

GSA Repository Item (Hotinski et al.)**Additional Model Description and Parameters**

The model used was MOM v. 1, a three-dimensional primitive equation model derived from that described by Bryan (1979) and Bryan and Lewis (1979), using rigid lid and Boussinesq approximations. Lateral boundary conditions are no-slip with insulating walls for heat and salt, and there is a no-flux boundary at the ocean for heat and salt at the ocean floor. Constant vertical and horizontal viscosity coefficients were used with values of $20 \text{ cm}^2/\text{s}$ and $2.5 \times 10^9 \text{ cm}^2/\text{s}$. Mixed boundary conditions were used for temperature and salinity forcing. Upper ocean layer temperatures were relaxed toward specified forcing values using Newtonian damping with a restoring time scale of 16 days, corresponding to a transfer velocity of 3 m/s for a 50 m thick mixed layer. Surface salinities were similarly relaxed to average zonal values to produce a salt flux calculated from a moisture flux (evaporation minus precipitation) predicted by GENESIS (land and ocean) using the approach of Weaver et al. (1994).

Oxygen and phosphate were added to the model code as reactive tracers and set to uniform initial values (170 and $2.15 \text{ } \mu\text{mol/L}$, respectively) consistent with modern observational data from the deep ocean (Levitus and Boyer, 1994; Conkright et al., 1994). The model is a closed system with respect to phosphate, but is open with respect

to oxygen at the surface, where oxygen concentrations are restored to saturation concentrations at predicted temperatures and salinity of water in the model's uppermost layer. Both production of oxygen during photosynthesis and loss of oxygen during decomposition are linked to phosphate uptake and release by the Redfield ratio of 169:1 (Takahashi et al., 1985). Phosphate and oxygen have no-flux boundaries at the bottom of the simulated ocean, and there is no flux of phosphate into the ocean's surface. Oxygen concentrations in surface waters are relaxed toward values in equilibrium with model-predicted temperatures and salinities according to the relationship of Weiss (1970).

Carbon export from the photic zone is given by the following equation (Yamanaka and Tajika, 1996):

$$F_{\text{exp}} = r * L_f * [\text{PO}_4] * [\text{PO}_4] / (h + [\text{PO}_4]). \quad (1)$$

The export flux is linearly related to the phosphate concentration, except when phosphate declines to a value near the half-saturation constant, h , of $0.02 \mu\text{mol/L}$ (Maier-Reimer, 1993). Uptake is modified by a light limitation term, L_f , which is the cosine of latitude, and a rate constant, r , of 0.8 yr^{-1} (Yamanaka & Tajika, 1996).

The exported organic flux is instantaneously remineralized below the euphotic zone according to the power law of Martin et al. (1987):

$$F(z) = F_{\text{exp}} (z/z_e)^{-0.858} \quad (2)$$

where F_{exp} is organic matter exported from the euphotic zone, $F(z)$ is the remaining particulate flux at depth z , and z_e is the depth of base of the euphotic zone.

The model was run for with a timestep of 200 s for momentum calculations and 150,000 s for tracers, and timesteps did not vary with depth.

Effect of Elevated Phosphate Concentration

Increasing the mean nutrient concentration of the ocean is one way of increasing export production without changing circulation. To evaluate the effects of increased nutrients, the reduced-gradient experiment was rerun with a mean phosphate concentration 50% higher than the modern value. Dissolved oxygen concentrations in this simulation are dramatically reduced in intermediate waters (Fig. 1A), and there is an intensified zone of anoxia off the western coast of Pangea. More negative values of O_2 (as low as $-600 \mu\text{mol/L}$) in this region indicate extensive sulfate reduction in intermediate waters (Fig. 1B). Deep-water oxygen values are also negative, values approaching $-200 \mu\text{mol/L}$. These values, however, do not approach the $-1000 \mu\text{mol } O_2/\text{L}$ estimated by Knoll et al. (1996), which would require even higher ocean phosphate inventories.

Model Limitations

Our Permian ocean model contains a number of simplifications, but these do not significantly alter our findings. First, we used a simple vertical mixing scheme and a flat bathymetry. However, new reduced-gradient experiments with more realistic bathymetry, including a shallow (500 m deep) Tethys ocean, show no significant

deviations from the results presented here, and although our constant vertical diffusivity of $1 \text{ cm}^2/\text{s}$ is high relative to modern ocean observations (Ledwell et al., 1993), use of a smaller coefficient should only decrease estimates of deep ocean oxygenation. Kutzbach et al. (1990) found symmetrical thermohaline circulation in an idealized, symmetrical Permian ocean with symmetrical surface forcings. The asymmetry in our model results from both more complex paleogeography and asymmetrical surface forcing. The Southern Hemisphere landmass in the Permian reconstruction has an extensive, continuous north-south-oriented boundary while landmasses in the Northern Hemisphere are less meridionally oriented and are broken by seaways. This configuration supports stronger meridional transport in the Southern Hemisphere, an effect enhanced by a higher surface density gradient in the Southern Hemisphere and higher density south polar water.

Second, the GENESIS atmospheric model that provided the high-gradient surface forcings has a built-in hemispheric asymmetry in albedo, which tends to make the southern hemisphere too cool in pre-Cenozoic climate simulations (D. Pollard, personal communication, 2000). As our reduced-gradient temperature forcing is specified to be symmetric, this asymmetry does not affect the result of the warm-pole scenario but may enhance overturning asymmetry in the high-gradient case. Third, models with coarse vertical resolution have been shown to trap nutrients in shallow water and underestimate the export of organic matter to deep waters (Najjar et al., 1992; Aumont et al., 1999). Since the deep ocean is anoxic almost everywhere in reduced-gradient simulations, our finding of widespread anoxia does not seem susceptible to this spatial bias. Lastly, the

oceanic inventory of phosphate in each of our simulations is invariant. In an open-system scenario the reduction in productivity, and thus organic carbon burial, between the high- and low-gradient cases would increase ocean phosphate levels (Hotinski et al., 2000), and subsequent anoxia might also raise nutrient levels by regeneration of phosphate from sediments (Ingall and Jahnke, 1994; 1997; Van Cappellen and Ingall, 1994). Such additional increases in phosphate would enhance anoxia in our simulations.

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Figure Captions

Figure 1. Maps of steady-state oxygen concentrations ($\mu\text{mol/L}$) for (A) upper intermediate and (B) deep ocean in reduced-gradient scenario with 150% modern phosphate.

Figure 1 (Repository) - Hotinski et al.

