on average, and large phytoplankton (“diatoms”) are primarily limited by nutrient
 331 availability. In the model, this results from a nutrient supply by combined wind-induced upwelling
 332 and eddy-induced subduction (Gruber et al., 2011; Nagai et al., 2015; Renault et al., 2016a) that is
 333 unable to saturate the potential net uptake of nitrogen by phytoplankton at prevailing light levels.
 334 Significant eddy fluxes also occur due to submesoscale eddies and fronts in the CCS (Kessouri
 335 et al., 2020), but are not resolved in this model. Large phytoplankton only experience light limita-
 336 tion in the northern CCS, where the seasonal cycle alternates between winter light limitation and
 337 nutrient limitation for the rest of the year.

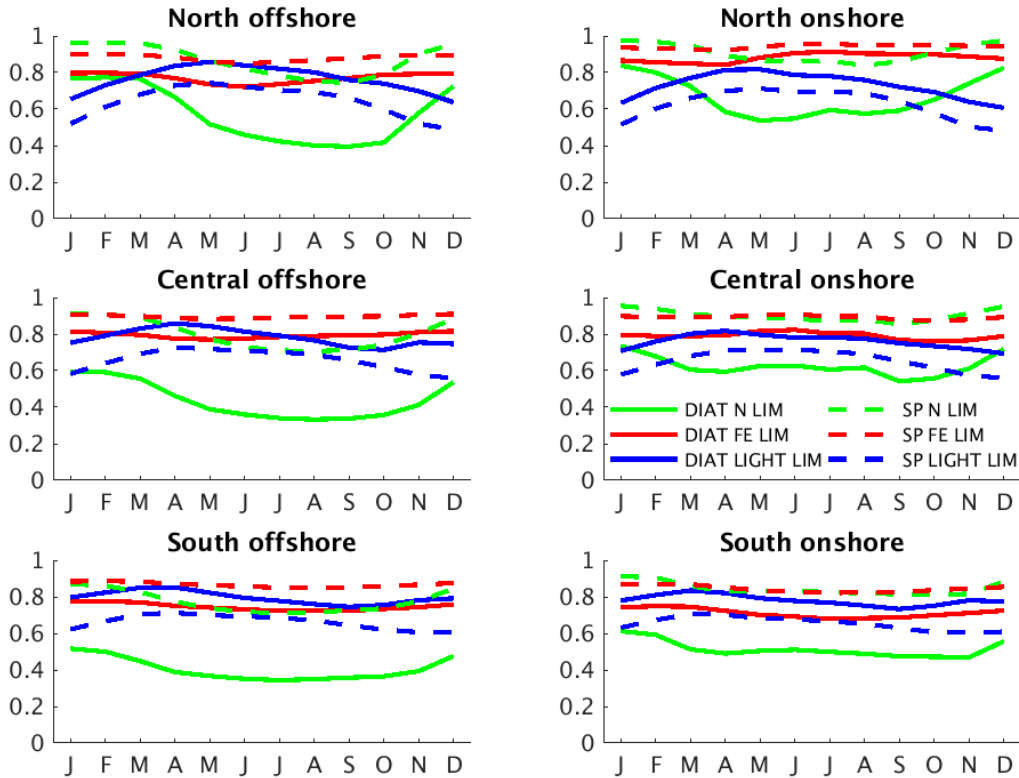


Figure 7: Seasonal cycle of growth limitation factors for light (γ term in (1), blue) and for nutrients (λ term in (1): NO_3^- (green) and Fe (red) for diatoms (solid) and small phytoplankton (dashed)). Factors are NPP-weighted and averaged over the photic zone for each region shown in Fig. 3. Limitation factors close to 1 mean no limitation; values close to 0 mean complete limitation.

338 While on the regional scales used for this analysis nitrogen limitation appears more stringent
 339 than Fe limitation throughout the CCS, significant Fe limitation occurs on smaller scales and
 340 shorter durations (see below). The seasonal amplitude of plankton growth rates is relatively small
 341 ($\approx 10\%$), indicating that the amplitude of seasonal production ($\approx 100\%$) is governed by seasonal
 342 controls on biomass rather than growth rates. The trade-off between light and nutrient limitation
 343 spatially and seasonally is a ubiquitous feature of phytoplankton distributions and phenology. Mov-
 344 ing deeper in the water column, light becomes more limiting as photosynthetically active radiation

345 (PAR, light of 400 - 700 nm wavelengths) is absorbed and scattered, while nutrient concentra-
346 tions are greatest below the surface mixed layer and rapidly decrease upward across the seasonally
347 variable pycnocline; *i.e.*, there is a well-defined nutricline.

348 The competing influences of nutrient and light limitation on depth of optimal plankton growth
349 (*i.e.*, highest growth rates) are reflected in the depth of the deep chlorophyll maximum (DCM).
350 In the CCS, the depth of the DCM deepens from the coast to the open ocean, suggesting that the
351 growth-maximizing combination of light and nutrients is found deeper offshore, consistent with
352 a deepening nutricline that intensifies nutrient limitation at the surface and light limitation where
353 nutrients are abundant, for both small and large phytoplankton. We use the observed pattern of the
354 DCM depth as an indicator of whether the model achieves a realistic trade-off between these two
355 countervailing growth condition gradients (Fig. 8).

356 However, for DCM depth to be a reliable indicator of phytoplankton growth rate trade-offs,
357 we must rule out two alternative interpretations relating to isopycnal advection and photoacclima-
358 tion. First, the offshore deepening of the DCM can have a physical origin. Because it closely
359 follows the plunging of isopycnal surfaces offshore, the vertical peak in biomass and associated
360 chlorophyll could be caused by eddy subduction carrying high surface chlorophyll away from the
361 coast along deepening isopycnals (Gruber et al., 2011; Nagai et al., 2015; Renault et al., 2016a).
362 To evaluate this possibility, we compared the depths of maximum chlorophyll concentration and
363 phytoplankton biomass to the depth at which the product of light and nutrient limitation factors
364 are maximized (Fig. 8). These maps are virtually indistinguishable, suggesting the DCM follows
365 growth rates rather than advection. As a more stringent test, we performed a short (5-year) simu-
366 lation in which surface PAR was reduced by 10%. The results revealed a significant shoaling of
367 both the biomass and *Chl* peaks, but no detectable change in isopycnal depths, confirming that
368 these depths do in fact reflect a nutrient-light trade-off rather than advection along density sur-
369 faces. Second, the peak depth of *Chl* may also be decoupled from that of biomass and growth
370 rates due to photoacclimation, or a shift in the amounts of light-harvesting pigments and photopro-
371 tective compounds produced by phytoplankton in response to the light-environment. This process
372 is included in the ROMS-BEC representation of phytoplankton physiology following the model
373 of Geider et al. (1998), which relates changes in chlorophyll synthesis and nutrient uptake in re-
374 sponse to changing PAR. Indeed, we find that the depth of the DCM is slightly deeper than that of
375 the maximum plankton biomass. However, the offshore and latitude gradients of the depth of peak
376 biomass and chlorophyll are very similar. DCM deepening offshore is consistent with optimized
377 growth conditions in the model, and reproduces the pattern observed in the available CalCOFI data
378 (Fig. 8).

379 In summary, the *Chl* maximum does not primarily reflect photoacclimation or isopycnal trans-
380 port, although it is affected by those processes. Instead, it is found at approximately the same depth
381 as that of peak biomass (panel c), which in turn is found at the depth that maximizes growth rate
382 through the combined impacts of nutrients and light (panel d). This is the same depth at which NPP,
383 the product of biomass and growth rate, is also maximum. Thus, these correspondences indicate
384 that the model DCM is a reflection of the essential trade-off between light and nutrient limitation,
385 and the fidelity to the observed DCM implies this trade-off is adequately captured.

386 Growth rates are modulated by a complex and evolving pattern of nutrient limitation by reactive
387 nitrogen ($NO_3^- + NH_4^+$) and soluble dissolved *Fe*, with no appreciable limitation by $Si(OH)_4$ and
388 PO_4^{3-} in the CCS. The limitation factors are mapped as a biomass-weighted fraction of time that
389 each of the nutrients is most limiting (Fig. 9). The spatial pattern among nutrients largely reflects

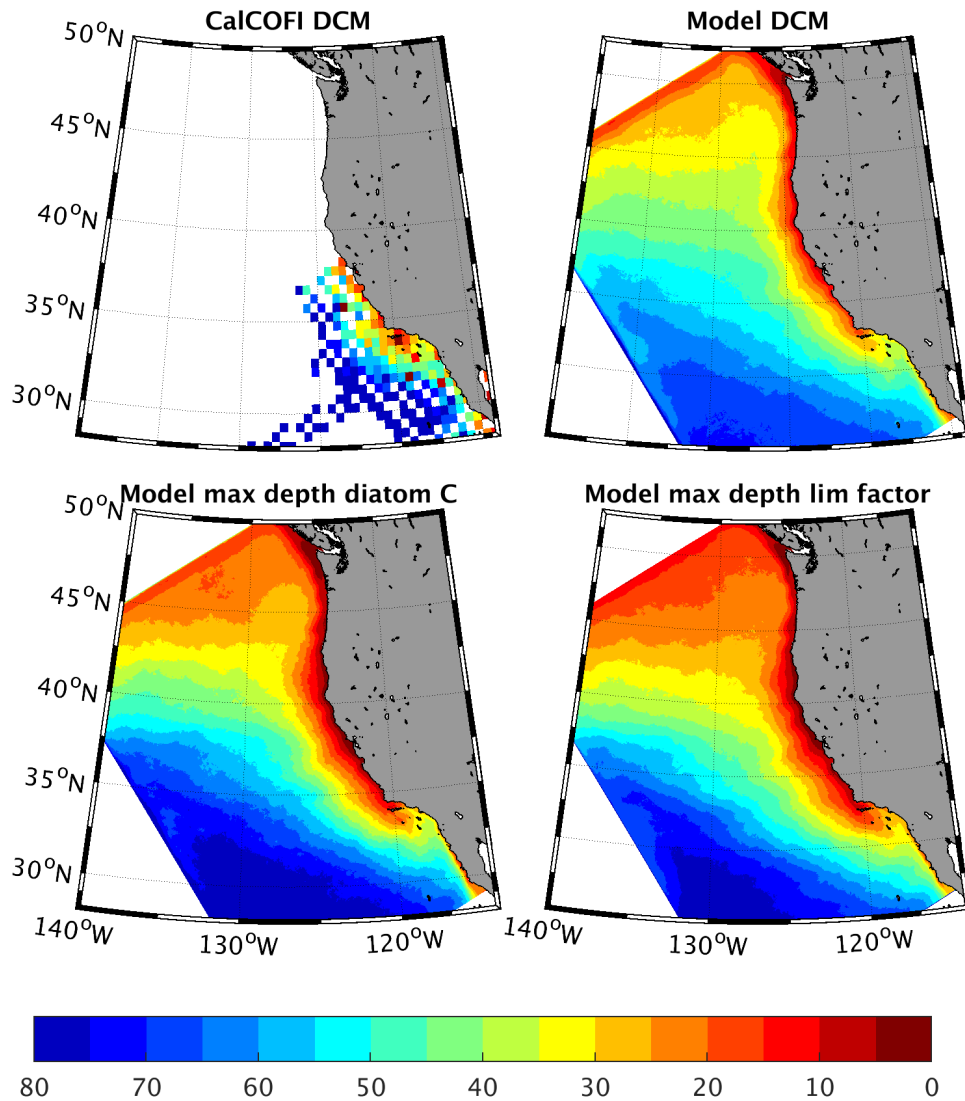


Figure 8: Depth of the vertical maximum of Chlorophyll in CalCOFI (top left) and ROMS (top right), and of model diatom biomass (bottom left) and nutrient limitation factor (bottom right). The correspondence between model fields demonstrates that the model DCM tracks nutrient limitation, while the fidelity to observations implies the model captures a realistic trade-off between nutrient and light limitation.

390 the areas where NO_3^- supply routinely exceeds maximum potential uptake seasonally. Thus, the
 391 waters entering the CCS from the subarctic High Nutrient – Low Chlorophyll (HNLC) region are
 392 most frequently *Fe* limited. Along the coast, the seasonal upwelling of excess nitrogen leads to
 393 significant periods of *Fe* limitation as well. In the coastal zone off Monterey Bay, *Fe* limitation
 394 has been diagnosed via incubation experiments, in a band of water slightly offshore, with nitrogen
 395 limitation both in more shoreward and open coastal zones (Firme et al., 2003). This pattern is
 396 consistent with that predicted by the model (Fig. 9b inset). Most of the rest of the model domain
 397 is perpetually nitrogen limited.

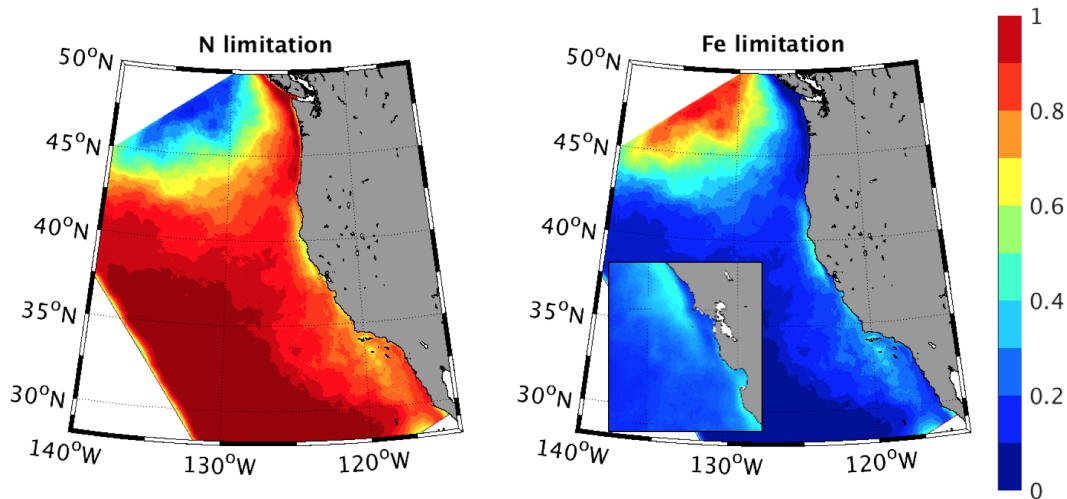


Figure 9: Frequency of limitation by NO_3^- (left) and Fe (right) for the model’s dominant primary producer, diatoms. The limitation factors are weighted by biomass (as in Fig. 7), using 5-day output, and plotted as the fraction of time that each nutrient limitation factor is the lowest among nutrients. The inset shows offshore band of relatively frequent Fe limitation along the central CCS, similar to that observed by (Firme et al., 2003).

398 3.1.3 Nutrient concentrations

399 Surface nutrient concentrations provide an important measurable test of system behavior. For
 400 nutrients that limit phytoplankton growth, accurately simulating their distributions is a necessary
 401 condition for a mechanistic prediction of NPP. Moreover, they provide an integrative measure of
 402 net community production (equal to NPP less community heterotrophic respiration) and export of
 403 organic matter to the thermocline. We therefore compared model predicted distributions of the two
 404 primary limiting nutrients, NO_3^- and Fe , to available observations.

405 Coastal measurements of dissolved Fe reveal a spatially patchy distribution, reflecting its short
 406 residence time with respect to removal by plankton uptake and particle-active scavenging. Existing
 407 data are too sparse to yield a clear climatological pattern for model validation. However, the
 408 primary coastal region where the model predicts most frequent Fe limitation, in the central CCS,
 409 has been relatively well sampled, including on two cruises off Monterey Bay that also tested for
 410 Fe limitation (Firme et al. (2003); see Fig. 9, inset). Given the lack of a clear large-scale pattern
 411 of surface Fe levels, we used a more statistically-based validation metric, focusing on the relative
 412 frequency of Fe measurements versus concentration and distance from the shore (Fig. 10). On
 413 average, the data and model both show a decline in the mean and median Fe levels with offshore
 414 distance. This reduction is driven largely by the decreasing frequency of high concentrations, while
 415 the most commonly observed Fe levels remain consistent at $\approx 0.5 \cdot 10^{-6} \text{ mol m}^{-3}$ regardless of
 416 distance from shore. The thinning tail of high concentrations in Fe distribution with cross-shelf
 417 distance occurs in both modeled and observed fields, but is more pronounced in the measurements.

418 In the CCS, the nutrient most often limiting NPP is reactive nitrogen, of which by far the largest
 419 and most commonly measured pool is NO_3^- . We therefore compare the model simulated patterns

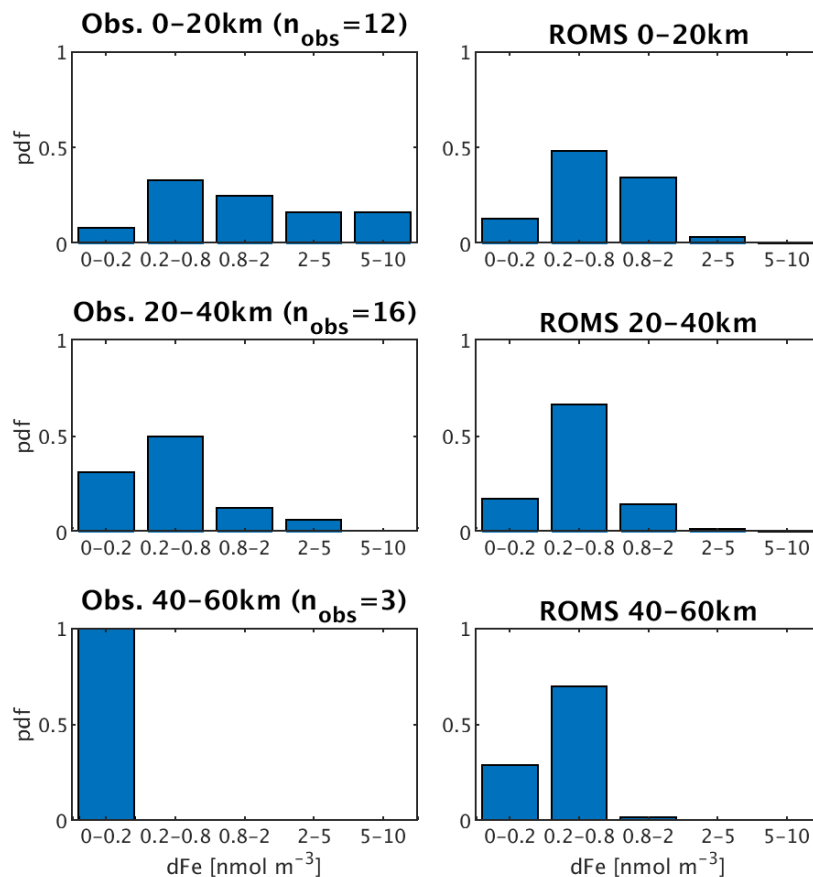


Figure 10: Histogram of surface Fe concentrations (10^{-6} mol m^{-3}) from observations (left) and ROMS (right). Because Fe concentrations are patchy in nature and sparsely observed, values are binned by distance from shore (0-20 km, upper row; 20-40 km middle row; 40-60 km lower row) to reveal a cross-shelf gradient, in the same latitude band off Monterey Bay for which observations were made on summer cruises (Firme et al., 2003).

420 of NO_3^- to climatological values from the World Ocean Database (Fig. 11). A depth of 50 m is
 421 chosen because it approximates the average depth of maximum biomass and NPP (see Fig. 8) and
 422 is generally near the base of the photic zone on the continental shelf. We included all historical
 423 measurements for this analysis because the data density in the model period (1994-2010) was much
 424 sparser, and no significant differences were found between the average NO_3^- in this period relative
 425 to 1955-2013.

426 ROMS-BEC captures regional patterns well for NO_3^- (Fig. 11). Annual mean concentrations
 427 of ≈ 15 mmol m^{-3} along the coast decline to values below the half-saturation level for model
 428 diatom growth ($NO_3^- < 1$ mmol m^{-3}) within 500 km from shore. The offshore gradient is similar
 429 throughout most of the CCS, except in the Southern California Bight (SCB), where coastal sur-
 430 face values are much lower. Similar model fidelity was found for other macronutrients ($Si(OH)_4$,
 431 PO_4^{3-}), but not shown because they do not reach limiting concentrations. The coastal zone exhibits
 432 strong variability in NO_3^- at 50 m, with standard deviations of 5-10 mmol m^{-3} throughout most

433 of the coastal zone, but with a slight northward increase in variance (Fig. 11c). The variability of
 434 NO_3^- in the climatology (WOA) exhibits a similar spatial pattern, but with $\approx 50\%$ of the magni-
 435 tude. Thus, approximately half of the variation in surface nutrients most commonly limiting NPP is
 436 associated not with the seasonal cycle, but with interannual variability. The model also reproduces
 437 observed magnitudes and patterns of NO_3^- variability (Fig. 11d). We use the interannual anomaly
 438 fields in the model and in observations to test the importance of nutrient supply as a mechanism
 439 driving changes in NPP over time.

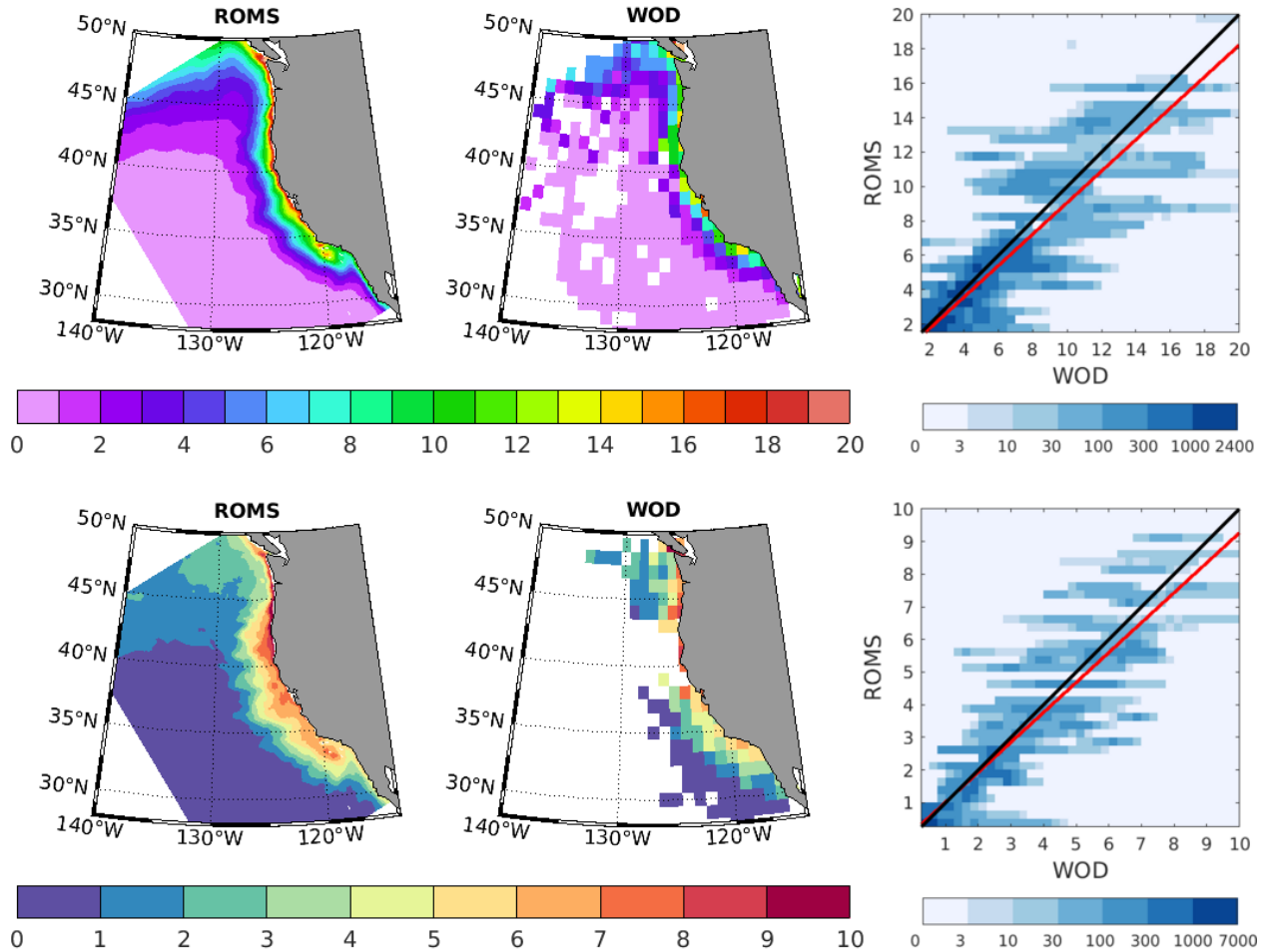


Figure 11: Long-term mean (upper row) and historical variability (bottom row) of nitrate (mmol m^{-3}) near the base of the photic zone (50 m) in ROMS (left column), WOD (center column), and their correlations (right column). Variability is mapped as the standard deviation and consists of roughly equal contributions from seasonal and interannual variability (see text). The full period of the World Ocean Database (1955-2013) is used to yield the most robust estimate of variance. The correlations between ROMS and WOD are highly significant ($p \ll 0.01$) for both mean and variability, with squared Pearson correlation coefficients (R^2) of 0.84 and 0.80, respectively.

440 **3.1.4 Interannual variability in surface nutrients and NPP**

441 Factors that limit NPP during the mean seasonal cycle may drive interannual and longer term
442 productivity changes. We examined correlations between NPP and both light and nutrient concen-
443 trations in model simulations and observations, where available. Interannual anomalies in NO_3^- in
444 ROMS-BEC are found to be well correlated ($R^2 \approx 0.5$) to the density of water at 100 m (Fig. 12).
445 This reflects the role of pycnocline heave and vertical mixing in supplying macronutrients, and
446 alleviating local nutrient limitation. Observations show a similar magnitude of correlation in the
447 southern and central CCS, but a weaker correlation to the north. This may reflect the role of nu-
448 trient supply processes that are either missing or represented only climatologically in our model,
449 and not connected to pycnocline heave. Because the weaker correlations are in the northern do-
450 main where nutrients can enter from subarctic surface waters, the climatological NO_3^- used for the
451 boundary conditions is a likely culprit. However, the presence of river NO_3^- sources or a variable
452 Juan de Fuca flux could also weaken the correlation in the observations relative to the model.

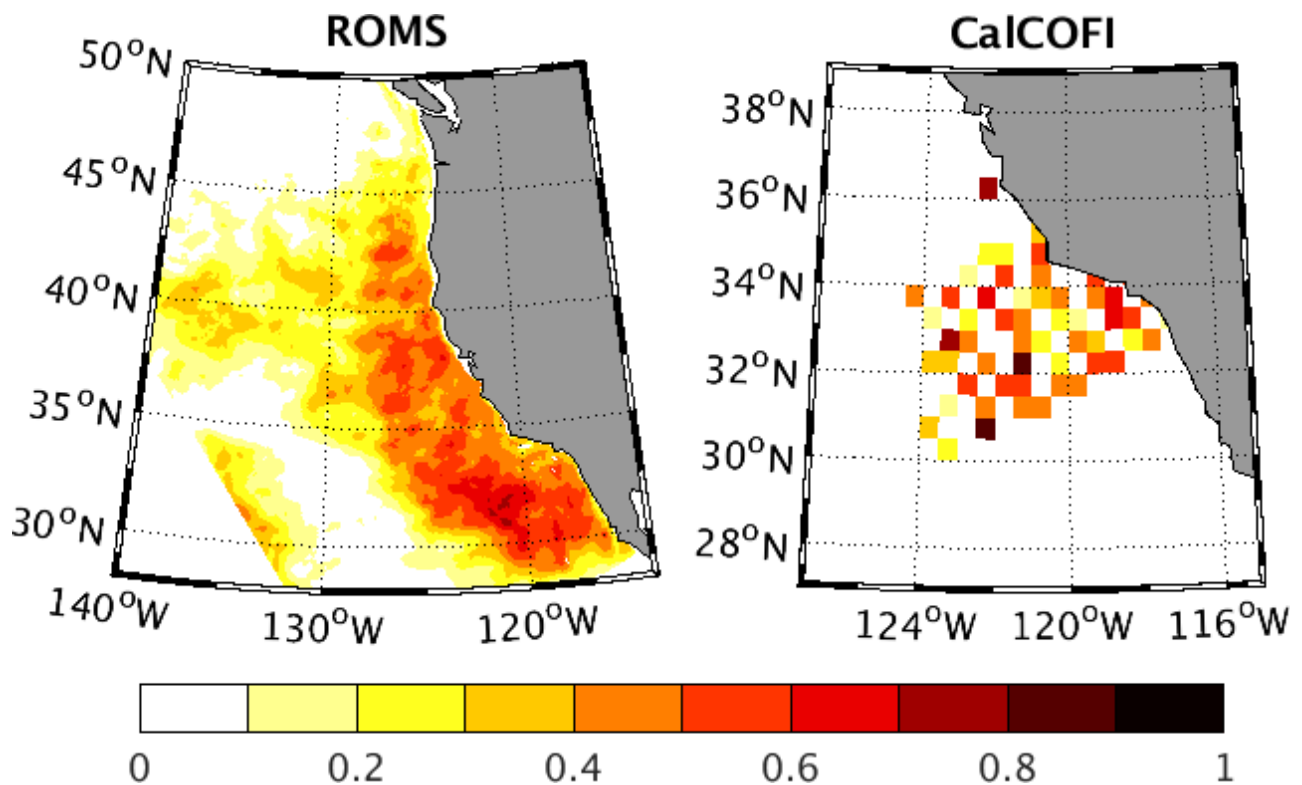


Figure 12: Correlation (R^2) between NPP and density, from ROMS-BEC (left) and CalCOFI mea-
surements. The NPP rate is integrated over depth from ^{14}C bottle incubation data, and density is
interpolated to 100 m as an index of nutrient supply (see Fig. 13). In both variables, the mean
seasonal cycle is removed leaving interannual variations. Relationships have similar strength in
data and model, and indicate that $\approx 50\%$ of interannual variability in NPP can be attributed to
anomalous nutrient supply due to pycnocline heaving.

453 Predicted correlations between NPP and density can be tested directly by combining bottle in-
454 cubations and hydrographic observations in the southern CCS (Fig. 13). Relationships between
455 nutrient and density anomalies (subtracting the mean seasonal cycle), are of similar strength, ac-

456 counting for $\approx 50\%$ of the variability in both CalCOFI observations and model simulations. Inter-
 457 annual anomalies in NPP in ROMS-BEC are also significantly correlated with surface PAR (not
 458 shown) due to variable cloudiness, though it accounts for a smaller fraction of the variance (\approx
 459 20%). The role of light is confined closer to the coastal upwelling where surface NO_3^- is high, and
 460 light availability thus limits phytoplankton growth.

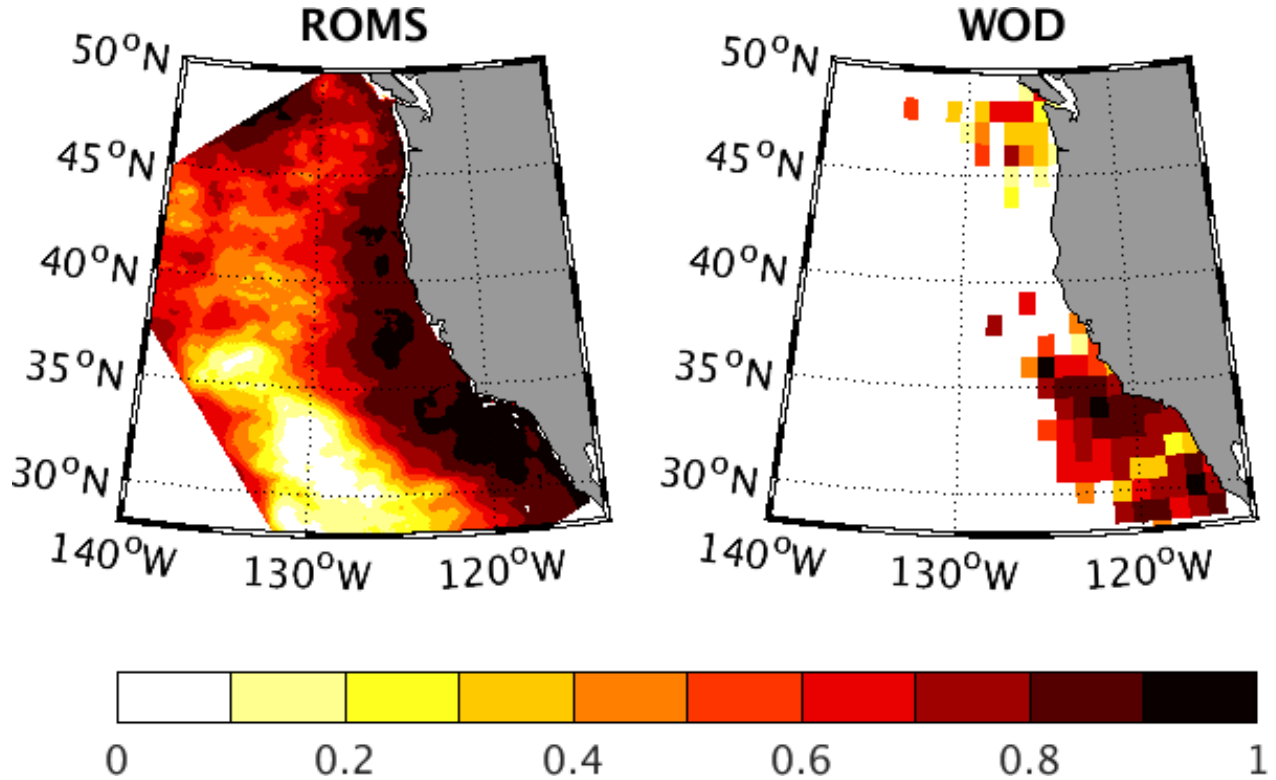


Figure 13: Correlation (R^2) between NO_3^- and density at 100 m depth in ROMS-BEC (left) and WOD (right). For both variables, the mean seasonal cycle is removed leaving interannual variations. Interannual variations in subsurface (100 m) NO_3^- highly correlated with density ($R^2 \geq 0.8$) in most of CCS. Correlation is weaker in northern CCS in data than in model.

461 3.1.5 Carbon fluxes from the photic zone

462 Of the net production of organic matter by phytoplankton, a substantial fraction can be respired
 463 by zooplankton and higher trophic levels. The remainder is available to be transported away, by
 464 particle sinking and transport of dissolved organic matter, *e.g.*, via eddy subduction (Sec. 3.1.2).
 465 The fraction of NPP that is regenerated within the surface ocean depends on food web processes,
 466 such as grazing rates. Although data is not available to evaluate large-scale patterns of grazing
 467 rates, an indirect comparison can be made through the export flux and the fraction of NPP that
 468 is exported in sinking particles rather than recycled (the so-called pe-ratio (Dunne et al., 2005;
 469 Murray et al., 1996)).

470 The fraction of NPP that is exported varies from 5-25%, consistent with the range of values
 471 inferred in field studies (Fig. 14). The model predicts highest pe-ratios in the coastal zone, where

472 productivity is high and sea surface temperature relatively low. These dependencies are also con-
 473 sistent with those inferred from a global analysis of rate measurements for NPP and net community
 474 production (NCP, assumed equal to total export of both particulate and dissolved organic carbon)
 475 (Dunne et al., 2005). For a more quantitative comparison, we compared modeled pe-ratios to those
 476 predicted by a statistical model fit to global observations by Dunne et al. (2005), over the sea-
 477 sonal cycle in each of our 6 standard CCS regions (Fig. 14). In both the empirical model and in
 478 ROMS-BEC, the mean value, phasing, and seasonal amplitude of changes in pe-ratio are similar.
 479 Empirically based estimates generate a similar result that remains consistent with model simula-
 480 tions, even when sea surface temperature is held constant. This suggests that temperature and its
 481 impacts on the relative growth rates of phytoplankton and their grazers are not the essential cause
 482 of variable pe-ratios.

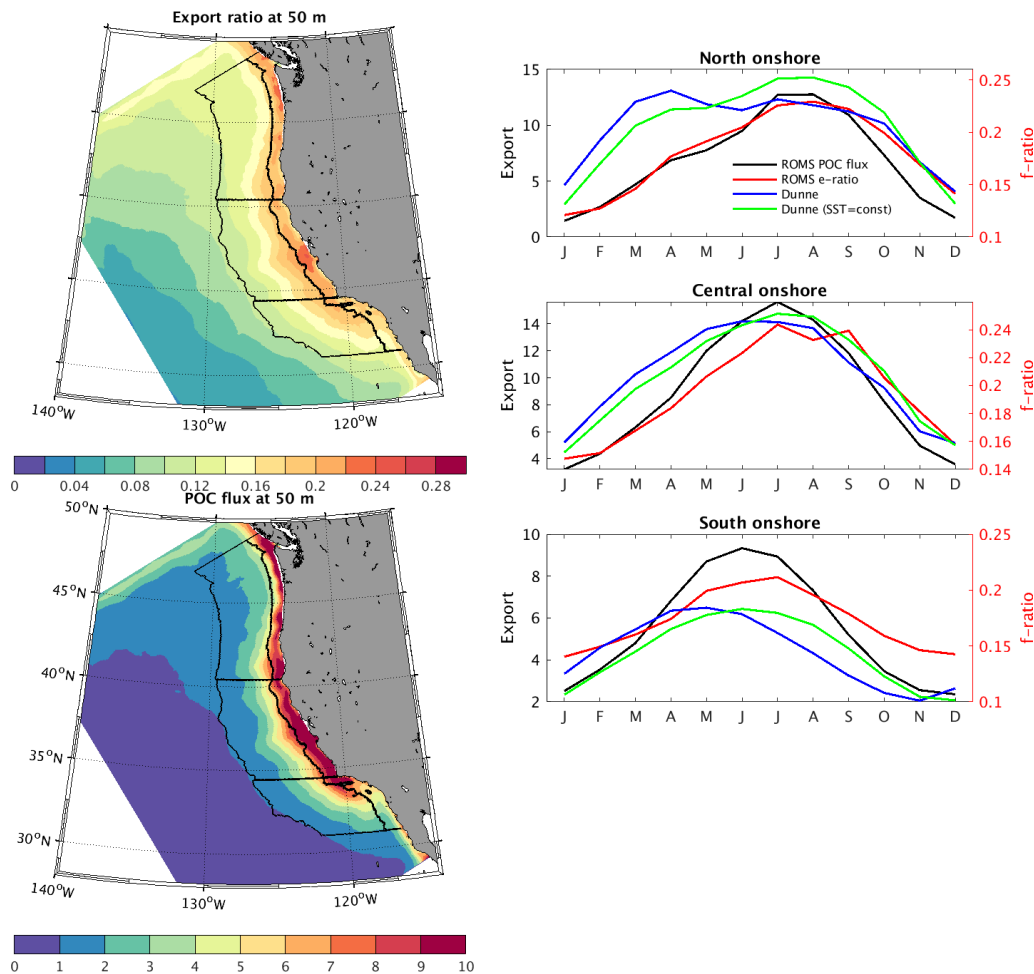


Figure 14: Annual mean and seasonal cycle of fraction of NPP that is exported in sinking particles (*i.e.*, pe-ratio; colored lines), and the export flux ($\text{mol } C \text{ m}^{-2} \text{ yr}^{-1}$; black line). Values diagnosed from model simulations (red line) are compared to an empirically derived algorithm (Dunne et al., 2005) based on *Chl* and SST (blue line), and from the same algorithm applied with constant SST (green line). Observed annual net community production, which should approximate export on an annual basis are measured to $5\text{-}10 \text{ mol } C \text{ m}^{-2} \text{ yr}^{-1}$ (Munro et al., 2013).

483 Annual mean export flux represents the transfer of biogenic material from the surface to depth,

484 and thus the influence of CCS productivity on air-sea CO_2 flux and thermocline properties. Model
485 simulated export production ranges from $\approx 1 - 10 \text{ mol } C \text{ m}^{-2} \text{ y}^{-1}$ (Fig. 14b). Few measurements
486 of export or net community production are available to evaluate the overall pattern of this flux. In
487 the SCB region, measurements of O_2 have been used to estimate NCP rates of 3 - 17 (mean 6.5)
488 $\text{mol } C \text{ m}^{-2} \text{ y}^{-1}$ (Munro et al., 2013), in line with modeled rates and with similar spatial patterns
489 of export (greatest along the Northern coast of the SCB). Munro et al. (2013) also combine in-
490 formation from ^{15}N uptake experiments and nitrate based new production ratios (Dugdale et al.,
491 1992; Eppley et al., 1992) with radiocarbon incubations to generate very similar estimates of the
492 magnitude and spatial variability of production in the SCB. Particle-based estimates of export (ne-
493 glecting the role of dissolved carbon) are lower than observed NCP as well as ROMS-BEC export,
494 $2 - 4 \text{ mol } C \text{ m}^{-2} \text{ y}^{-1}$ (Collins et al., 2011; Stukel et al., 2011), though enhanced particle export
495 associated with mesoscale fronts (2 - 3 times greater rates over smaller spatial and temporal scales
496 (Stukel et al., 2017)) highlights the potentially important role of subduction by eddies and fronts in
497 explaining mismatches between observations. Importantly, ROMS-BEC generates such mesoscale
498 features which contribute to model export estimates. While the model assumes particulate matter
499 is redistributed vertically without being transported by the lateral circulation, the magnitudes and
500 pattern of export are not substantially different from models that include explicit 3-dimensional
501 particle transport (Frischknecht et al., 2018).

502 Model export production is also similar to regional nutrient budget analyses which suggest NCP
503 averaging $7 - 9 \text{ mol } C \text{ m}^{-2} \text{ y}^{-1}$ over the broader CalCOFI region (Roemmich and McCallister,
504 1989; Bograd and Lynn, 2001). Similar nutrient budget analysis indicates that annual NCP should
505 be $\approx 17 \text{ mol } C \text{ m}^{-2} \text{ y}^{-1}$ off Monterrey Bay, again in line with ROMS-BEC estimates of export
506 for that region (Fig. 14b). As noted above, the fidelity of modeled nitrate distributions across the
507 model domain provides a critical broad-scale measure of net biological drawdown, and thus of net
508 community and export production.

509 Surface ocean CO_2 partial pressure and sea-to-air flux (Fig. 15) is reduced by net commu-
510 nity production, increased by surface heat flux, reduced by freshwater fluxes, and modulated by
511 upwelling and lateral circulation. It thus represents another important metric of overall system
512 function. The role of the CCS and its sub-regions in the atmosphere-ocean balance of CO_2 has
513 previously been investigated in several studies, both empirically (*e.g.*, Hales et al. (2012)) and in
514 models (*e.g.*, Fiechter et al. (2014); Turi et al. (2014) and references therein). We evaluated the
515 patterns of annual mean and summertime surface pCO_2 in the model hindcast simulation against
516 observations in the SOCATv6 database from 1995-2010 (Bakker and coauthors, 2016), and the as-
517 sociated air-sea fluxes over the modeled coastal region. Similar to observations over the simulated
518 period, strong CO_2 supersaturations are simulated in a narrow band of coastal water within about
519 100 km of the shore (Fig. 15). The values are highest along the central coast (35N-43N), lowest
520 in the northern CCS, and intermediate in the southern domain. The highest values are associated
521 with major topographic features, as previously noted by Fiechter et al. (2014).

522 The lack of apparent regional or seasonal bias in the the surface pCO_2 , together with the good
523 model-data agreement in surface buoyancy fluxes (Renault et al., 2021) and biological rates in these
524 simulations, suggests that the balance of processes regulating the model's surface CO_2 fluxes is
525 reliably captured. Consistent with previous studies, the net integrated CO_2 flux to the atmosphere
526 (an uptake of $1.4 \text{ Tg } C \text{ yr}^{-1}$) is found to be a relatively small residual of larger compensating
527 outgassing and ingassing fluxes. A detailed accounting of factors driving pCO_2 and air-sea flux
528 variability has been described (Fiechter et al., 2014; Turi et al., 2014). Our model predicts a

529 somewhat larger net uptake than previous studies, likely because our model domain extends farther
530 into the northern CCS where the combination of heat loss and fresh water forcing suppresses
531 surface pCO_2 . An extension of similarly detailed analysis of air-sea flux variability to the northern
532 CCS domain is left to future work.

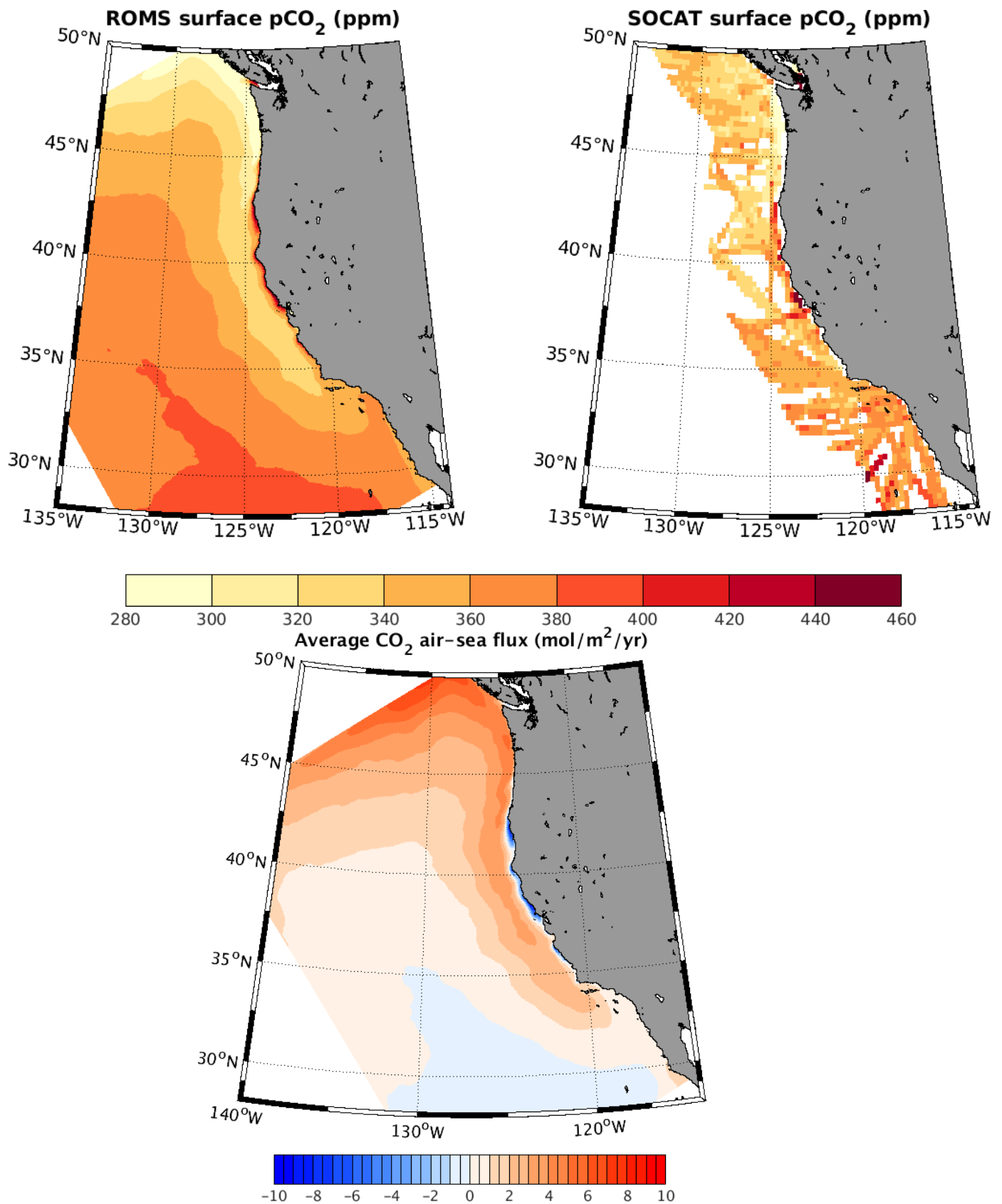


Figure 15: Surface ocean CO_2 partial pressure (ppm) and air-sea flux ($mol\ m^{-2}\ yr^{-1}$). (a) Annual surface pCO_2 from ROMS. (b) Surface pCO_2 from SOCATv6 gridded coastal dataset, averaged over all months from 1995-2010. (c) Annual air-sea CO_2 flux from ROMS. Negative values indicate outgassing to the atmosphere.

533 3.2 Thermocline

534 Here we test the model’s representation of biogeochemical properties below the mixed layer and
535 photic zone, *i.e.*, in the thermocline. We focus on distributions and variability of O_2 and arag-
536 onite saturation state (Ω_A) as these properties influence habitability for calcification and aerobic
537 respiration by marine animals. The values of Ω_A are calculated from model dissolved inorganic
538 carbon (DIC) and total alkalinity (Alk) using carbonate system equilibrium equations (CO2SYS)
539 (van Heuven et al., 2011). The variability of O_2 has been analyzed in greater detail using previous
540 simulations of this model with different atmospheric and physical boundary conditions (Durski
541 et al., 2017).

542 To evaluate the vertical structure of biogeochemical tracers, we turn to repeat hydrographic
543 sections (Fig. 16). Transects through three cross-sections spanning the Southern California Bight
544 (CalCOFI line 80), the central California coast (MBARI line 67), and the central Oregon coast
545 (Newport line) show the typical vertical and cross-shelf gradients of NO_3^- and O_2 . The downward
546 enhancement of NO_3^- and depletion of O_2 is a signal of the broad-scale vertical redistribution of
547 these elements by the formation and degradation of organic matter within the CCS, as well as the
548 gradients imported from the Pacific basin through the boundary conditions. The shoaling of the
549 isopleths of both quantities follows that of the isopycnal surfaces by upwelling along the coast. The
550 distributions of NO_3^- and O_2 are generally well reproduced by ROMS-BEC. The model somewhat
551 underestimates the slope of these isopleths very nearshore. This tendency is also reflected in, and
552 likely derived from, the same underestimate in the zonal tilt of isopycnal surfaces (Renault et al.,
553 2021).

554 Along isopycnal surfaces, O_2 generally increases with latitude and with distance from shore,
555 reflecting the contrasting properties carried by the broad offshore California current from the O_2 -
556 rich subarctic, and the narrow near-shore California Undercurrent that transports low- O_2 waters of
557 tropical origin northward along the slope. Both northern and southern end-member water types can
558 be seen on the isopycnal surface 26.5 (Fig. 17), which also comprises the source of water upwelling
559 onto the continental shelf along much of the US west coast (Pierce et al., 2012). On this and other
560 density surfaces, the distribution of O_2 in ROMS-BEC is consistent with climatological observa-
561 tions, suggesting that the balance of distinct water masses and the respiratory modifications they
562 experience in the interior of the domain are relatively well represented in the model. Thermocline
563 nutrient distributions exhibit a similar model skill (not shown).

564 The O_2 in the thermocline of the CCS is highly variable, with standard deviations of ≈ 20 mmol
565 m^{-3} that are on average 15-20% of the mean O_2 across the historical measurements (Fig. 17). The
566 magnitudes and patterns of variance are well captured by the ROMS hindcast. In both the model
567 hindcast and in observations, the standard deviation of monthly O_2 is ≈ 5 times larger than that
568 of the climatological seasonal cycle. Thus, a large majority of the O_2 variability is explained
569 by non-seasonal time-scales, including large eddy-driven fluctuations (Frenger et al., 2018) and
570 low-frequency climate variability (Buil and Lorenzo, 2017; McClatchie et al., 2010). Both model
571 and observations indicate that O_2 variation peaks slightly offshore, in a pattern resembling that of
572 eddy kinetic energy (Renault et al., 2021), and reflecting the role of eddies in transporting hypoxic
573 waters offshore (Frenger et al., 2018). The high interannual to decadal O_2 variability observed in
574 the central North Pacific, which reaches its maximum on the isopycnal 26.5, is not included in the
575 climatological boundary conditions, and thus likely accounts for the model bias toward low O_2
576 variance in the north of the domain.

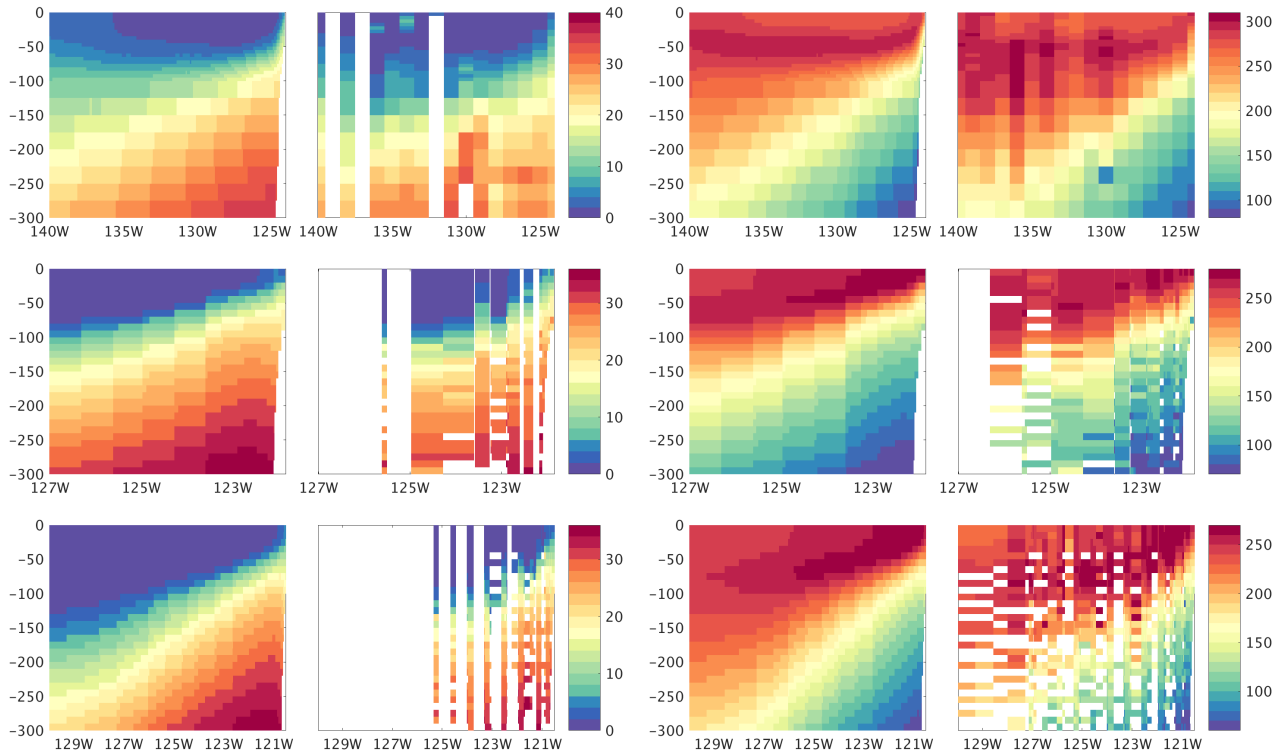


Figure 16: Vertical sections of annual mean NO_3^- (left) and O_2 (right) (mmol m^{-3}), from ROMS (left) and WOD (right) at the latitudes with regular observations by repeat hydrographic surveys. The lines span the northern CCS (upper row, $\approx 44.5^\circ\text{N}$, nearest the Newport OR), the central CCS (middle row, MBARI line 67), and the southern CCS (bottom row, CalCOFI line 80). Locations of observations are shown in Fig. 3.

577 Variations in O_2 on depth surfaces in the thermocline are also dominated by interannual anomalies.
 578 The fluctuating O_2 at depths of 100-200 m are highly correlated with density ($R^2 \geq 0.5$)
 579 throughout the CCS, reflecting the importance of isopycnal heaving of the background O_2 gradient (Ito et al., 2019). Similar correlations are observed in ROMS and the World Ocean Database
 580 (Fig. 18). In the model, the largest such anomaly is associated with the ENSO event in 1997-98, in
 581 which deepened isopycnals yield high O_2 conditions that last for ≈ 1 year. The signal is recorded
 582 in the central CCS as well, although the magnitude of the anomalies is reduced by $\approx 50\%$ relative
 583 to the better-sampled event in the SCB.
 584

585 The variability of O_2 within an isopycnal surface can be used to account for this portion of
 586 variance, leaving only lateral circulation and respiration. We find that along $\sigma_\theta = 26.5 \text{ kg m}^{-3}$,
 587 a large fraction of O_2 variability is correlated with salinity (Fig. 19), commonly used as a proxy
 588 for tropical low-oxygen and high-salinity water transported poleward in the coastal undercurrent
 589 (Meinvielle and Johnson, 2013). Interannual variability in respiration rates on this surface also
 590 accounts for $\approx 20\%$ of isopycnal O_2 variance, and is in turn correlated with the depth of the density
 591 layer (Deutsch et al., 2011). Historical observations show a declining strength of correlation
 592 between S and O_2 with latitude, suggesting that variability from dynamics other than the CUC become
 593 an increasingly important source of O_2 variability to the north. Indeed, the variability of O_2 in
 594 central mode water from the open North Pacific is most pronounced on this density surface (?Buil

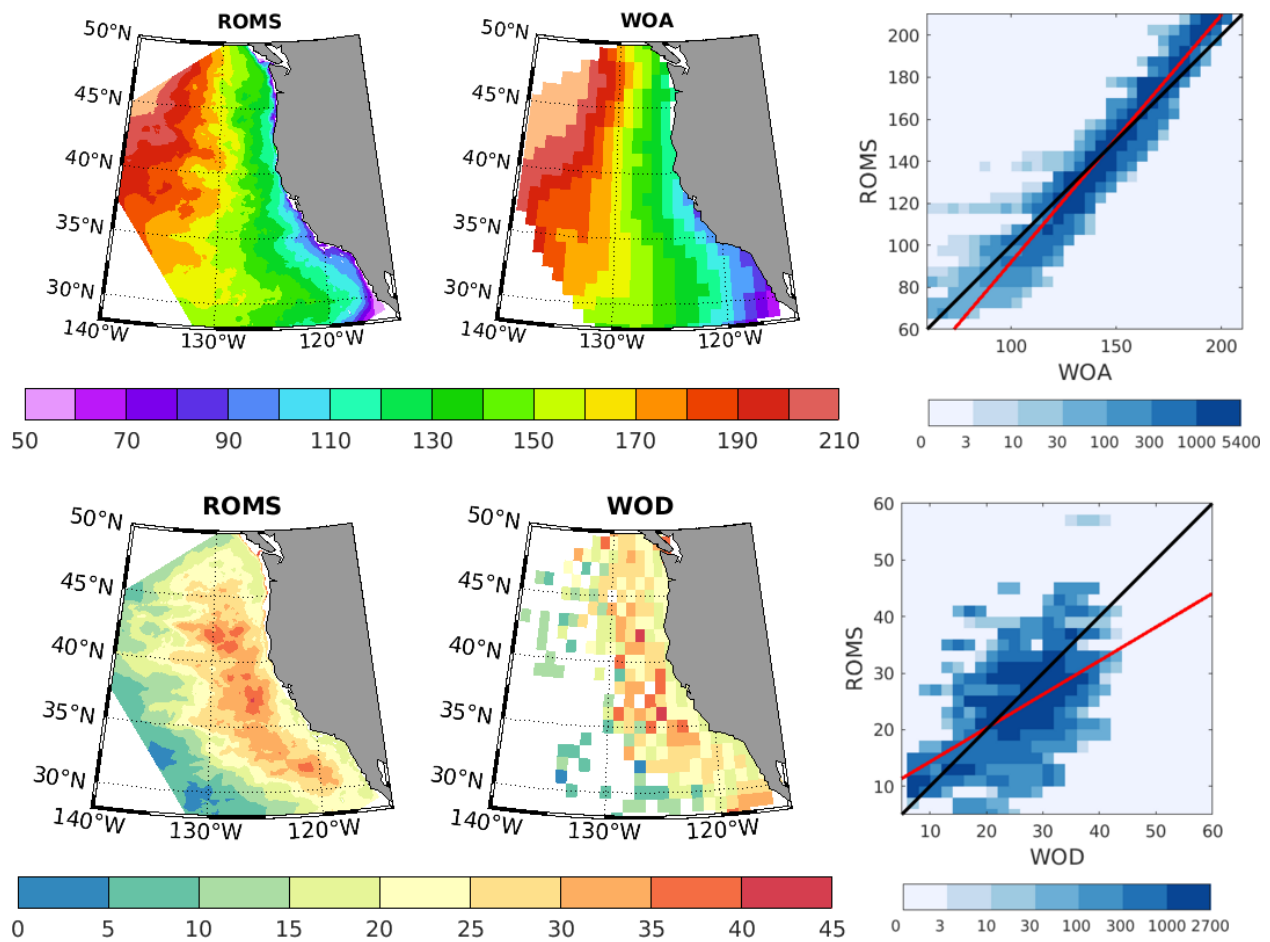


Figure 17: Thermocline O_2 in model simulations and observations of the climatological mean (upper) and standard deviation (lower). All maps are interpolated to a potential density surface (26.5) surface chosen as the density class of waters upwelling onto the shelf in summer (a.k.a. “source water”). The mean O_2 maps are from summer months (JJA), however other seasons reveal similar model fidelity. Observed mean summer O_2 is from objectively mapped climatology (WOA). Variability is mapped as the standard deviation of monthly values from the World Ocean Database covering 1955-2013 (lower right), and it is predominantly due to interannual variability rather than the seasonal cycle (see text). The correlations between ROMS and WOD are highly significant ($p \ll 0.01$) for both mean O_2 and its variability, with squared Pearson correlation coefficients (R^2) of 0.95 for the mean state and 0.35 for the spatial pattern of temporal variability.

595 and Lorenzo, 2017), and the North Pacific Current is thought to transport isopycnal O_2 anomalies
 596 into the northern CCS and contribute to the significant interdecadal fluctuations observed in the
 597 CCS [insert citation in comment here]. However, this longer term variance along isopycnals is not
 598 represented in the climatological biogeochemical boundary conditions in the model. Thus the rela-
 599 tively constant correlation between S and O_2 along isopycnal surfaces across latitude in the model
 600 may stem from the lack of O_2 variability on isopycnal surfaces as those waters enter the domain
 601 from the open North Pacific. A complete attribution of the observed magnitudes of O_2 variance

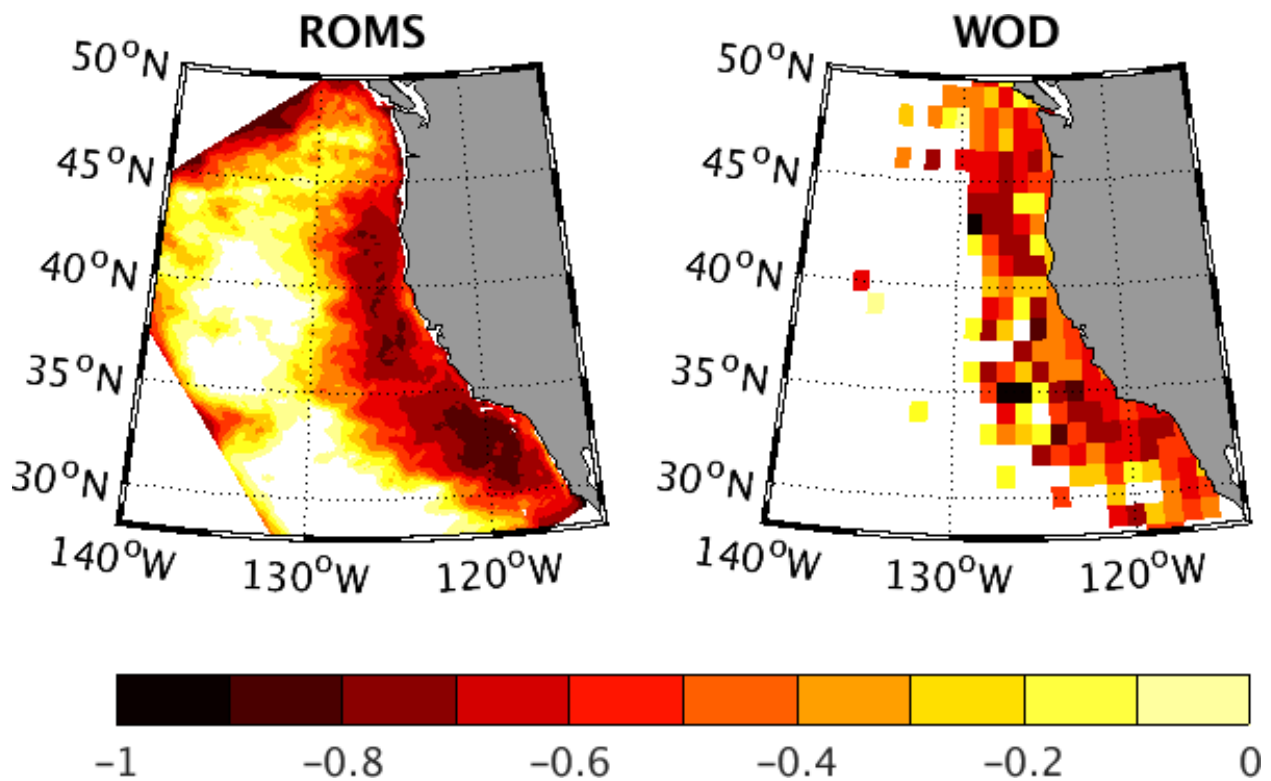


Figure 18: Correlation between O_2 and density anomalies at 100 m, in ROMS-BEC (left) and WOD (right). Interannual variations in subsurface O_2 are highly correlated with density ($R^2 \geq 0.5$) throughout the CCS, and the strength of the relationship is similar in model and observations.

602 in source waters to the CCS will require inclusion of anomalies entering from the broader North
 603 Pacific (Deutsch et al., 2006; Kwon et al., 2016) and is left for future work.

604 In addition to low coastal O_2 , the CCS is characterized by shallow depth horizons for carbonate
 605 saturation (Fig. 20). Below ≈ 100 m depth, carbonate concentrations are commonly undersaturated
 606 with respect to aragonite mineral formation, and thus inhibit shell formation by calcifying organ-
 607 isms. The aragonite saturation state, Ω_A , is predicted to fall below saturation ($\Omega_A < 1$) along
 608 most of the coast in summer, consistent with observations in NOAA coastal surveys (e.g., Feely
 609 et al. (2008)). Coastal hydrographic surveys reveal a strong mesoscale patchiness to the carbonate
 610 saturation state, likely reflecting mesoscale eddies and submesoscale features. In the multi-annual
 611 mean distribution of Ω_A , the model hindcast captures the scale and intensity of undersaturated
 612 conditions well.

613 Both low O_2 and low Ω_A have been implicated as primary factors mediating the influence of
 614 climate on organism fitness and species habitat in the CCS (Howard et al., 2020; Busch and McEl-
 615 hany, 2016). We compare decadal trends in both these properties from the hindcast simulations
 616 to the observed changes over time. For Ω_A , the measurements are too sparse and the distribution
 617 too patchy to define a robust trend, even over the short model period. For each property, time
 618 series are shown for the regions with the most data coverage (Fig. 21): northern CCS for aragonite
 619 saturation, and southern CCS for O_2 . In both cases, the trend in the data is within the uncertainty
 620 in the measurements.

621 As a metric of variability in these habitat constraints, we computed the volume of water subject

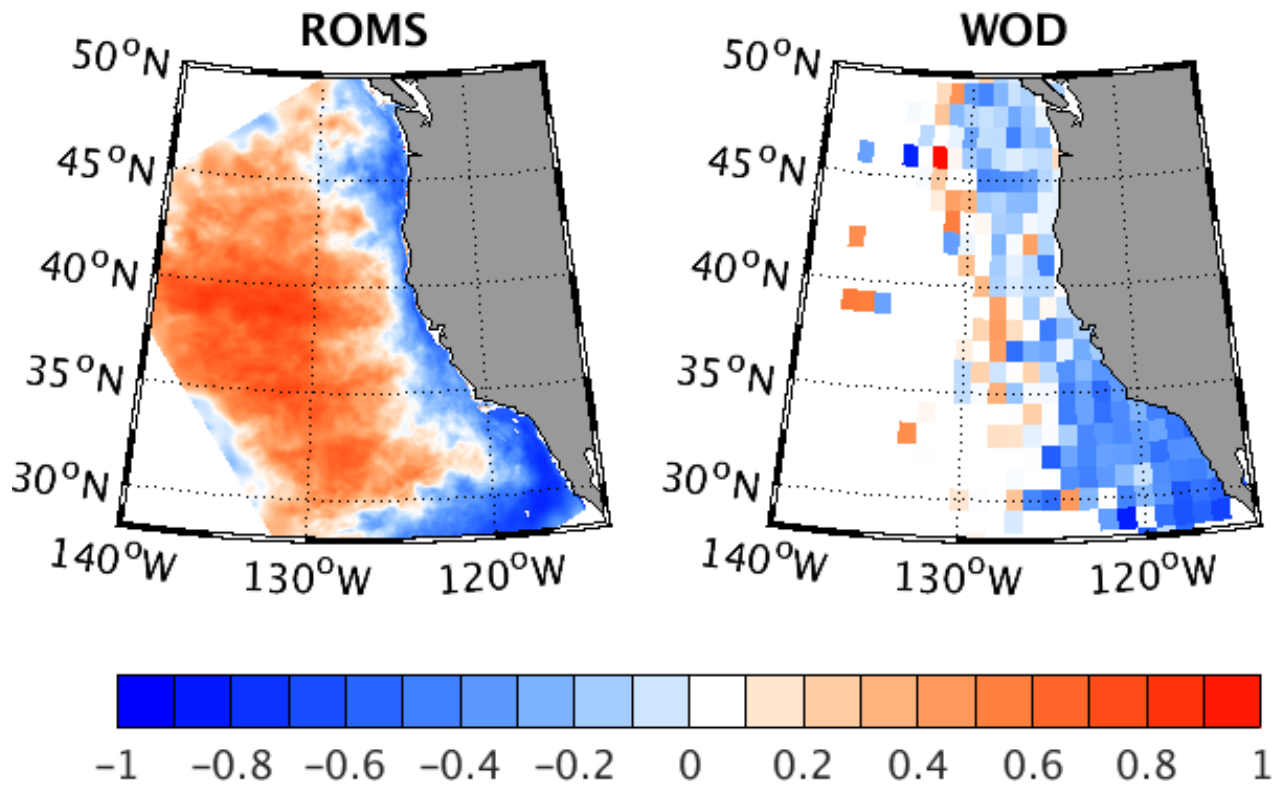


Figure 19: Correlation coefficient between O_2 and salinity anomalies in (left) ROMS, and (right) WOD along the isopycnal surface $\sigma_\theta = 26.5 \text{ kg m}^{-3}$. Prevailing negative values indicate that high S occurs when O_2 is low. To maximize data availability, the observational correlation is based on monthly anomalies in WOD from the period 1955-2013. The correlations are of similar magnitude when confined to the simulated period, 1995-2010, but are only available in the Southern California Bight.

622 to hypoxic or corrosive conditions. We use a constant O_2 level of 100 mmol m^{-3} as a simple
 623 indicator of hypoxic constraints, recognizing that this value varies among species, and depends
 624 on other factors, including temperature (Deutsch et al., 2015). Corrosive conditions are defined
 625 by simple thermodynamic undersaturation ($\Omega_A < 1$), though biological sensitivities may begin at
 626 higher thresholds. Water volumes are computed as the sum of grid cell volumes with $O_2 < 100$
 627 mmol m^{-3} or $\Omega_A < 1$ that are on the continental shelf ($-z < 200 \text{ m}$).

628 The volume of hypoxic and corrosive water in the CCS varies strongly over latitude and time
 629 (Fig. 22). For both properties, restriction of putative habitat volume is stronger to the north of Pt.
 630 Conception, opposite the latitudinal gradient of O_2 and Ω_A . The corrosive volumes are much larger
 631 than hypoxic volumes, exceeding 90% of water volume in northern latitudes during the summer
 632 upwelling season, consistent with NOAA survey data. On an annual basis, waters with a more
 633 stringent criterion for calcification ($\Omega_A < 2$) are about twice as voluminous still, primarily because
 634 the length of the season with low carbonate is broadened. Hypoxic conditions occupy a smaller
 635 fraction of shelf waters, but reach 30% of shelf water volume over a broad latitude range. The
 636 fractional coverage by hypoxia peaks around 45°N , on the Oregon coast, and it is quite small in the
 637 Southern California Bight, where O_2 declines most sharply below 200 m rather than on the shelf.

638 An analogous figure is shown in Renault et al. (2021) for the along-coast and temporal vari-

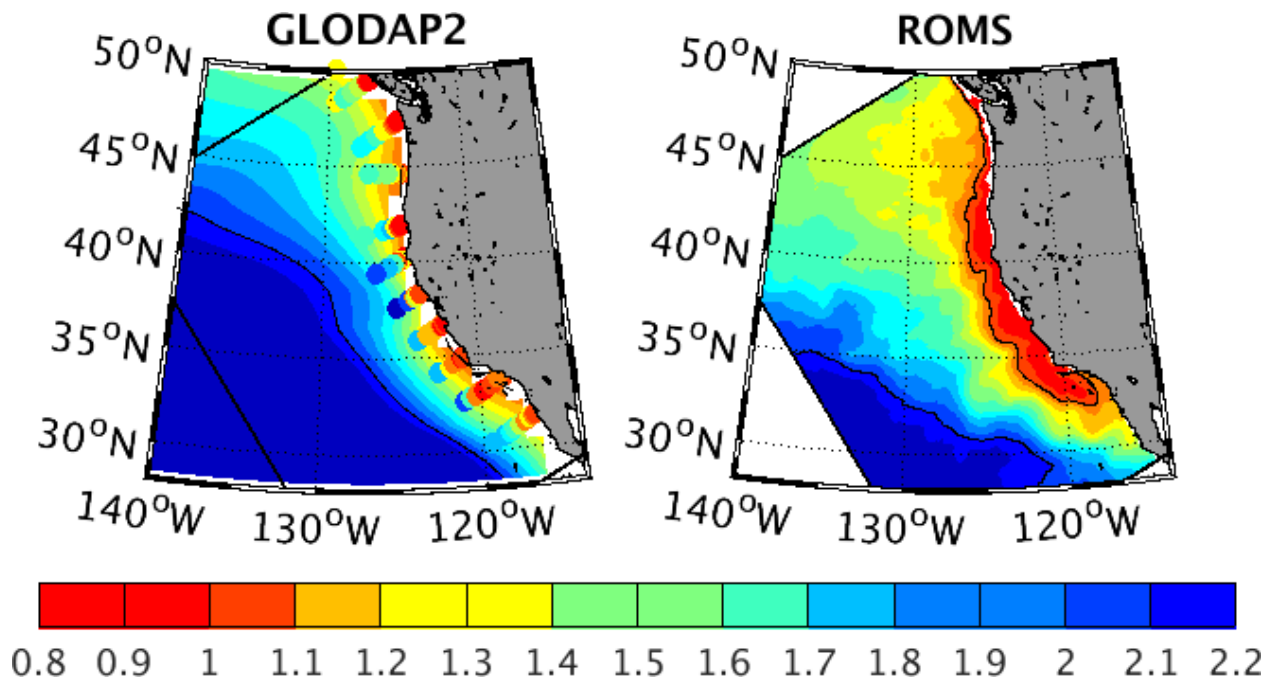


Figure 20: Aragonite saturation state (Ω_A) at 100 m from observations (left) and ROMS-BEC (right). Observations are from large-scale objectively analyzed fields (color field, GLODAP2 (Lauvset and coauthors, 2016)), and from NOAA coastal surveys in the summer of 2007 (circles, Feely et al. (2008)). Model distribution is averaged over summer (JJA) from 2004-2010 climatology.

639 ability of the sea-surface height and depth of the $\sigma_\theta = 26.5 \text{ kg m}^{-3}$ depth anomalies. Relative
 640 to the quantities shown in Fig. 22, they exhibit more along-coast coherence and a more dominant
 641 seasonal cycle, with less evident interannual variability than shown here, apart from the 1997-98
 642 ENSO event. This indicates somewhat smoother physical fields than biogeochemical ones, reflec-
 643 tive of non-conservative biogeochemical processes acting on top of the broader patterns of physical
 644 circulation influence.

645 Variability in both habitat constraints is largely synchronous (Fig. 22c,d), reflecting the strong
 646 control on both O_2 and Ω_A by the effects of cumulative organic matter respiration. For both
 647 volumes, the fractional variation is similar and substantial, reaching $\approx 50\%$ of the mean across
 648 much of the latitude range. Variability of corrosive volumes is greatly attenuated south of Pt.
 649 Conception. Simulated hypoxic volumes increased sharply during the early 2000s off Oregon,
 650 when major ecosystem die-offs were attributed to the onset of extreme hypoxia there (Chan et al.,
 651 2008).

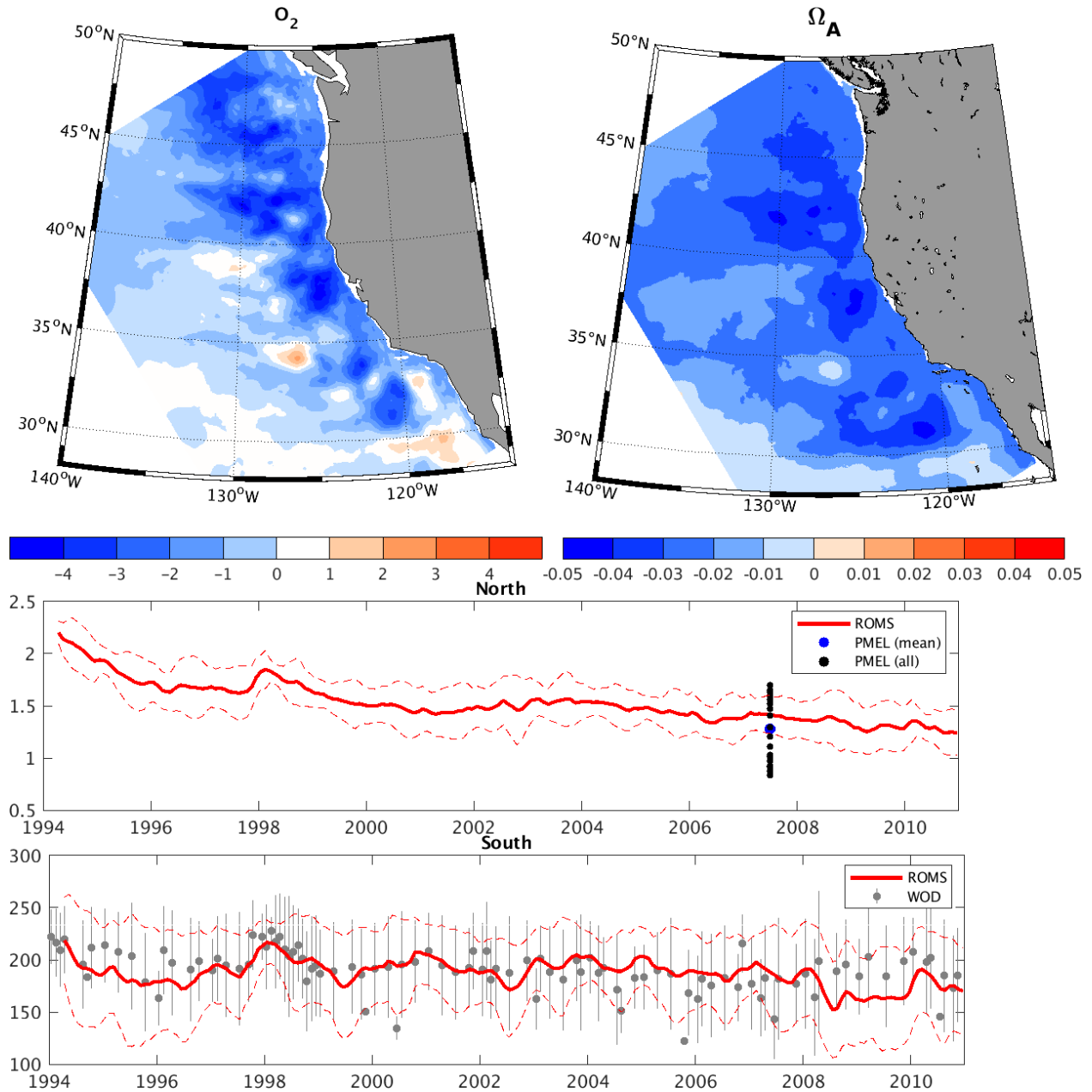


Figure 21: Trends in thermocline O_2 and Ω_A over the simulated period. Maps of the linear trend are shown in upper panels. Time series (lower panels) are shown for the regions with the most data coverage for each tracer: northern CCS for carbonate, and southern CCS for O_2 . For Ω_A all available profiles are shown. For O_2 , the mean value and standard deviation are plotted as box-whisker for each month in the WOD.

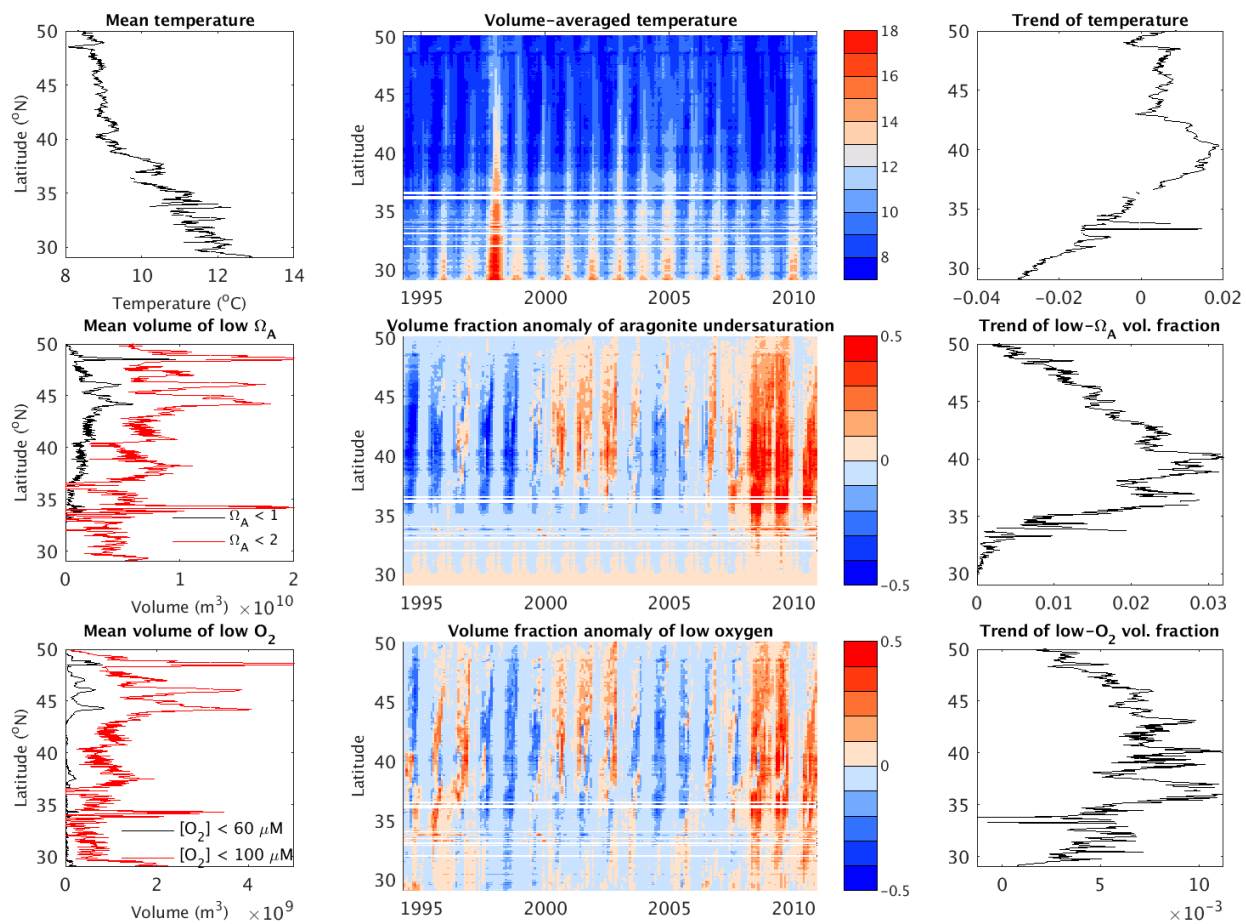


Figure 22: Top row: mean temperature and temporal trend ($^{\circ}\text{C year}^{-1}$) of water on the shelf ($-z < 200$ m). Middle and bottom rows: volume of corrosive and hypoxic water over time and latitude. Water volumes are computed as sum of grid cell volumes with $O_2 < 100$ mmol m^{-3} (bottom) and $\Omega_A < 1$ (middle) that are on the shelf ($-z < 200$ m). Mean values for each month are shown in the left column, and anomalies, computed as a fractional deviation from the climatological mean monthly volume at each latitude, in the middle column. Trends over time are shown in the right column.

4 Conclusions

We present model simulations of ecosystem and biogeochemical cycles in the CCS that reproduce the broad patterns of processes and states observed in this region over the past couple of decades. Our results demonstrate that productivity of the CCS reflects a complex interplay of factors. The limitation by the physical supply and removal of macronutrients (nitrate) provides the dominant seasonal and spatial pattern of NPP, but with significant constraints from light and Fe on a seasonal basis, especially in the northern CCS. Interannual variations in NPP are reasonably well predicted by fluctuations in pycnocline depth that modulate the rates of surface nutrient supply. Expanded datasets on near-surface Fe concentrations are needed to better establish its role as a limiting factor for growth in the CCS. A significant correlation between model NPP and surface irradiance suggests that changes in light are also influential. Together, these results suggest that an index of NPP that accounts for both regional pycnocline structure and cloud cover would be more skillful than one based only on coastal winds (*e.g.*, Bakun (1990); Jacox et al. (2018)). Our results highlight the value of continued measurements of the depth of the chlorophyll maximum.

Biogeochemical properties of subsurface waters in the CCS are also well reproduced by model simulations. The amplitude of interannual variability in NO_3^- at the base of the photic zone and of O_2 in the thermocline are also both strongly correlated to undulations of the pycnocline. The largest such anomalies in our simulation period were associated with the 1997-98 ENSO event, whose amplitude of density and O_2 anomalies remains coherent over a wide latitude band, albeit with declining magnitude. For NO_3^- , the overall variance is somewhat lower, and the strength of density correlations is somewhat higher in model output than in observations in the northern domain. This suggests an important role for anomalies entering the CCS from the subarctic North Pacific, an HNLC region. Basin-scale changes in biogeochemical properties are known to be exceptionally high at the gyre boundary $\approx 45^\circ N$ (Mecking et al., 2008), and these remote anomalies are likely to play an important but uncertain role in the variability observed in the CCS. Similarly, inputs from terrestrial and riverine sources of nutrients and organic carbon could contribute to interannual variability in the system that is currently difficult to constrain from observational data. The impact of nearshore influences on the mean state suggests that its contribution to variability could also be substantial. Evaluating these remote influences from both the open ocean and from boundary inputs, using empirically-based time-dependent biogeochemical boundary conditions is an important avenue for future research.

The variability of biogeochemical properties leads to significant changes in the volume of waters characterized by biologically stressful conditions of hypoxia and carbonate undersaturation. In the volume anomalies for both habitat constraints, there is strong coherence across the CCS. Years with unusually large volumes of hypoxic or corrosive water offer few obvious latitudinal refuges. The onset of these conditions tends to propagate from the central CCS ($\approx 40^\circ N$), arriving in the northern CCS with a 2-3 month delay. Thus, monitoring hypoxia and CO_2 system parameters in the central CCS may offer some seasonal predictability for northern ecosystem impacts.

690 **Codes and Simulation Data** The physical and biogeochemical codes used for our simulations
691 are at <https://github.com/UCLA-ROMS/Code>. Simulation model output archive data can be made
692 available by email requests to the Corresponding Authors.

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700 ses, including the National Ocean Data Center (NOAA), the remote sensing products from NASA,
701 and the California Cooperative Oceanic Fisheries Investigations (CalCOFI).

702 5 Appendix: Biogeochemical Model

703 Here, for completeness, we summarize the equations of the Biogeochemical Elemental Cycling
 704 (BEC) model in the implementation used for this work. This formulation is based on the original
 705 version presented in Moore et al. (2004).

706 5.1 Variables and parameters

707 5.1.1 Prognostic variables

Name	Description	Units
N_{NH_4}	ammonium	mmol N/m^3
N_{NO_2}	nitrite	mmol N/m^3
N_{NO_3}	nitrate	mmol N/m^3
N_{N_2O}	nitrous oxide	mmol N/m^3
N_{N_2}	nitrogen	mmol N/m^3
N_{don}	dissolved organic nitrogen	mmol N/m^3
P_{PO_4}	Phosphorus	mmol P/m^3
P_{dop}	dissolved organic phosphorus	mmol P/m^3
O_2	dissolved oxygen	mmol O_2/m^3
Fe	iron	nmol Fe/m^3
Fe_{sp}	small phytoplankton iron	mmol Fe/m^3
Fe_{diat}	diatom iron	mmol Fe/m^3
Fe_{diaz}	diazotroph iron	mmol Fe/m^3
Fe_{dofe}	dissolved organic iron	mmol Fe/m^3
Ca_{sp}	small phytoplankton calcium carbonate	mmol $CaCO_3/m^3$
Si_{diat}	diatom silicate	mmol SiO_2/m^3
Si_{SiO_2}	Silicate	mmol SiO_2/m^3
Alk	total alkalinity	mmol/ m^3
C_{DIC}	dissolved inorganic carbon	mmol C/m^3
C_{sp}	small phytoplankton carbon	mmol C/m^3
C_{diat}	diatom carbon	mmol C/m^3
C_{diaz}	diazotroph carbon	mmol C/m^3
C_{doc}	dissolved organic carbon	mmol C/m^3
Chl_{sp}	small phytoplankton chlorophyll	mmol Chl/m^3
Chl_{diat}	diatom chlorophyll	mmol Chl/m^3
Chl_{diaz}	diazotroph chlorophyll	mmol Chl/m^3
C_{zoo}	zooplankton carbon	mmol C/m^3
POC_{sed}	organic carbon in sediment	mmol C/m^2
$CaCO_{3,sed}$	inorganic carbon in sediment	mmol C/m^2
Si_{sed}	Silicate in sediment	mmol Si/m^2
I_{SW}	Penetrative solar heat flux	W/m^2

5.1.2 Local variables

Name	Description	Units
$J_{sp,no3}^u$	nitrate uptake by small phytoplankton	mmol $N/m^3/sec$
$J_{sp,nh4}^u$	ammonium uptake by small phytoplankton	mmol $N/m^3/sec$
$J_{diat,no3}^u$	nitrate uptake by diatoms	mmol $N/m^3/sec$
$J_{diat,nh4}^u$	ammonium uptake by diatoms	mmol $N/m^3/sec$
J_{ammox}	rate of NH_4 oxidation to NO_2	mmol $N/m^3/sec$
J_{nit}	rate of NO_2 oxidation to NO_3	mmol $N/m^3/sec$
$J_{no3,no2}^{denit}$	amount of NO_3 converted to NO_2 by water column denitrif.	mmol $N/m^3/sec$
J_{sed}^{denit}	amount of NO_3 converted to N_2 by benthic denitrif.	mmol $N/m^3/sec$
$J_{no2,n2o}^{denit}$	amount of NO_2 converted to N_2O by water column denitrif.	mmol $N/m^3/sec$
$J_{no2,n2}^{denit}$	amount of NO_2 converted to N_2 by water column denitrif.	mmol $N/m^3/sec$
J_{diaz}^{Nfix}	total N_2 -fixation by diazotrophs	mmol $N/m^3/sec$
$J_{diaz}^{Nexcrete}$	N excreted by diazotrophs	mmol $N/m^3/sec$
J_{diaz}^{photoN}	N fixed by diazotroph (non-excreted)	mmol $N/m^3/sec$
$J_{sp,po4}^u$	phosphate uptake by small phytoplankton	mmol $P/m^3/sec$
$J_{diat,po4}^u$	phosphate uptake by diatoms	mmol $P/m^3/sec$
$J_{diaz,po4}^u$	phosphate uptake by diazotrophs	mmol $P/m^3/sec$
P_{diaz}^{remain}	remaining diazotrophs phosphate	mmol $P/m^3/sec$
$J_{diaz,dop}^{loss}$	non-grazing mortality of diazotrophs routed to POC	mmol $P/m^3/sec$
$J_{diaz,dip}^{loss}$	non-grazing mortality of diazotrophs routed to PIC	mmol $P/m^3/sec$
$J_{sp,caco3}^{prod}$	$CaCO_3$ production by small phytoplankton	mmol $CaCO_3/m^3/sec$
$J_{sp,C}^{photo}$	carbon uptake by photosynthesis in small phytoplankton	mmol $C/m^3/sec$
$J_{diat,C}^{photo}$	carbon uptake by photosynthesis in diatoms	mmol $C/m^3/sec$
$J_{diaz,C}^{photo}$	carbon uptake by photosynthesis in diazotrophs	mmol $C/m^3/sec$
J_{sp}^{grz}	grazing loss for small phytoplankton	mmol $C/m^3/sec$
$J_{sp,dic}^{grz}$	grazed small phytoplankton routed to DIC	mmol $C/m^3/sec$
$J_{sp,doc}^{grz}$	grazed small phytoplankton routed to DOC	mmol $C/m^3/sec$
$J_{sp,poc}^{grz}$	grazed small phytoplankton routed to POC	mmol $C/m^3/sec$
$J_{sp,zoo}^{grz}$	grazed small phytoplankton routed to new zooplankton biomass	mmol $C/m^3/sec$
J_{diat}^{grz}	grazing loss for diatoms	mmol $C/m^3/sec$
$J_{diat,dic}^{grz}$	grazed diatoms routed to DIC	mmol $C/m^3/sec$
$J_{diat,doc}^{grz}$	grazed diatoms routed to DOC	mmol $C/m^3/sec$
$J_{diat,poc}^{grz}$	grazed diatoms routed to POC	mmol $C/m^3/sec$
$J_{diat,zoo}^{grz}$	grazed diatoms routed to new zooplankton biomass	mmol $C/m^3/sec$
J_{diaz}^{grz}	grazing loss for diazotrophs	mmol $C/m^3/sec$
$J_{diaz,dic}^{grz}$	grazed diazotrophs routed to DIC	mmol $C/m^3/sec$
$J_{diaz,doc}^{grz}$	grazed diazotrophs routed to DOC	mmol $C/m^3/sec$
$J_{diaz,poc}^{grz}$	grazed diazotrophs routed to POC	mmol $C/m^3/sec$
$J_{diaz,zoo}^{grz}$	grazed diazotrophs routed to new zoo biomass	mmol $C/m^3/sec$

J_{sp}^{agg}	aggregation loss of small phytoplankton	mmol $C/m^3/sec$
J_{diat}^{agg}	aggregation loss of diatoms	mmol $C/m^3/sec$
J_{sp}^l	non-grazing mortality of small phytoplankton	mmol $C/m^3/sec$
$J_{sp,dic}^l$	non-grazing mortality of small phytoplankton routed to DIC	mmol $C/m^3/sec$
$J_{sp,doc}^l$	non-grazing mortality of small phytoplankton routed to DOC	mmol $C/m^3/sec$
$J_{sp,poc}^l$	non-grazing mortality of small phytoplankton routed to POC	mmol $C/m^3/sec$
J_{diat}^l	non-grazing mortality of diatoms	mmol $C/m^3/sec$
$J_{diat,dic}^l$	non-grazing mortality of diatoms routed to DIC	mmol $C/m^3/sec$
$J_{diat,doc}^l$	non-grazing mortality of diatoms routed to DOC	mmol $C/m^3/sec$
$J_{diat,poc}^l$	non-grazing mortality of diatoms routed to POC	mmol $C/m^3/sec$
J_{diaz}^l	non-grazing mortality of diazotrophs	mmol $C/m^3/sec$
$J_{diaz,dic}^l$	non-grazing mortality of diazotrophs routed to DIC	mmol $C/m^3/sec$
$J_{diaz,doc}^l$	non-grazing mortality of diazotrophs routed to DOC	mmol $C/m^3/sec$
$J_{diaz,poc}^l$	non-grazing mortality of diazotrophs routed to POC	mmol $C/m^3/sec$
J_{zoo}^l	zooplankton mortality	mmol $C/m^3/sec$
$J_{zoo,dic}^l$	zooplankton mortality routed to DIC	mmol $C/m^3/sec$
$J_{zoo,doc}^l$	zooplankton mortality routed to DOC	mmol $C/m^3/sec$
$J_{zoo,poc}^l$	zooplankton mortality routed to POC	mmol $C/m^3/sec$
f_{zoo}^d	fractional factor for routing of zoo losses	no units
J_{sp}^{agg}	aggregation of small phytoplankton	mmol $C/m^3/sec$
J_{diat}^{agg}	aggregation of diatoms	mmol $C/m^3/sec$
J_{diaz}^{agg}	aggregation of diazotrophs	mmol $C/m^3/sec$
$J_{don,nh4}$	amount of dissolved organic N remineralized	mmol $N/m^3/sec$
$J_{doc,dic}$	amount of dissolved organic C remineralized	mmol $C/m^3/sec$
$J_{dop,po4}$	amount of dissolved organic P remineralized	mmol $P/m^3/sec$
$J_{dofe,fe}$	amount of dissolved Fe remineralized	mmol $Fe/m^3/sec$
J_{other}^{remin}	remineralization in the sediments by processes other than oxic remin. and denitrif.	mmol $C/m^3/sec$
J_{poc}^{prod}	amount of particulate organic C produced	mmol $C/m^3/sec$
J_{pic}^{prod}	amount of particulate inorganic C produced	mmol $C/m^3/sec$
J_{PSi}^{prod}	amount of particulate organic Si produced	mmol $Si/m^3/sec$
J_{pfe}^{prod}	amount of particulate organic Fe produced	mmol $Fe/m^3/sec$
J_{poc}^{remin}	remineralized particulate organic C	mmol $C/m^3/sec$
J_{pic}^{remin}	remineralized particulate inorganic C	mmol $C/m^3/sec$
J_{PSi}^{remin}	remineralized particulate organic Si	mmol $SiO_2/m^3/sec$
J_{pfe}^{remin}	remineralized particulate organic Fe	mmol $Fe/m^3/sec$
J_{Fe}^{scav}	scavenging of inorganic Fe	mmol $Fe/m^3/sec$
J_{Fe}^{sed}	remineralized Fe from sediment	mmol $Fe/m^3/sec$

PAR	photosynthetically active radiation	W/m ²
$\Phi_{poc}^{mineral}$	incoming mineral associated particulate C flux	mmol C cm/m ³ /sec
Φ_{poc}^{out}	particulate C-flux buried in the sediments	mmol C /m ² /sec
Φ_{poc}^{free}	free associated particulate C flux	mmol C /m ² /sec
Φ_{pic}^{sol}	soluble and pic associated particulate C flux	mmol C /m ² /sec
Φ_{pic}^{nonsol}	non-soluble and pic associated particulate C flux	mmol C /m ² /sec
Φ_{psi}^{sol}	soluble and SiO ₂ associated particulate C flux	mmol SiO ₂ /m ² /sec
Φ_{psi}^{nonsol}	non-soluble and SiO ₂ associated particulate C flux	mmol SiO ₂ /m ² /sec
Φ_{dust}^{sol}	soluble and dust associated particulate C flux	mmol C /m ² /sec
Φ_{dust}^{nonsol}	non-soluble and dust associated particulate C flux	mmol C /m ² /sec
Φ_{pfe}^{sol}	soluble and dust associated particulate C flux	mmol Fe /m ² /sec
Φ_{pfe}^{nonsol}	non-soluble and dust associated particulate C flux	mmol Fe /m ² /sec
$\Phi_{poc}^{remin, sed}$	remineralized organic C flux from sediment	mmol C/m ² /sec
$\Phi_{pic}^{remin, sed}$	remineralized inorganic C flux from sediment	mmol C/m ² /sec
$\Phi_{Si}^{remin, sed}$	remineralized Si flux from sediment	mmol C/m ² /sec
$\Phi_{O_2}^{air}$	O ₂ air-sea flux	mmol O ₂ /m ² /sec
$\Phi_{N_2O}^{air}$	N ₂ O air-sea flux	mmol N/m ² /sec
$\Phi_{N_2}^{air}$	N ₂ air-sea flux	mmol N/m ² /sec
$\Phi_{CO_2}^{air}$	CO ₂ air-sea flux	mmol C/m ² /sec
Sc	Schmidt Number	-
Ws	Wind speed at 10 m	m/sec
PV	Piston Velocity	m/sec

712

713 5.2 Ecosystem parameters

Parameters	Description	Values	Units
Model grid			
Δz	Model layer thickness	variable	meters
k_ρ	Index of model vertical level at tracer points	1-60	no units
Carbon			
PC_{ref}^{sp}	max phyto C-specific growth rate at Tref (GD98) for small phytoplankton	3.0	1/d
PC_{ref}^{diat}	max phyto C-specific growth rate at Tref (GD98) for diatoms	3.0	1/d

PC_{ref}^{diaz}	max phyto C-specific growth rate at Tref (GD98) for diazotrophs	0.4	1/d
α_{chl}	chlorophyll-specific initial slope of P vs. I curve for diatoms and small phytoplankton	0.3	mmol C m ² /(mg Chl W day)
α_{chl}^{diaz}	chlorophyll-specific initial slope of P vs. I curve for diazotrophs	0.036	mmol C m ² /(mg Chl W day)

Nutrient limitation

$k_{NO_3}^{sp}$	Small phyto. half saturation constant for NO_3 uptake	0.5	mmol N /m ³
$k_{NO_3}^{diat}$	diatom half saturation constant for NO_3 uptake	2.5	mmol N /m ³
$k_{NH_4}^{sp}$	Small phyto. half saturation constant for NH_4 uptake	0.01	mmol N /m ³
$k_{NH_4}^{diat}$	Diatom half saturation constant for NH_4 uptake	0.1	mmol N /m ³
k_{Fe}^{sp}	Small phyto. half saturation constant for Fe uptake	0.035e-3	mmol Fe /m ³
k_{Fe}^{diat}	Diat half saturation constant for Fe uptake	0.08e-3	mmol Fe /m ³
k_{Fe}^{diaz}	Diazotroph half saturation constant for Fe uptake	0.1e-3	mmol Fe /m ³
$k_{PO_4}^{sp}$	Small phyto. half saturation constant for PO_4 uptake	0.01	mmol P /m ³
$k_{PO_4}^{diat}$	Diatom half saturation constant for PO_4 uptake	0.1	mmol P /m ³
$k_{PO_4}^{diaz}$	Diazotroph half saturation constant for PO_4 uptake	0.005	mmol P /m ³

$k_{SiO_2}^{diat}$	Diatom half saturation constant for SiO_2 uptake	1	$mmol SiO_2 / m^3$
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Fixed Stoichiometry

$Q_{N:C}$	Nitrogen to Carbon ratio	0.137	$mmol N / mmol C$
$Q_{P:C}$	Small phyto. and diatom P:C ratio	0.00855	$mmol P / mmol C$
$Q_{P:C}^{diaz}$	Diazotroph P:C ratio	0.002735	$mmol P / mmol C$

Dissolved organic matter

τ_{dom}	Dissolved organic matter remineralization inverse timescale	0.01	1/d
$\tau_{remin, sed}$	remineralization inverse timescale in sediment	0.003	1/d

N₂ fixation

α_{ex}^{diaz}	Ratio of N excreted by diazotrophs to total N fixed	0.3	no units
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Chl stoichiometry and production

$Q_{N:chl,max}^{sp,growth}$	Max ratio of Chl produced to N photosynthesized for small phyto.	2.5	$mg Chl / mmol N$
$Q_{N:chl,max}^{diat,growth}$	Max ratio of Chl produced to N photosynthesized for diatom	4.0	$mg Chl / mmol N$
$Q_{N:chl,max}^{diaz,growth}$	Max ratio of Chl produced to N photosynthesized for diaz.	3.4	$mg Chl / mmol N$

Light

f_{Isw}	Fraction of incoming radiation used for photosynthesis	0.45	no units
μ_w	Radiation attenuation coefficient per unit chlorophyll	0.03	1/m/(mg <i>Chl</i> /m ³)
μ_{Chl}	Radiation attenuation coefficient for water	0.04	1/m

***Fe* stoichiometry and cycling**

$Q_{Fe:C}^{sp,g0}$	Ratio used in the calculation of the <i>Fe</i> to <i>C</i> ratio of uptake for small phyto.	6e-6	mmol <i>Fe</i> /mmol <i>C</i>
$Q_{Fe:C}^{sp,g1}$	Maximum <i>Fe</i> : <i>C</i> ratio of uptake by small phyto. if $Fe < 2 \cdot k_{Fe}^{sp}$.	2.5e-6	mmol <i>Fe</i> /mmol <i>C</i>
$Q_{Fe:C}^{diat,g0}$	Ratio used in the calculation of the <i>Fe</i> to <i>C</i> ratio of uptake for diatoms	6e-6	mmol <i>Fe</i> /mmol <i>C</i>
$Q_{Fe:C}^{diat,g1}$	Maximum <i>Fe</i> : <i>C</i> ratio of uptake by diatoms if $Fe < 2 \cdot k_{Fe}^{diat}$	2.5e-6	mmol <i>Fe</i> /mmol <i>C</i>
$Q_{Fe:C}^{diaz,g0}$	Ratio used in the calculation of the <i>Fe</i> to <i>C</i> ratio of uptake for diazotrophs	42e-6	mmol <i>Fe</i> /mmol <i>C</i>
$Q_{Fe:C}^{diaz,g1}$	Maximum <i>Fe</i> : <i>C</i> ratio of uptake by diazotrophs if $Fe < 2 \cdot k_{Fe}^{diaz}$	14e-6	mmol <i>Fe</i> /mmol <i>C</i>
$\beta_{fe}^{max,scal1}$	Scaling parameter used in the calculation of <i>Fe</i> scavenging	3.0	no units
$\beta_{fe}^{thresh,1}$	<i>Fe</i> threshold parameter used in the calculation of <i>Fe</i> scavenging	0.6e-3	mmol <i>Fe</i> /m ³

$\beta_{Fe}^{thresh,2}$	<i>Fe</i> threshold parameter used in the calculation of <i>Fe</i> scavenging	0.5e-3	mmol <i>Fe</i> /m ³
$J_{Fe}^{scav,0}$	<i>Fe</i> scavenging reference rate	0.12	mmol <i>Fe</i> /m ³ /s
<hr/> <i>CaCO</i>₃ stoichiometry and production			
$Q_{CaCO_3:C}^{sp,max}$	Maximum calcification to <i>C</i> photosynthesis ratio	0.4	mmol <i>CaCO</i> ₃ /mmol <i>C</i>
$f_{CaCO_3}^{prod}$	Initial calcification to <i>C</i> photosynthesis ratio	0.026	mmol <i>CaCO</i> ₃ /mmol <i>C</i>
$T_1^{CaCO_3}$	Temperature parameter used to modify the initial calcification rate	1.	°C
$T_2^{CaCO_3}$	Temperature parameter used to modify the initial calcification rate	-2.	°C
<hr/> <i>Si</i> stoichiometry			
$Q_{Si:C}^{diat,max}$	Maximum $Q_{Si:C}$ ratio for diatoms	0.685	mmol <i>Si</i> /mmol <i>C</i>
$Q_{Si:C}^{diat,g0}$	Default Si:C ratio of growth for diatoms	0.137	mmol <i>Si</i> /mmol <i>C</i>
$Q_{Si:C}^{diat,g1}$	Ratio used to calculate the <i>Si</i> to <i>C</i> multiplicative ratio of growth for diatoms	2.5	no units
<hr/> Grazing			
$J_{sp}^{g,max}$	maximum grazing loss for small phytoplankton	2.5	1/d
$J_{diat}^{g,max}$	maximum grazing loss for diatoms	1.95	1/d
$J_{diaz}^{g,max}$	maximum grazing loss for diazotrophs	1.2	1/d

β_z^{grz}	grazing coefficient, used in density dependent grazing modification	1.05	mmol C/m ³
$\beta_{thres}^{grz,sp,0}$	Small phytoplankton threshold concentration for grazing	0.001	mmol C/m ³
$\beta_{thres}^{grz,diat,0}$	Diatom threshold concentration for grazing	0.02	mmol C/m ³
$\beta_{thres}^{grz,diaz,0}$	Diazotroph threshold concentration for grazing	0.01	mmol C/m ³
$\alpha_{sp}^{grz,zoo}$	Fraction of small phyto. grazing going to zooplankton	0.3	no units
$\alpha_{diat}^{grz,zoo}$	Fraction of diatom grazing going to zooplankton	0.3	no units
$\alpha_{diaz}^{grz,zoo}$	Fraction of diazotroph grazing going to zooplankton	0.21	no units
$\alpha_{sp}^{grz,poc}$	Default fraction of small phytop. grazing going to POC	0.22	no units
$\alpha_{diat}^{grz,poc}$	Fraction of diatom grazing going to POC	0.26	no units
$\alpha_{diaz}^{grz,poc}$	Fraction of diazotroph grazing going to POC	0.0	no units
$\alpha_{sp}^{grz,doc}$	Fraction of small phyto. grazing going to DOC	0.34	no units
$\alpha_{diat}^{grz,doc}$	Fraction of diatom grazing going to DOC	0.13	no units
$\alpha_{diaz}^{grz,doc}$	Fraction of diazotroph grazing going to DOC	0.24	no units
$\alpha_{sp}^{grz,dic}$	Fraction of small phytop. grazing going to DIC	0.36	no units
$\alpha_{diat}^{grz,dic}$	Fraction of diatom grazing going to DIC	0.31	no units
$\alpha_{diaz}^{grz,dic}$	Fraction of diazotroph grazing going to DIC	0.55	no units

Losses

λ_{sp}^{mort}	Small phyto. mortality 1	0.15	1/d
λ_{diat}^{mort}	Diatom mortality 1	0.15	1/d
λ_{diaz}^{mort}	Diazotroph mortality 1	0.16	1/d
$\alpha_{diat}^{l,poc}$	Fraction of diatom loss going to POC	0.05	no units
$\alpha_{diaz}^{l,poc}$	Fraction of diazotroph loss going to POC	0	no units
f_{labile}	Fraction of Labile dis- solved organic matter for loss calculations	0.70	no units
λ_{zoo}^{mort}	Zooplankton linear mortality	0.08	1/d
λ_{zoo}^{mort2}	Zooplankton quadratic mortality	0.42	1/(mmol C m ³ d)
$\beta_{zoo}^{thres0,l}$	Zooplankton threshold concentrations for mor- tality	0.03	mmol C/m ³

Aggregation

$\tau_{diat}^{agg,min}$	Minimum aggregation rate for diatoms	0.01	1/d
$\tau_{diaz}^{agg,min}$	Minimum aggregation rate for diazotrophs	0.01	1/d
$\tau_{sp}^{agg,max}$	Maximum aggregation rate for small phyto.	0.75	1/d
$\tau_{diat}^{agg,max}$	Maximum aggregation rate for diatoms	0.75	1/d
$\tau_{diaz}^{agg,max}$	Maximum aggregation rate for diazotrophs	0.75	1/d
λ_{sp}^{mort2}	Small phyto. quadratic mortality	0.0035	1/(mmol C m ³ d)
λ_{diat}^{mort2}	diatom quadratic mor- tality	0.0035	1/(mmol C m ³ d)
λ_{diaz}^{mort2}	diazotroph quadratic mortality	0.16	1/(mmol C m ³ d)

N cycle rates

τ^{ammox}	NH_4 oxidation inverse timescale	0.06	1/d
τ^{nit}	NO_2 oxidation inverse timescale	0.33	1/d

PAR_{max}^{nitrif}	Light threshold for onset of nitrification	4.0	W / m ²
$Q_{N:C}^{denit}$	Ratio of inorganic <i>N</i> consumed to POC remineralized during denitrification	104/106	mmol <i>N</i> / mmol <i>C</i>
f_{n2o}^{denit}	Fraction of denitrification that goes to <i>N</i> ₂ <i>O</i>	0.95	no units
τ_{n2o}^{cons}	<i>N</i> ₂ <i>O</i> reduction inverse timescale	0.33333	1/d
a_0^{bohl}	Parameter used to calculate sedimentary denitrification (Bohlen et al., 2012)	0.06	no units
a_1^{bohl}	Parameter used to calculate sedimentary denitrification (Bohlen et al., 2012)	0.19	no units
a_2^{bohl}	Parameter used to calculate sedimentary denitrification (Bohlen et al., 2012)	0.99	no units

Particle cycling

ρ_{pic}	Organic carbon to inorganic carbon mass ratio in PIC	$0.07 \cdot \frac{100.09}{12.01}$	no units
ρ_{psi}	Organic carbon to inorganic Si mass ratio in <i>PSiO</i> ₂	$0.035 \cdot \frac{60.08}{12.01}$	mmol <i>SiO</i> ₂ / mmol <i>C</i>
ρ_{dust}	Organic carbon to dust mass ratio in dust	$0.07 \cdot \frac{10^6}{12.01}$	no units
λ_{poc}	remin. length scale for <i>poc</i> , modified by T	130	meters
λ_{pic}	remin. length scale for <i>pic</i>	600	meters
λ_{psi}	remin. length scale for <i>psi</i> , modified by T	210	meters
λ_{dust}	remin. length scale for <i>dust</i>	600	meters
λ_{nonsol}	remin. length scale for non soluble material	40000	meters
$Q_{10,poc}$	Temperature dependency factor for <i>poc</i> remin. length scale	1.12	no units

$Q_{10,psi}$	Temperature depen- dency factor for psi remin. length scale	4.0	no units
γ_{pic}	Production fraction routed to the nonsolu- ble subclass	0.55	no units
γ_{psi}	Production fraction routed to the nonsolu- ble subclass	0.37	no units
Φ_{poc}^{ref}	particulate C flux of reference	2.0e-3	mmolC/m ² /sec

Oxygen

O_2^{min}	Oxygen minimum threshold	1.0	mmol O_2 /m ³
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714 **5.3 Model equations**

715 **5.3.1 Tracer equations**

716 Here the symbol d/dt denotes the sum of the local time derivative and the physical transport.

$$\frac{d}{dt}(N_{no3}) = J^{nit} - J_{no3,no2}^{denit} - J_{sed}^{denit} - (J_{diat,no3}^u + J_{sp,no3}^u) \quad (A1)$$

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$$\begin{aligned} \frac{d}{dt}(N_{nh4}) = & Q_{N:C} J_{poc}^{remin} + J_{don,nh4} + Q_{N:C} (J_{diat,dic}^l + J_{sp,dic}^l \\ & + J_{diaz,dic}^l) + Q_{N:C} J_{zoo,dic}^l + Q_{N:C} (J_{diat,dic}^{grz} + J_{sp,dic}^{grz} + J_{diaz,dic}^{grz}) \\ & - J^{ammox} - (J_{diat,nh4}^u + J_{sp,nh4}^u) + Q_{N:C} \Phi_{poc}^{remin, sed} / \Delta z \end{aligned} \quad (A2)$$

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$$\frac{d}{dt}(N_{no2}) = J^{ammox} - J^{nit} - J_{no2,n2o}^{denit} + J_{no3,no2}^{denit} \quad (A3)$$

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$$\frac{d}{dt}(N_{n2o}) = 0.5 J_{no2,n2o}^{denit} - J_{n2o,n2}^{denit} + \Phi_{N_2O}^{air} / \Delta z \quad (A4)$$

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$$\frac{d}{dt}(N_{n2}) = J_{n2o,n2}^{denit} + 0.5 J_{sed}^{denit} + \Phi_{N_2}^{air} / \Delta z \quad (A5)$$

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$$\begin{aligned}
\frac{d}{dt}(Fe) = & J_{dofe,fe} - J_{Fe}^{scav} + (Q_{Fe:C}^{zoo} J_{zoo,dic}^l) + J_{pFe}^{remin} - J_{diaz,Fe}^u \\
& + Q_{Fe:C}^{sp}(J_{sp,dic}^l + J_{sp,dic}^{grz}) + Q_{Fe:C}^{diat}(J_{diat,dic}^l + J_{diat,dic}^{grz}) \\
& + Q_{Fe:C}^{diaz}(J_{diaz,dic}^l + J_{diaz,dic}^{grz}) - J_{sp,Fe}^u - J_{diat,Fe}^u \\
& + J_{sp,zoo}^{grz}(Q_{Fe:C}^{sp} - Q_{Fe:C}^{zoo}) + J_{diat,zoo}^{grz}(Q_{Fe:C}^{diat} - Q_{Fe:C}^{zoo}) \\
& + J_{diaz,zoo}^{grz}(Q_{Fe:C}^{diaz} - Q_{Fe:C}^{zoo}) + J_{Fe}^{Sed}
\end{aligned} \tag{A6}$$

727

728

$$\begin{aligned}
\frac{d}{dt}(P_{po4}) = & Q_{P:C}(J_{poc}^{remin} + J_{zoo,dic}^l + J_{sp,dic}^l + J_{diat,dic}^l + J_{sp,dic}^{grz} + J_{diat,dic}^{grz}) \\
& + J_{diaz,dip}^{loss} - (J_{sp,P}^u + J_{diat,P}^u + J_{diaz,P}^u) + J_{dop,po4} + Q_{P:C} \Phi_{poc}^{remin, sed} / \Delta z
\end{aligned} \tag{A7}$$

729

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$$\begin{aligned}
\frac{d}{dt}(Si_{sio2}) = & Q_{Si:C}(0.5 J_{diat}^{grz} + 0.95 J_{diat}^l) \\
& - J_{diat,Si}^u + J_{pSi}^{remin} + \Phi_{Si}^{remin, sed} / \Delta z
\end{aligned} \tag{A8}$$

731

732

733 If ($O_2 > O_2min$)

$$\begin{aligned}
\frac{d}{dt}(O_2) = & (J_{sp,C}^{photo} + J_{diat,C}^{photo} + J_{diaz,C}^{photo}) / R_{C:O}^d + ((-J_{poc}^{remin} - J_{doc,dic} - J_{zoo,dic}^l - J_{sp,dic}^l \\
& - J_{sp,dic}^{grz} - J_{diat,dic}^l - J_{diat,dic}^{grz} - J_{diaz,dic}^l - J_{diaz,dic}^{grz}) / R_{C:O}^d) \\
& - 1.5 J_{ammox} - 0.5 J_{nit} - \Phi_{poc}^{remin, sed} / (R_{C:O}^d \Delta z) + \Phi_{O_2}^{air} / \Delta z
\end{aligned} \tag{A9}$$

734

735 If ($O_2 \leq O_2min$)

$$\frac{d}{dt}(O_2) = (J_{sp,C}^{photo} + J_{diat,C}^{photo} + J_{diaz,C}^{photo}) / R_{C:O}^d + \Phi_{O_2}^{air} / \Delta z \tag{A10}$$

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$$\begin{aligned}
\frac{d}{dt}(DIC) = & J_{doc,dic} + 0.33 J_{sp}^{grz} Q_{C:caco3} \\
& + J_{zoo,dic}^l + J_{sp,dic}^l + J_{sp,dic}^{grz} + J_{diat,dic}^l + J_{diat,dic}^{grz} - J_{sp,C}^{photo} - J_{diat,C}^{photo} \\
& - J_{sp,caco3}^{prod} + J_{diaz,dic}^{grz} + J_{diaz,dic}^l - J_{diaz,C}^{photo} + (J_{poc}^{remin} + J_{pic}^{remin}) \\
& + (\Phi_{poc}^{remin, sed} + \Phi_{pic}^{remin, sed}) / \Delta z + \Phi_{CO_2}^{air} / \Delta z
\end{aligned} \tag{A11}$$

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$$\begin{aligned} \frac{d}{dt}(Alk) = & -\frac{dN_{no3}}{dt} + \frac{dN_{nh4}}{dt} + 2J_{pic}^{remin} + 0.33 J_{sp}^{grz} Q_{C:caco3} - J_{sp,caco3}^{prod} \\ & + 2\Phi_{pic}^{remin, sed} / \Delta z \end{aligned} \quad (A12)$$

740

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$$\frac{d}{dt}(C_{sp}) = J_{sp,C}^{photo} - (J_{sp}^{grz} + J_{sp}^l + J_{sp}^{agg}) \quad (A13)$$

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$$\frac{d}{dt}(Chl_{sp}) = J_{sp,chl}^{pa} - Q_{Chl:C}^{sp}(J_{sp}^{grz} + J_{sp}^l + J_{sp}^{agg}) \quad (A14)$$

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$$\frac{d}{dt}(Fe_{sp}) = gQ_{Fe:C}^{sp} J_{sp,C}^{photo} - Q_{Fe:C}^{sp}(J_{sp}^{grz} + J_{sp}^l + J_{sp}^{agg}) \quad (A15)$$

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$$\frac{d}{dt}(Ca_{sp}) = J_{sp,caco3}^{prod} - Q_{caco3:C}^{sp}(J_{sp}^{grz} + J_{sp}^l + J_{sp}^{agg}) \quad (A16)$$

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$$\frac{d}{dt}(C_{diat}) = J_{diat,C}^{photo} - (J_{diat}^{grz} + J_{diat}^l + J_{diat}^{agg}) \quad (A17)$$

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$$\frac{d}{dt}(Fe_{diat}) = gQ_{Fe:C}^{diat} J_{diat,C}^{photo} - Q_{Fe:C}^{diat}(J_{diat}^{grz} + J_{diat}^l + J_{diat}^{agg}) \quad (A18)$$

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$$\frac{d}{dt}(Chl_{diat}) = J_{diat,chl}^{pa} - Q_{Chl:C}^{diat}(J_{diat}^{grz} + J_{diat}^l + J_{diat}^{agg}) \quad (A19)$$

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$$\frac{d}{dt}(Si_{diat}) = gQ_{Si:C}^{diat} J_{diat,C}^{photo} - Q_{Si:C}^{diat}(J_{diat}^{grz} + J_{diat}^l + J_{diat}^{agg}) \quad (A20)$$

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$$\frac{d}{dt}(C_{diaz}) = J_{diaz,C}^{photo} - (J_{diaz}^{grz} + J_{diaz}^l + J_{diaz}^{agg}) \quad (A21)$$

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759

$$\frac{d}{dt}(Fe_{diaz}) = gQ_{Fe:C}^{diaz} J_{diaz,C}^{photo} - Q_{Fe:C}^{diaz} (J_{diaz}^{grz} + J_{diaz}^l + J_{diaz}^{agg}) \quad (A22)$$

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761

$$\frac{d}{dt}(Chl_{diaz}) = J_{diaz,chl}^{pa} - Q_{Chl:C}^{diaz} (J_{diaz}^{grz} + J_{diaz}^l + J_{diaz}^{agg}) \quad (A23)$$

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$$\frac{d}{dt}(C_{zoo}) = (J_{sp,zoo}^{grz} + J_{diat,zoo}^{grz} + J_{diaz,zoo}^{grz}) - J_{zoo}^l \quad (A24)$$

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765

$$\begin{aligned} \frac{d}{dt}(C_{doc}) = & J_{sp,doc}^l + J_{diat,doc}^l + J_{diaz,doc}^l + J_{zoo,doc}^l \\ & + (J_{sp,doc}^{grz} + J_{diat,doc}^{grz} + J_{diaz,doc}^{grz}) - J_{doc,dic} \end{aligned} \quad (A25)$$

766
767

$$\begin{aligned} \frac{d}{dt}(N_{don}) = & Q_{N:C} (J_{sp,doc}^l + J_{diat,doc}^l + J_{diaz,doc}^l + J_{zoo,doc}^l \\ & + J_{sp,doc}^{grz} + J_{diat,doc}^{grz} + J_{diaz,doc}^{grz}) - J_{don,nh4} + J_{diaz,N}^{excrete} \end{aligned} \quad (A26)$$

768
769

$$\begin{aligned} \frac{d}{dt}(P_{dop}) = & Q_{P:C} (J_{sp,doc}^l + J_{diat,doc}^l + J_{sp,doc}^{grz} + J_{diat,doc}^{grz} + J_{zoo,doc}^l) \\ & + J_{diaz,dop}^l - J_{dop,po4} \end{aligned} \quad (A27)$$

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771

$$\begin{aligned} \frac{d}{dt}(Fe_{dofe}) = & Q_{Fe:C}^{sp} (J_{sp,doc}^l + J_{sp,doc}^{grz}) + Q_{Fe:C}^{diat} (J_{diat,doc}^l + J_{diat,doc}^{grz}) + \\ & Q_{Fe:C}^{diaz} (J_{diaz,doc}^l + J_{diaz,doc}^{grz}) + Q_{Fe:C}^{zoo} J_{zoo,doc}^l - J_{dofe,fe} \end{aligned} \quad (A28)$$

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774 5.3.2 Treatment of particulate organic matter

775 General model

776 Particulate organic matter is produced and instantaneously distributed over the depth of the water
 777 column following the exponential solution to the steady-state 1-dimensional production-remineralization
 778 equation:

$$\frac{\partial \Phi(z)}{\partial z} = J^{remin}(z) - J^{prod}(z) = -\frac{\Phi(z)}{\lambda} + J^{prod}(z), \quad (\text{A29})$$

779 where Φ is a flux, λ is the remineralization length-scale, J^{prod} is the known production rate within
 780 the layer, and J^{remin} the remineralization rate within the layer, which needs to be determined. For
 781 a single layer, assuming the flux at the top of the layer $\Phi(k)$ is known, and the production $J^{prod}(k)$
 782 is constant within the layer, the solution to equation A29, can be cast to determine the flux out of
 783 the layer, $\Phi(k-1)$, for each element i :

$$\Phi_i(k-1) = \Phi_i(k) e^{-\frac{\Delta z}{\lambda_i}} + J^{prod}(k) (1 - e^{-\frac{\Delta z}{\lambda_i}}) \lambda, \quad (\text{A30})$$

784 Particulate organic carbon (POC) is partitioned between a free and mineral associated compo-
 785 nent:

$$\Phi_{poc}(z) = \Phi_{poc}^{free}(z) + \Phi_{poc}^{mineral}(z) \quad (\text{A31})$$

786 $\Phi_{poc}^{mineral}$ can be associated with CaCO_3 , SiO_2 or dust. Each mineral-associated POC flux is
 787 further partitioned into a “soluble” component, which remineralizes with the length-scale of the
 788 associated mineral, and a “non-soluble” component which remineralizes with a length-scale of
 789 40,000 m.

$$\begin{aligned} \Phi_{poc}^{mineral}(z) = & \rho_{pic} (\Phi_{pic}^{sol}(z) + \Phi_{pic}^{nonsol}(z)) \\ & + \rho_{psi} (\Phi_{psi}^{sol}(z) + \Phi_{psi}^{nonsol}(z)) \\ & + \rho_{dust} (\Phi_{dust}^{sol}(z) + \Phi_{dust}^{nonsol}(z)) \end{aligned} \quad (\text{A32})$$

790 For all components of the fluxes except Fe, the flux out of the layer, $\Phi(k-1)$, is computed first,
 791 with knowledge of the source within the layer, $J^{prod}(k)$, and of the remineralization length-scale.
 792 Remineralization in the layer is then calculated from conservation, *i.e.*, from A29. Below, for each
 793 component, we list the equations used to determine the production terms $J^{prod}(k)$, followed by
 794 the fluxes out, $\Phi(k-1)$, and finally the remineralization terms, $J^{remin}(z)$, which enter the tracer
 795 conservation equations.

796 Production

$$\begin{aligned} J_{poc}^{prod} = & (J_{sp,poc}^{grz} + J_{diat,poc}^{grz} + J_{diaz,poc}^{grz}) + (J_{sp}^{agg} + J_{diat}^{agg}) \\ & + (J_{sp,poc}^l + J_{diat,poc}^l + J_{diaz,poc}^l) + J_{zoo,poc}^l \end{aligned} \quad (\text{A33})$$

$$J_{pic}^{prod} = (0.67 J_{sp}^{grz} + J_{sp}^{agg} + J_{sp}^l) Q_{caco3:C}^{sp} \quad (A34)$$

$$J_{psi}^{prod} = (0.5 J_{diat}^{grz} + 0.05 J_{diat}^l + J_{diat}^{agg}) Q_{Si:C}^{diat} \quad (A35)$$

797

$$J_{dust}^{prod} = 0 \quad (A36)$$

798 Available production for free POC is then:

$$J_{poc,avail}^{prod} = J_{poc}^{prod} - \rho_{pic} J_{pic}^{prod} - \rho_{psi} J_{psi}^{prod} \quad (A37)$$

799 Fluxes out

800 Temperature dependency is used to modify the remineralization length scales of particulate organic
801 carbon, POC, and opal, SiO₂:

$$T_{poc} = (Q_{10,poc})^{\frac{T-30}{10}} \quad (A38)$$

802

$$T_{psi} = (Q_{10,psi})^{\frac{T-30}{10}} \quad (A39)$$

803 Free POC flux equation:

$$\Phi_{poc}^{free}(k-1) = \Phi(k)_{poc}^{free} e^{-\frac{\Delta z T_{poc}}{\lambda_{poc}}} + J_{poc,avail}^{prod}(k) (1 - e^{-\frac{\Delta z T_{poc}}{\lambda_{poc}}}) \frac{\lambda_{poc}}{T_{poc}} \quad (A40)$$

804 Soluble mineral-associated POC flux equation:

$$\Phi_{pic}^{sol}(k-1) = \Phi(k)_{pic}^{sol} e^{-\frac{\Delta z}{\lambda_{pic}}} + J_{pic}^{prod}(k) (1 - \gamma_{pic}) (1 - e^{-\frac{\Delta z}{\lambda_{pic}}}) \lambda_{pic} \quad (A41)$$

$$\Phi_{psi}^{sol}(k-1) = \Phi(k)_{psi}^{sol} e^{-\frac{\Delta z T_{psi}}{\lambda_{psi}}} + J_{psi}^{prod}(k) (1 - \gamma_{psi}) (1 - e^{-\frac{\Delta z T_{psi}}{\lambda_{psi}}}) \frac{\lambda_{psi}}{T_{psi}} \quad (A42)$$

$$\Phi_{dust}^{sol}(k-1) = \Phi(k)_{dust}^{sol} e^{-\frac{\Delta z}{\lambda_{dust}}} \quad (A43)$$

805 Non-soluble mineral-associated POC flux equation:

$$\Phi_{pic}^{nonsol}(k-1) = \Phi(k)_{pic}^{nonsol} e^{-\frac{\Delta z}{\lambda_{nonsol}}} + J_{pic}^{prod}(k) \gamma_{pic} \Delta z \quad (A44)$$

$$\Phi_{psi}^{nonsol}(k-1) = \Phi(k)_{psi}^{nonsol} e^{-\frac{\Delta z}{\lambda_{nonsol}}} + J_{psi}^{prod}(k) \gamma_{psi} \Delta z \quad (A45)$$

$$\Phi_{dust}^{nonsol}(k-1) = \Phi(k)_{dust}^{nonsol} e^{-\frac{\Delta z}{\lambda_{nonsol}}} \quad (A46)$$

806 **Remineralization**

807 Remineralization is computed from conservation, i.e, A29:

$$J^{remmin} = J^{prod} + \frac{\partial \Phi(z)}{\partial z} \quad (\text{A47})$$

808 Numerically, for each individual layer, we have:

$$J_{poc}^{remmin}(k) = J_{poc}^{prod}(k) + \frac{\Phi_{poc}^{free}(k) - \Phi_{poc}^{free}(k-1) + \Phi_{poc}^{mineral}(k) - \Phi_{poc}^{mineral}(k-1)}{\Delta z} \quad (\text{A48})$$

809

$$J_{pic}^{remmin}(k) = J_{pic}^{prod}(k) + \frac{\Phi_{pic}^{sol}(k) - \Phi_{pic}^{sol}(k-1) + \Phi_{pic}^{nonsol}(k) - \Phi_{pic}^{nonsol}(k-1)}{\Delta z} \quad (\text{A49})$$

810

$$J_{psi}^{remmin}(k) = J_{psi}^{prod}(k) + \frac{\Phi_{psi}^{sol}(k) - \Phi_{psi}^{sol}(k-1) + \Phi_{psi}^{nonsol}(k) - \Phi_{psi}^{nonsol}(k-1)}{\Delta z} \quad (\text{A50})$$

811

$$J_{dust}^{remmin}(k) = \frac{\Phi_{dust}^{sol}(k) - \Phi_{dust}^{sol}(k-1) + \Phi_{dust}^{nonsol}(k) - \Phi_{dust}^{nonsol}(k-1)}{\Delta z} \quad (\text{A51})$$

812 **Particulate Fe**

813 Production is as follows:

$$\begin{aligned} J_{pfe}^{prod} = & (J_{sp}^{agg} + J_{sp,poc}^{grz} + J_{sp,poc}^l) Q_{Fe:C}^{sp} \\ & + (J_{diat}^{agg} + J_{diat,poc}^{grz} + J_{diat,poc}^l) Q_{Fe:C}^{diat} \\ & + (J_{diaz}^{agg} + J_{diaz,poc}^{grz}) Q_{Fe:C}^{diaz} + 0.1 J_{Fe}^{scav} \\ & + J_{zoo}^l J_{zoo}^d Q_{Fe:C}^{zoo} \end{aligned} \quad (\text{A52})$$

814 Particulate Fe remineralization is assumed to be proportional to POC remineralization plus a
815 release from dust :

$$J_{pfe}^{remmin}(k) = J_{poc}^{remmin}(k) \frac{\Phi_{pfe}^{sol}(k) + \Phi_{pfe}^{nonsol}(k)}{\Phi_{poc}^{free}(k) + \Phi_{poc}^{mineral}(k)} + 626.712 J_{dust}^{remmin}(k) \quad (\text{A53})$$

816 The flux out can then be computed from conservation:

$$\Phi_{pfe}^{sol}(k-1) = \Phi_{pfe}^{sol}(k) + \Delta z \left(J_{pfe}^{prod}(k) - J_{poc}^{remmin}(k) \frac{\Phi_{pfe}^{sol}(k) + \Phi_{pfe}^{nonsol}(k)}{\Phi_{poc}^{free}(k) + \Phi_{poc}^{mineral}(k)} \right) \quad (\text{A54})$$

$$\Phi_{pfe}^{nonsol}(k-1) = \Phi_{pfe}^{nonsol}(k) \quad (\text{A55})$$

817 **Bottom flux to sediment**

$$\Phi_{poc}^{out} = \Phi_{poc}(k_{bottom}) \quad (\text{A56})$$

818 5.3.3 Biogeochemical rates

819 Carbon

$$T_{func} = 2^{0.1 * T - 3} \quad (A57)$$

$$J_{sp,C}^{photo} = PC_{ref}^{sp} f_{nut}^{sp} T_{func} \left(1 - e^{-\frac{\alpha_{chl} Q_{chl:C}^{sp} PAR}{PC_{ref}^{sp} f_{nut}^{sp} T_{func}}} \right) C_{sp} \quad (A58)$$

$$J_{diat,C}^{photo} = PC_{ref}^{diat} f_{nut}^{diat} T_{func} \left(1 - e^{-\frac{\alpha_{chl} Q_{chl:C}^{diat} PAR}{PC_{ref}^{diat} f_{nut}^{diat} T_{func}}} \right) C_{diat} \quad (A59)$$

$$J_{diaz,C}^{photo} = PC_{ref}^{diaz} f_{nut}^{diaz} T_{func} \left(1 - e^{-\frac{\alpha_{chl} Q_{chl:C}^{diaz} PAR}{PC_{ref}^{diaz} f_{nut}^{diaz} T_{func}}} \right) C_{diaz} \quad (A60)$$

823 Remineralization

$$J_{don,nh4} = \tau_{dom} N_{don} \quad (A61)$$

$$J_{doc,dic} = \tau_{dom} C_{doc} \quad (A62)$$

$$J_{dop,po4} = \tau_{dom} P_{dop} \quad (A63)$$

$$J_{dofe,fe} = \tau_{dom} Fe_{dofe} \quad (A64)$$

$$\Phi_{poc}^{remin,sed} = \begin{cases} \tau_{remin,sed} POC_{sed}, & \text{if } k = k_{bottom} \\ 0 & \text{elsewhere} \end{cases} \quad (A65)$$

$$\Phi_{pic}^{remin,sed} = \begin{cases} \tau_{remin,sed} CaCO3_{sed}, & \text{if } k = k_{bottom} \\ 0 & \text{elsewhere} \end{cases} \quad (A66)$$

$$\Phi_{Si}^{remin,sed} = \begin{cases} \tau_{remin,sed} Si_{sed}, & \text{if } k = k_{bottom} \\ 0 & \text{elsewhere} \end{cases} \quad (A67)$$

830 Nutrient limitation

$$f_{nut}^{sp} = \min(V_{NO_3}^{sp} + V_{NH_4}^{sp}, V_{Fe}^{sp}, V_{PO_4}^{sp}) \quad (A68)$$

$$f_{nut}^{diat} = \min(V_{NO_3}^{diat} + V_{NH_4}^{diat}, V_{Fe}^{diat}, V_{SiO_3}^{diat}, V_{PO_4}^{diat}) \quad (A69)$$

$$f_{nut}^{diaz} = \min(V_{Fe}^{diaz}, V_{PO_4}^{diaz}) \quad (A70)$$

833

$$V_{NO_3}^{sp} = \frac{N_{NO_3}/k_{no3}^{sp}}{1 + N_{NO_3}/k_{no3}^{sp} + N_{NH_4}/k_{nh4}^{sp}} \quad (A71)$$

834

$$V_{NO_3}^{diat} = \frac{N_{NO_3}/k_{no3}^{diat}}{1 + N_{NO_3}/k_{no3}^{diat} + N_{NH_4}/k_{nh4}^{diat}} \quad (A72)$$

835

$$V_{NH_4}^{sp} = \frac{N_{NH_4}/k_{nh4}^{sp}}{1 + N_{NO_3}/k_{no3}^{sp} + N_{NH_4}/k_{nh4}^{sp}} \quad (A73)$$

836

$$V_{NH_4}^{diat} = \frac{N_{NH_4}/k_{nh4}^{diat}}{1 + N_{NO_3}/k_{no3}^{diat} + N_{NH_4}/k_{nh4}^{diat}} \quad (A74)$$

837

$$V_{Fe}^{sp} = \frac{Fe}{Fe + k_{fe}^{sp}} \quad (A75)$$

838

$$V_{Fe}^{diat} = \frac{Fe}{Fe + k_{fe}^{diat}} \quad (A76)$$

839

$$V_{Fe}^{diaz} = \frac{Fe}{Fe + k_{fe}^{diaz}} \quad (A77)$$

840

$$V_{PO_4}^{sp} = \frac{PO_4}{PO_4 + k_{po4}^{sp}} \quad (A78)$$

841

$$V_{PO_4}^{diat} = \frac{PO_4}{PO_4 + k_{po4}^{diat}} \quad (A79)$$

842

$$V_{PO_4}^{diaz} = \frac{PO_4}{PO_4 + k_{po4}^{diaz}} \quad (A80)$$

843

$$V_{SiO_2}^{diat} = \frac{Si_{sio3}}{Si_{sio2} + k_{sio2}^{diat}} \quad (A81)$$

844 NO_3 and NH_4 uptake

$$J_{sp,no3}^u = Q_{N:C} \frac{V_{NO_3}^{sp}}{V_{NO_3}^{sp} + V_{NH_4}^{sp}} J_{sp,C}^{photo} \quad (A82)$$

845

$$J_{sp,nh4}^u = Q_{N:C} \frac{V_{NH_4}^{sp}}{V_{NO_3}^{sp} + V_{NH_4}^{sp}} J_{sp,C}^{photo} \quad (A83)$$

846

$$J_{diat,no3}^u = Q_{N:C} \frac{V_{NO_3}^{diat}}{V_{NO_3}^{diat} + V_{NH_4}^{diat}} J_{diat,C}^{photo} \quad (A84)$$

847

$$J_{diat,nh4}^u = Q_{N:C} \frac{V_{NH_4}^{diat}}{V_{NO_3}^{diat} + V_{NH_4}^{diat}} J_{diat,C}^{photo} \quad (A85)$$

848 **N₂ fixation**

$$J_{diaz,N}^{photo} = Q_{N:C} J_{diaz,C}^{photo} \quad (A86)$$

849

$$J_{diaz}^{nfix} = J_{diaz,N}^{photo} / (1 - \alpha_{ex}^{diaz}) \quad (A87)$$

850

$$J_{diaz,N}^{excrete} = \alpha_{ex}^{diaz} J_{diaz}^{nfix} \quad (A88)$$

851 **PO₄ uptake**

$$J_{sp,po4}^u = Q_{P:C} J_{sp,C}^{photo} \quad (A89)$$

852

$$J_{diat,po4}^u = Q_{P:C} J_{diat,C}^{photo} \quad (A90)$$

853

$$J_{diaz,po4}^u = Q_{P:C}^{diaz} J_{diaz,C}^{photo} \quad (A91)$$

854 **Chl stoichiometry and production**

$$Q_{Chl:C}^{sp} = Chl_{sp} / C_{sp} \quad (A92)$$

855

$$Q_{Chl:C}^{diat} = Chl_{diat} / C_{diat} \quad (A93)$$

856

$$Q_{Chl:C}^{diaz} = Chl_{diaz} / C_{diaz} \quad (A94)$$

$$J_{sp,chl}^{pa} = Q_{N:chl,max}^{sp,growth} \frac{J_{sp,C}^{photo} / C_{sp}}{\alpha_{chl} Q_{Chl:C}^{sp} PAR} Q_{N:C} J_{sp,C}^{photo} \quad (A95)$$

$$J_{diat,chl}^{pa} = Q_{N:chl,max}^{diat,growth} \frac{J_{diat,C}^{photo} / C_{diat}}{\alpha_{chl} Q_{Chl:C}^{diat} PAR} Q_{N:C} J_{diat,C}^{photo} \quad (A96)$$

$$J_{diaz,chl}^{pa} = Q_{N:chl,max}^{diaz,growth} \frac{J_{diaz,C}^{photo} / C_{diaz}}{\alpha_{chl} Q_{Chl:C}^{diaz} PAR} Q_{N:C} J_{diaz,C}^{photo} \quad (A97)$$

857 **Light**

$$PAR_B(z) = \max(0, f_{I_{sw}} I_{SW}) e^{\int_z^0 (\mu_{chl}(Chl_{sp} + Chl_{diat} + Chl_{diaz}) + \mu_w) dz} \quad (A98)$$

858

$$PAR = \frac{1}{\mu_{chl}(Chl_{sp} + Chl_{diat} + Chl_{diaz}) + \mu_w} \frac{\partial PAR_B}{\partial z}(z) \quad (A99)$$

859 *Fe* stoichiometry and cycling

$$Q_{Fe:C}^{sp} = Fe_{sp}/C_{sp} \quad (A100)$$

860

$$Q_{Fe:C}^{diat} = Fe_{diat}/C_{diat} \quad (A101)$$

861

$$Q_{Fe:C}^{diaz} = Fe_{diaz}/C_{diaz} \quad (A102)$$

862

$$Q_{Fe:C}^{Zoo} = 2.5e^{-6} \quad (A103)$$

863

$$gQ_{Fe:C}^{sp} = \begin{cases} Q_{Fe:C}^{sp,g0}, & \text{if } Fe \geq 2k_{fe}^{sp} \\ \max(Q_{Fe:C}^{sp,g1}, \frac{Q_{Fe:C}^{sp,g0} Fe}{2k_{fe}^{sp}}), & \text{otherwise} \end{cases} \quad (A104)$$

864

$$gQ_{Fe:C}^{diat} = \begin{cases} Q_{Fe:C}^{diat,g0}, & \text{if } Fe \geq 2k_{fe}^{diat} \\ \max(Q_{Fe:C}^{diat,g1}, \frac{Q_{Fe:C}^{diat,g0} Fe}{2k_{fe}^{diat}}), & \text{otherwise} \end{cases} \quad (A105)$$

865

$$gQ_{Fe:C}^{diaz} = \begin{cases} Q_{Fe:C}^{diaz,g0}, & \text{if } Fe \geq 2k_{fe}^{diaz} \\ \max(Q_{Fe:C}^{diaz,g1}, \frac{Q_{Fe:C}^{diaz,g0} Fe}{2k_{fe}^{diaz}}), & \text{otherwise} \end{cases} \quad (A106)$$

$$J_{sp,Fe}^u = gQ_{Fe:C}^{sp} J_{sp,C}^{photo} \quad (A107)$$

866

$$J_{diat,Fe}^u = gQ_{Fe:C}^{diat} J_{diat,C}^{photo} \quad (A108)$$

867

$$J_{diaz,Fe}^u = gQ_{Fe:C}^{diaz} J_{diaz,C}^{photo} \quad (A109)$$

$$J_{Fe}^{scav} = \begin{cases} (J_{Fe}^{scav,0} \min((\Phi_{poc} + 8.33 \cdot 10^4 \Phi_{dust})/\Phi_{poc}^{ref}, \beta_{fe}^{max, scale1}) \\ + (Fe - \beta_{fe}^{thresh,1}) 6.0/1.4e^{-3}) 3.1709792e^{-8} Fe, & \text{if } Fe > \beta_{fe}^{thresh,1} \\ (J_{Fe}^{scav,0} \min((\Phi_{poc} + 8.33 \cdot 10^4 \Phi_{dust})/\Phi_{poc}^{ref}, \beta_{fe}^{max, scale1}) \\ + \frac{Fe}{\beta_{fe}^{thresh,2}}) 3.1709792e^{-8} Fe, & \text{if } Fe < \beta_{fe}^{thresh,2} \end{cases} \quad (A110)$$

$$J_{Fe}^{Sed} = \begin{cases} 0.75 \cdot 10^{(2.5-0.0165 O_2)} (0.001/86400)/\Delta z, & \text{if } k = k_{bottom} \text{ and } O_2 > O_2^{min} \\ 0.25 \cdot 10^{(2.5-0.0165 O_2)} (0.001/86400)/\Delta z, & \text{if } k = k_{bottom-1} \text{ and } O_2 > O_2^{min} \\ 10^{(2.5-0.0165 O_2)} (0.001/86400)/\Delta z & \text{elsewhere} \end{cases} \quad (A111)$$

868 75% of J_{Fe}^{Sed} is released in the bottom layer and 25 % in the layer above.

$$Q_{caco3:C}^{sp} = \begin{cases} CaCO_3_{sp}/C_{sp} & \text{if } CaCO_3_{sp}/C_{sp} \leq Q_{caco3:C}^{sp,max} \\ Q_{caco3:C}^{sp,max} & \text{otherwise} \end{cases} \quad (A112)$$

$$J_{sp,caco3}^{prod,max} = f_{caco3}^{prod} J_{sp,C}^{photo} f_{nut}^{sp} \quad (A113)$$

$$J_{sp,caco3}^{prodT} = \begin{cases} J_{sp,caco3}^{prodmax} & \text{if } T \geq T_1^{caco3} \\ J_{sp,caco3}^{prodmax} \frac{\max[T-T_2^{caco3}, 0]}{T_1^{caco3}-T_2^{caco3}} & \text{if } T < T_1^{caco3} \end{cases} \quad (A114)$$

$$J_{sp,caco3}^{prod} = \begin{cases} \min[(J_{sp,caco3}^{prodT} C_{sp}/3, 0.4 J_{sp,C}^{photo})] & \text{where } C_{sp} > 3.0 \\ J_{sp,caco3}^{prod,T} & \text{elsewhere} \end{cases} \quad (A115)$$

870 **Si stoichiometry**

$$Q_{Si:C}^{diat} = \max(Si_{diat}/C_{diat}, Q_{Si:C}^{diat,max}) \quad (A116)$$

$$gQ_{Si:C}^{diat} = \begin{cases} \min(Q_{Si:C}^{diat,g0} (2Q_{Si:C}^{diat,g1} k_{fe}^{diat} / Fe - Q_{Si:C}^{diat,g1} + 1), Q_{Si:C}^{diat,max}), & \text{if } 0 < Fe < 2k_{fe}^{diat} \text{ and } Si_{sio2} > 2k_{sio2}^{diat} \\ Q_{Si:C}^{diat,max} & \text{if } Fe = 0 \\ Q_{Si:C}^{diat,g0} & \text{otherwise} \end{cases} \quad (A117)$$

$$J_{diat,Si}^u = gQ_{Si:C}^{diat} J_{diat,C}^{photo} \quad (A118)$$

871 **Grazing**

$$\beta_{thres}^{grz,sp} = \begin{cases} \beta_{thres,0}^{grz,sp} & z \geq -100m \\ \beta_{thres,0}^{grz,sp} \cdot (200 + z)/100 & -100 m > z > -200 m \\ 0 & z \leq -200 m \end{cases} \quad (A119)$$

$$P'_{sp} = \max(C_{sp} - \beta_{thres}^{grz,sp}, 0) \quad (A120)$$

872

$$\beta_{thres}^{grz,diat} = \begin{cases} \beta_{thres,0}^{grz,diat} & z \geq -100m \\ \beta_{thres,0}^{grz,diat} \cdot (200 + z)/100 & -100 m > z > -200 m \\ 0 & z \leq -200 m \end{cases} \quad (A121)$$

$$P'_{diat} = \max(C_{diat} - \beta_{thres}^{grz,diat}, 0) \quad (A122)$$

$$\beta_{thres}^{grz,diaz} = \begin{cases} \beta_{thres,0}^{grz,diaz} & z \geq -100m \\ \beta_{thres,0}^{grz,diaz} \cdot (200 + z)/100 & -100 m > z > -200 m \\ 0 & z \leq -200 m \end{cases} \quad (A123)$$

$$P'_{diaz} = \max(C_{diaz} - \beta_{thres}^{grz,diaz}, 0) \quad (A124)$$

$$J_{sp}^{grz} = J_{sp}^{g,max} T_{func} \frac{P_{sp}'^2}{P_{sp}'^2 + (\beta_z^{grz})^2} C_{zoo} \quad (A125)$$

$$J_{diat}^{grz} = J_{diat}^{g,max} T_{func} \frac{P_{diat}'^2}{P_{diat}'^2 + 0.81(\beta_z^{grz})^2} C_{zoo} \quad (A126)$$

$$J_{diaz}^{grz} = J_{diaz}^{g,max} T_{func} \frac{P_{diaz}'^2}{P_{diaz}'^2 + (\beta_z^{grz})^2} C_{zoo} \quad (A127)$$

$$J_{sp}^{g,zoo} = \alpha_{sp}^{grz,zoo} J_{sp}^{grz} \quad (A128)$$

873

$$J_{diat}^{g,zoo} = \alpha_{diat}^{grz,zoo} J_{diat}^{grz} \quad (A129)$$

874

$$J_{diaz}^{g,zoo} = \alpha_{diaz}^{grz,zoo} J_{diaz}^{grz} \quad (A130)$$

$$J_{sp}^{g,poc} = \max(Q_{caco3:C}^{sp,max} Q_{caco3:C}^{sp}, \min(0.18 P'_{sp}, \alpha_{sp}^{grz,poc})) J_{sp}^{grz} \quad (A131)$$

875

$$J_{diat}^{g,poc} = \alpha_{diat}^{grz,poc} J_{diat}^{grz} \quad (A132)$$

876

$$J_{diaz}^{g,poc} = \alpha_{diaz}^{grz,poc} J_{diaz}^{grz} \quad (A133)$$

$$J_{sp}^{g,doc} = \alpha_{sp}^{grz,doc} J_{sp}^{grz} - J_{sp}^{grz,poc} \quad (A134)$$

877

$$J_{diat}^{g,doc} = \alpha_{diat}^{grz,doc} J_{diat}^{grz} \quad (A135)$$

878

$$J_{diaz}^{g,doc} = \alpha_{diaz}^{grz,doc} J_{diaz}^{grz} \quad (A136)$$

$$J_{sp}^{g,dic} = \alpha_{sp}^{grz,dic} J_{sp}^{grz} \quad (A137)$$

879

$$J_{diat}^{g,dic} = \alpha_{diat}^{grz,dic} J_{diat}^{grz} \quad (A138)$$

880

$$J_{diaz}^{g,dic} = \alpha_{diaz}^{grz,dic} J_{diaz}^{grz} \quad (A139)$$

881 **Losses**

$$J_{sp}^l = \lambda_{sp}^{mort} P'_{sp} \quad (A140)$$

882

$$J_{diat}^l = \lambda_{diat}^{mort} P'_{diat} \quad (A141)$$

883

$$J_{diaz}^l = \lambda_{diaz}^{mort} P'_{diaz} \quad (A142)$$

$$J_{sp}^{l,poc} = Q_{caco3:C}^{sp} J_{sp}^l \quad (A143)$$

884

$$J_{diat}^{l,poc} = \alpha_{diat}^{l,poc} J_{diat}^l \quad (A144)$$

885

$$J_{diaz}^{l,poc} = \alpha_{diaz}^{l,poc} J_{diaz}^l \quad (A145)$$

$$J_{sp}^{l,doc} = (1 - f_{labile})(J_{sp}^l - J_{sp}^{l,poc}) \quad (A146)$$

886

$$J_{diat}^{l,doc} = (1 - f_{labile})(J_{diat}^l - J_{diat}^{l,poc}) \quad (A147)$$

887

$$J_{diaz}^{l,doc} = (1 - f_{labile})(J_{diaz}^l - J_{diaz}^{l,poc}) \quad (A148)$$

$$J_{sp}^{l,dic} = f_{labile}(J_{sp}^l - J_{sp}^{l,poc}) \quad (A149)$$

888

$$J_{diat}^{l,dic} = f_{labile}(J_{diat}^l - J_{diat}^{l,poc}) \quad (A150)$$

889

$$J_{diaz}^{l,dic} = f_{labile}(J_{diaz}^l - J_{diaz}^{l,poc}) \quad (A151)$$

$$P_{diaz}^{remain} = Q_{P:C}^{diaz} (J_{diaz}^{grz} + J_{diaz}^l + J_{diaz}^{agg}) - Q_{P:C} (J_{diaz,poc}^{grz} + J_{diaz,zoo}^{grz}) \quad (A152)$$

890

$$J_{diaz,dop}^l = (1 - f_{labile}) P_{diaz}^{remain} \quad (A153)$$

891

$$J_{diaz,dip}^l = f_{labile} P_{diaz}^{remain} \quad (A154)$$

$$\beta_{zoo}^{thres,l} = \begin{cases} \beta_{zoo}^{thres0,l} & z \geq -100\text{m} \\ \beta_{zoo}^{thres0,l} \cdot (200 + z)/100 & -100 \text{ m} > z > -200 \text{ m} \\ 0 & z \leq -200 \text{ m} \end{cases} \quad (A155)$$

$$Z'_{zoo} = \max(C_{zoo} - \beta_{zoo}^{thres,l}, 0) \quad (A156)$$

$$J_{zoo}^l = \lambda_{zoo}^{mort2} T_{func} Z_{zoo}'^2 + \lambda_{zoo}^{mort} T_{func} Z_{zoo}' \quad (A157)$$

$$f_{zoo}^d = \frac{0.1333 J_{diat}^{grz} + 0.0333 J_{sp}^{grz}}{J_{diat}^{grz} + J_{sp}^{grz} + J_{diaz}^{grz}} \quad (A158)$$

$$J_{zoo,doc}^l = (1 - f_{labile}) (1 - f_{zoo}^d) J_{zoo}^l \quad (A159)$$

892

$$J_{zoo,dic}^l = f_{labile} (1 - f_{zoo}^d) J_{zoo}^l \quad (A160)$$

893

$$J_{zoo,poc}^l = f_{zoo}^d J_{zoo}^l \quad (A161)$$

894 Aggregation

$$J_{sp}^{agg} = \min(\tau_{sp}^{agg,max} P'_{sp}, \lambda_{sp}^{mort2} P_{sp}'^2) \quad (A162)$$

895

$$J_{diat}^{agg} = \max(\tau_{diat}^{agg,min} P'_{diat}, \min(\tau_{diat}^{agg,max} P'_{diat}, \lambda_{diat}^{mort2} P_{diat}'^2)) \quad (A163)$$

896

$$J_{diaz}^{agg} = 0.0 \quad (A164)$$

897 N-cycle rates

$$J_{ammox} = \begin{cases} \tau_{ammox} N_{nh4}, & \text{where } PAR_B(k) < PAR_{max}^{nitrif} \\ 0, & \text{elsewhere} \end{cases} \quad (A165)$$

$$J_{nit} = \begin{cases} \tau_{nit} N_{no2}, & \text{where } PAR_B(k) < PAR_{max}^{nitrif} \text{ and } O_2 > O_2^{min} \\ 0, & \text{elsewhere} \end{cases} \quad (A166)$$

$$\begin{aligned} J_{o2}^{cons} = & (J_{poc}^{remin} + J_{doc,dic} + J_{zoo,dic}^l + J_{sp,dic}^l \\ & + J_{sp,dic}^{grz} + J_{diat,dic}^l + J_{diat,dic}^{grz} \\ & + J_{diaz,dic}^l + J_{diaz,dic}^{grz} + J_{remin}^{poc}) / Q_{C:O} \\ & + 1.5 J_{ammox} + 0.5 J_{nit} \end{aligned} \quad (A167)$$

$$J_{no3,no2}^{denit} = \begin{cases} Q_{N:C}^{denit} J_{poc}^{remin}, & \text{where } O_2 \leq O_2^{min} \\ 0, & \text{elsewhere} \end{cases} \quad (A168)$$

$$J_{no2,n2o}^{denit} = \begin{cases} f_{n2o}^{denit} J_{no3,no2}^{denit}, & \text{where } O_2 \leq O_2^{min} \\ 0, & \text{elsewhere} \end{cases} \quad (A169)$$

898

$$J_{n2o,n2}^{denit} = \begin{cases} \tau_{n2o}^{cons} N_{n2o}, & \text{where } O_2 \leq O_2^{min} \\ 0, & \text{elsewhere} \end{cases} \quad (A170)$$

$$J_{sed}^{denit} = \begin{cases} (a_0^{bohl} + a_1^{bohl} (a_2^{bohl})^{O_2 - N_{no3}}) \frac{\Phi_{poc}^{out}}{\Delta z}, & \text{where } z = z_{bottom} \\ 0, & \text{elsewhere} \end{cases} \quad (\text{A171})$$

900 **Air-sea fluxes**

$$SC_{O_2} = 1638.0 + T_{surf}(-81.83 + T_{surf}(-0.008004(1.483 + T_{surf}))) \quad (\text{A172})$$

901

$$SC_{CO_2} = 2073.1 + T_{surf}(-125.62 + T_{surf}(-0.043219(3.6276 + T_{surf}))) \quad (\text{A173})$$

902

$$PV_{O_2} = 8.6 \cdot 10^{-7} W s^2 \sqrt{\frac{660}{SC_{CO_2}}} \quad (\text{A174})$$

903

$$PV_{CO_2} = 8.6 \cdot 10^{-7} W s^2 \sqrt{\frac{660}{SC_{CO_2}}} \quad (\text{A175})$$

904

$$\Phi_{O_2}^{air} = PV_{O_2}(O_{2sat} - O_2) \quad (\text{A176})$$

905 O_{2sat} is the oxygen saturation concentration computed from Garcia and Gordon (1992), page 1310,
906 eq. 8.

$$\Phi_{N_2O}^{air} = -PV_{O_2}N_2O \quad (\text{A177})$$

907

$$\Phi_{N_2}^{air} = -PV_{O_2}N_2 \quad (\text{A178})$$

908

$$\Phi_{CO_2}^{air} = PV_{CO_2}(CO_{2air}^* - CO_2^*) \quad (\text{A179})$$

909 with CO_{2air}^* and CO_2^* the concentrations corresponding to the partial pressure in air and water
910 computed from Dickson and Goyet (1994) (SOP No. 3, p25-26).

911 **Concentration in sediment**

$$POC_{sed} = \Phi_{poc}^{out} - J_{poc}^{remin, sed} \quad (\text{A180})$$

912

$$CaCO_3_{sed} = \Phi_{pic}^{out} - J_{pic}^{remin, sed} \quad (\text{A181})$$

913

$$Si_{sed} = \Phi_{Si}^{out} - J_{Si}^{remin, sed} \quad (\text{A182})$$

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