length as the ocean (0.01 yr), even in response to rather extreme (15 000 PgC) releases of (fossil fuel) CO_2 to the atmosphere.

2.1 Ocean biogeochemical cycling

The low vertical resolution of the ocean circulation model and need to maximize computational speed for long simulations and sensitivity analyses dictates that the "biological" part of the marine carbon cycle be relatively abstracted. A more important reason for starting off with as simple a description of surface biological productivity as possible is to reduce the number of free parameters in the model which must be optimized, particularly since this work represents the first attempt to assimilate marine geochemical observations within a prognostic 3-D circulation model of the ocean. Of course, as the nature of the assimilation problem becomes better defined and the capabilities and limitations of available optimization techniques better understood, we expect to subsequently re-calibrate the marine carbon cycle in conjunction with increasingly sophisticated multiple-nutrient biological schemes.

In this paper we estimate new (export) production directly from available surface nutrient concentrations, a robust tactic used in many early ocean carbon cycle models. In other words, what we have is "conceptually not a model of biology in the ocean but rather a model of biogenically induced chemical fluxes (from the surface ocean)" (Maier-Reimer, 1993). Overall, the biological export scheme is functionally similar to that of Parekh et al. (2005), and we adopt their notation where relevant. The main difference is that we currently consider only a single nutrient, phosphate (PO_4), rather than co-limitation with iron (Fe).

The governing equations in BIOGEM for the changes in phosphate and dissolved organic phosphorus (DOP) concentrations occurring in the surface ocean layer (but omitting the extra transport terms that are calculated by the ocean circulation model; Edwards and Marsh, 2005) are:

$$\frac{\partial \mathrm{PO_4}}{\partial t} = -\Gamma + \lambda \mathrm{DOP} \tag{1}$$

$$\frac{\partial \mathrm{DOP}}{\partial t} = \nu\Gamma - \lambda \mathrm{DOP} \tag{2}$$

$$\Gamma = u_0^{\mathrm{PO_4}} \cdot \frac{\mathrm{PO_4}}{\mathrm{PO_4} + K^{\mathrm{PO_4}}} \cdot (1 - A) \cdot \frac{I}{I_0} \tag{3}$$

where Γ is the biological uptake of PO_4. Γ is calculated from: (i) $u_0^{\mathrm{PO_4}}$, an assumed maximum uptake rate of phosphate (mol PO_4 kg^{-1} yr^{-1}) that would occur in the absence of any limitation on phytoplankton growth, and (ii) a Michaelis-Menten type kinetic limitation of nutrient uptake, of which $K^{\mathrm{PO_4}}$ is the half-saturation constant. Because of the degree of abstraction of ecosystem function inherent in our model, the appropriate values for either $u_0^{\mathrm{PO_4}}$ or $K^{\mathrm{PO_4}}$ are not obvious, and so are subsequently calibrated (see Sect. 3 and Table 1). We apply two modifiers on productivity representing the effects of sub-optimal ambient light levels and the fractional sea ice coverage of each grid cell (A) (Edwards and Marsh, 2005). A full treatment of the effects of light limitation on phytoplankton growth is beyond the scope of this current paper. The strength of local insolation (I) is therefore simply normalized to the solar constant (I_0) to give a limitation term that is linear in annual incident insolation. A proportion (ν) of PO_4 taken up by the biota is partitioned into dissolved organic phosphorus (DOP). The relatively labile dissolved organic molecules are subsequently remineralization with a time constant of $1/\lambda$. The values of ν and λ are assigned following the assumptions of the OCMIP-2 protocol (Najjar and Orr, 1999): ν=0.66 and λ=0.5 yr^{-1}.

The particulate organic matter fraction is exported vertically and without lateral advection out of the surface ocean layer at the next model time-step. Because there is no explicit standing plankton biomass in the model, the export flux of particulate organic phosphorus ($F^{\mathrm{POP}}_{z=h_e}$, in units of mol PO_4 m^{-2} yr^{-1}) is equated directly with PO_4 uptake (Eq. 1):

$$F^{\mathrm{POP}}_{z=h_e} = \int_{h_e}^{0} \rho \cdot (1 - \nu) \cdot \Gamma \, dz \tag{4}$$

where ρ is the density of seawater and h_e the thickness of the euphotic zone (175 m in the 8-level version of this ocean model).

In the production of organic matter, dissolved inorganic carbon (DIC) is taken out of solution in a 106:1 molar ratio with PO_4 (Redfield et al., 1963) while O_2 takes a -170:1 ratio with PO_4 (Anderson and Sarmiento, 1994). The effect on total alkalinity (ALK) of the biological uptake and remineralization of nitrate (NO_3) is accounted for via a modification of ALK in a -1:1 ratio with the quantity of NO_3 transformed. Because we do not model the nitrogen cycle explicitly in this paper, we link ALK directly to PO_4 uptake and remineralization through the canonical 16:1 N:P ratio (Redfield et al., 1963). For convenience, we will describe the various transformations involving organic matter in terms of carbon (rather than phosphorus) units, the relationship between organic matter export fluxes being simply:

$$F^{\mathrm{POC}}_{z=h_e} = 106 \cdot F^{\mathrm{POP}}_{z=h_e} \tag{5}$$

We represent the remineralization of particulate organic carbon (POC) as a process occurring instantaneously throughout the water column. We partition POC into two distinct fractions with different fates in the water column, following Ridgwell (2001) but adopting an exponential decay as an alternative to a power law. The POC flux at depth z in the water column is:

$$F^{\mathrm{POC}}_{z} = F^{\mathrm{POC}}_{z=h_e} \cdot \left(\left(1 - r^{\mathrm{POC}}\right) + r^{\mathrm{POC}} \cdot \exp\left(\frac{z_{h_e} - z}{l^{\mathrm{POC}}} \right) \right) \tag{6}$$

The parameters: l^{POC}, the length-scale and r^{POC}, the initial partitioning of POC into a labile fraction are both calibrated