

## Reply to Comment by Roberta M. Hotinski, Lee R. Kump, and Karen L. Bice on “Could the Late Permian deep ocean have been anoxic?”

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[1] Two recent studies [Hotinski *et al.*, 2001; Zhang *et al.*, 2001] have modeled and discussed Late Permian ocean circulation and oxygenation. These studies have reached significantly different conclusions. Zhang *et al.* [2001] find that a “thermal mode” ocean circulation driven by cooling in polar latitudes is unlikely to support deep-sea anoxia, but a “haline mode” ocean circulation, a shallow overturning cell driven by enhanced evaporation from the subtropics, perhaps could lead to global-scale deep-sea anoxia. Hotinski *et al.* [2001], using a different model, find that a “thermal mode” ocean circulation with a high equator-to-pole gradient gives well-oxygenated Permian deep water, while a “thermal mode” ocean circulation with a very low equator-to-pole gradient leads to global deep-sea anoxia. This discrepancy prompted a comment by Hotinski *et al.* [2002].

[2] In response, we would like to begin by emphasizing the common underlying feature of the two studies. In both models, deep ocean oxygen concentrations are significantly reduced if the overturning circulation becomes weaker with warmer, less oxygenated waters ventilating the deep ocean. In some circumstances this can lead to a dominance of respiration of organic matter and a very significant depletion of deep ocean oxygen. However, as noted above, the two studies differ greatly in the nature of the ocean circulation in which such a scenario can be achieved.

[3] Hotinski *et al.* [2002] argue that the difference between the two model results are “in large part the result of dissimilar definitions of ‘weak’ or ‘reduced’ meridional temperature gradients.” We agree that the two studies apply different pole-equator temperature gradients at the surface. The high-latitudes surface temperature used to force the

model in our study [Zhang *et al.*, 2001, Figure 5b] is much warmer than that of modern climate and is based on an atmospheric circulation model of the Late Permian [Kutzbach and Gallimore, 1989], corresponding to a warm climate with enhanced greenhouse effect ( $5 \times \text{CO}_2$ ) and reduced solar luminosity (1% decrease). The high-latitudes surface temperature is more than 10°C (15°C) warmer than modern climate at 70°N (70°S). The warmer high-latitudes surface temperature is also similar to that of another modeling study of the Late Permian climate [Kutzbach and Ziegler, 1993], and is consistent with the Late Permian paleobotanical data interpreted by Rees *et al.* [1999]. The tropical surface temperature is about 3°C warmer than present, resulting in pole-equator temperature gradient that is weaker than that of modern climate. On the other hand, our modeled Late Permian polar sea surface temperature (SST) is cool relative to of the “low-gradient” scenario of Hotinski *et al.* [2001] forced by the estimated Paleocene-Eocene thermal maximum (PETM) SST [Bice and Marotzke, 2001].

[4] If we use much higher polar surface temperature to force the model to give the estimated PETM polar SST as in the work by Hotinski *et al.* [2001], the thermal mode ocean circulation would be weakened. However, our recent theoretical study [Zhang *et al.*, 2002] showed that when the pole-equator surface temperature gradient is reduced to a certain threshold, the weakened “thermal mode” ocean circulation would switch to the “haline mode” circulation, and the critical freshwater flux required for the ocean to switch to the haline mode circulation is much smaller with the reduced pole-equator surface temperature gradient. In that case, the deep ocean oxygen distribution would be similar to what we found with the haline mode circulation. Now the issue is when the pole-equator surface temperature gradient is larger than the threshold of mode switching, whether the weakened thermal mode circulation (with a maximum of  $\sim 25$  Sv as shown by Hotinski *et al.* [2001]) originating from warm polar surface water can drive global deep-sea anoxia. We note that the weakened thermal mode

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circulation of *Hotinski et al.* [2001, Figure 2d] is of comparable strength to that in the modern North Atlantic Ocean. Why then, is the modern North Atlantic deep ocean at depths of 3000 m to 3500 m (which is mainly dominated by the North Atlantic deep water) so well oxygenated ( $\sim 246 \mu\text{mol L}^{-1}$ ), whereas wide-spread, deep-sea anoxia is sustained in the late Permian scenario of *Hotinski et al.* [2001] in a circulation of similar vigor? The difference in oxygen solubility due to changes in the temperature at the deep water source region partially accounts for this. The warm polar sea surface temperature ( $12^\circ\text{C}$ ) of *Hotinski et al.* [2001] results in lower oxygen concentration ( $\sim 250 \mu\text{mol L}^{-1}$ ) at the deep water source region. However, comparing the modern North Atlantic value deep water source region of  $\sim 336 \mu\text{mol L}^{-1}$ , the oxygen solubility effect can only account for  $\sim 86 \mu\text{mol L}^{-1}$  reduction in oxygen concentration at the deep water source region, far from sufficient to induce deep-sea anoxia.

[5] *Zhang et al.* [2001] suggested that use of “negative oxygen” by *Hotinski et al.* [2001] leads to the impression of widespread anoxia in their well-ventilated, “slow,” Permian Ocean circulation. *Hotinski et al.* [2002] argue that allowing negative oxygen concentrations represents the transport and oxidation of  $\text{H}_2\text{S}$ , a product of the oxidation of organic matter by sulfate. The negative oxygen approach enhances the potential for anoxia, relative to the scheme used by *Zhang et al.* [2001], where we chose to limit oxygen to nonnegative concentrations. In a simulation of the modern ocean using our model in which negative oxygen values are permitted, we find very exten-

sive regions where the dissolved oxygen is negative in the Pacific and Indian oceans. However, these regions are not anoxic in the present ocean. Hence the interpretation of similar regions as anoxic condition in a Permian simulation with the negative oxygen approach must be viewed with extreme caution. In addition, in the absence of oxygen, the primary oxidant of particulate organic material is nitrate and significant denitrification of the ocean could occur. This could lead to extensive nitrogen limitation of new and export production [e.g., *Falkowski*, 1997] before sulfate becomes the primary oxidant. Besides, oxidation of  $\text{H}_2\text{S}$  must not be necessarily by use of elemental  $\text{O}_2$ . Because of these complications, *Zhang et al.* [2001] chose not to allow oxygen concentrations to become negative, leading, we feel, to a clearer interpretation. However, we recognize that the complex interactions under anoxic condition must ultimately be understood and modeled in more detail. With different parameterizations of the anoxic condition in the biogeochemical models of the two studies, it is difficult to compare the results directly. Further investigations on the processes under the anoxic condition in the real ocean and their parameterizations in the biogeochemical models are deserved.

[6] Clearly the choice of physical and biogeochemical parameterizations and boundary conditions in both models strongly impacts upon the results and inferences. Our studies illustrate some of the range of possibilities, though by no means all. We have been stimulated by these contrasting inferences and look forward to continued progress and future dialogue.

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