Is AOU a good measure of respiration in the oceans?

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</tr>
<tr>
<td>Publisher</td>
<td>American Geophysical Union (AGU)</td>
</tr>
<tr>
<td>Version</td>
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</tr>
<tr>
<td>Accessed</td>
<td>Thu Dec 06 06:14:46 EST 2018</td>
</tr>
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Is AOU a good measure of respiration in the oceans?

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Received 1 July 2004; accepted 20 August 2004; published 14 September 2004.

[1] Apparent Oxygen Utilization (AOU) is widely used to infer respiration in the oceans by assuming that surface oxygen concentration is close to saturation with the overlying atmosphere. However, significant disequilibrium of oxygen has been observed in high latitude surface oceans where the deep waters are formed. We explicitly calculate True Oxygen Utilization (TOU) in a global ocean physical-biogeochemical model to evaluate the ability of AOU to represent respiration. We find significant differences between AOU and TOU in the deep waters, suggesting a systematic overestimation of respiration when inferred from AOU. The surface heat flux and the entrainment of thermocline waters together drive the surface undersaturation of oxygen in the regions of water mass formation, and their influences are significantly enhanced by sea ice cover at high latitudes.


1. Introduction

[2] Dissolved oxygen is generally depleted in thermocline and abyssal oceans due to the respiration of organic material. Air-sea equilibration of oxygen is generally rapid relative to ocean circulation timescales, and observed surface oxygen concentrations are indeed close to saturation in many areas of the surface oceans [e.g., Broecker and Peng, 1982; Chester, 2000]. Depletion of oxygen relative to saturation has been used as a measure of respiration in the interior ocean. AOU is defined as the difference between the saturation oxygen concentration, \( O_{2,\text{sat}} \), and the observed oxygen concentration, \( O_2 \) as defined in equation (1).

\[
AOU = O_{2,\text{sat}} - O_2
\]  

where \( O_{2,\text{sat}} \) depends on potential temperature, \( T \), and salinity, \( S \). Estimates of respiration inferred from AOU could contain errors due to several processes as pointed out by Shiller [1981], Broecker and Takahashi [1985] and Broecker et al. [1991], which include air-sea disequilibrium in the regions of water mass formation, non-linearity in the solubility of oxygen, and respiration involving denitrification. The impact of denitrification can be diagnosed from the distribution of \( N^* \) [Gruber and Sarmiento, 1997; Sabine et al., 1999] whereas the impact of the other effects have not been well-understood nor quantified. In this study, we explicitly determine the decoupling of AOU from respiration in a 3-dimensional physical-biogeochemical model.

[3] True Oxygen Utilization (TOU) [Broecker and Peng, 1982] is defined as the difference between the preformed oxygen content, \( O_{2,\text{pre}} \), and the observed oxygen content.

\[
TOU = O_{2,\text{pre}} - O_2
\]

\[
\Delta O_2 = TOU - AOU = O_{2,\text{pre}} - O_{2,\text{sat}}
\]

Preformed oxygen is set at the time of subduction and is transported into the interior ocean by physical circulation. Thus the preformed properties typically represent the mixed layer properties during winter seasons. Differences between AOU and TOU can arise from the surface disequilibrium at the time of subduction and water mass formation, and from mixing in the interior ocean combined with the nonlinearity in the solubility of oxygen. \( \Delta O_2 \) is exactly equal to the degree of saturation at the surface; see equation (3). \( \Delta O_2 \) is mapped into the interior ocean through advection and mixing. While \( O_{2,\text{sat}} \) can be determined from observed temperature and salinity, it is impossible to determine \( \Delta O_2 \) from direct observation since the true preformed oxygen content remains unknown.

2. Decoupling of AOU From TOU

[4] Here, we explicitly simulate \( \Delta O_2 \) using the MIT ocean general circulation and biogeochemistry model [Marshall et al., 1997a, 1997b; P. Parekh et al., Decoupling of phosphorus and iron in the global ocean, submitted to Global Biogeochemical Cycles, 2004, hereinafter referred to as Parekh et al., submitted manuscript, 2004] by calculating the global distribution of preformed oxygen, \( O_{2,\text{pre}} \) as an additional tracer. The model also includes explicit representation of the biogeochemical cycles of DIC, Alkalinity, \( O_2 \), \( PO_4 \), DOP and Fe. Horizontal resolution is 2.8° and there are 15 vertical layers. The model is forced dynamically with monthly climatological surface boundary conditions of wind stress and buoyancy flux [Trenberth et al., 1989; Conkright et al., 2002]. The model employs explicit iron chemistry forced with modeled monthly dust deposition field [Mahowald et al., 2003; Parekh et al., submitted manuscript, 2004]. Air-sea gas transfer is parameterized following [Wanninkhof, 1992]. Biological productivity can be limited by the availability of Fe or \( PO_4 \). All biological processes are assumed to occur in Redfieldian stoichiometry [Anderson and Sarmiento, 1994]. Preformed oxygen is set equal to simulated \( O_2 \) in the surface layer, and transported as a conservative, passive tracer in the ocean interior.

[5] Typically, estimates of oxygen utilization, and regenerated carbon and nutrients, rely on the assumption that \( \Delta O_2 \)
Figure 1. Simulated, annual-mean surface disequilibrium, $\Delta O_2$. Contour interval is 10 $\mu$M. Solid line represents undersaturation, $\Delta O_2 < 0$. See color version of this figure in the HTML.

Figure 2. $\Delta O_2$ is shown in the Atlantic (23W) and the Pacific (165W) of the control run. Contour interval is 10 $\mu$M. Dashed line represents zero and negative values. These results lead to local errors of up to 55 $\mu$M of regenerated carbon. See color version of this figure in the HTML.

Figure 3. Schematic diagram of the physical processes driving the surface oxygen away from saturation. Heat loss at the surface increases the solubility of oxygen, which leads toward undersaturation. Air-sea gas transfer is inhibited by sea ice. Thermocline and deep waters are depleted in oxygen relative to surface waters. Convective mixing, over a background gradient of $O_2$ can lead to undersaturation by bringing up the $O_2$-depleted intermediate and deep waters. See color version of this figure in the HTML.

3. What Controls the Disequilibrium of Oxygen in the Polar Oceans?

What drives the large undersaturation in polar surface oceans? Figure 3 schematically illustrates the processes affecting the preformed oxygen near the region of deep water formation. Heat loss and the associated solubility increase could drive the surface condition toward undersaturation. Convective mixing or entrainment of intermediate and deep waters could also enhance undersaturation by bringing oxygen depleted waters to the surface. The magnitude of $\Delta O_2$ is set by the relative timescales of these processes with respect to the air-sea gas exchange. A large $\Delta O_2$ is expected when the rate of solubility increase due to heat loss or the rate of entrainment flux are relatively rapid compared to that of the air-sea gas exchange. Air-sea gas exchange can be inhibited by sea ice cover effectively reducing the rate of air-sea gas equilibration, significantly amplifying the effects of heat flux and entrainment.

In a suite of sensitivity studies, we repeat the numerical simulation suppressing each of the three key processes in the model: (A; No-ICE) the inhibition of air-sea gas transfer by sea ice, (B; No-BIO) the effect of respiration on oxygen in the interior ocean, and (C; No-HF) the effect of heat fluxes on the solubility of oxygen in the surface ocean, keeping all other parameters fixed. The results of the sensitivity
experiments are summarized in Table 1, and Figure 4 shows the distribution of $\Delta O_2$ in the Southern Ocean.

[8] In experiment (A), without the effect of sea ice on air-sea gas transfer, we find that the magnitude of the surface disequilibrium is significantly reduced by $-64\%$ in the Weddell Sea, and by $-86\%$ in the Labrador Sea, suggesting a substantial impact of sea ice on the gas exchange rates in the polar oceans. This effect has little impact on $\Delta O_2$ in the western boundary currents.

[9] In experiment (B) where we suppress the effect of respiration on dissolved oxygen, surface undersaturation is reduced significantly in the high latitude Southern Ocean (approximately $-65\%$; see Table 1 and Figure 4b), though in the Labrador Sea, the response is much weaker (approximately $-24\%$). The air-sea disequilibrium is driven by the mixing and entrainment of intermediate and deep waters which have accumulated a large oxygen deficit due to respiration. This effect is more pronounced in the polar Southern Ocean where upwelling deep waters are older. This effect has significant impacts on $\Delta O_2$ in the western boundary currents, particularly in the Pacific.

[10] In experiment (C) where we suppress the temperature dependence of the solubility of oxygen, surface disequilibrium is reduced by $-27\%$ in the Weddell Sea and by $-70\%$ in the Labrador Sea (see Table 1 and Figure 4c). Heat losses and associated solubility increase play a relatively minor role in the high latitude Southern Ocean whereas this effect dominates in the North Atlantic. This effect also has impacts on $\Delta O_2$ in the western boundary currents.

[11] In the polar oceans, the effects of the heat loss and the entrainment of old, deep waters together drive the surface waters toward undersaturation, and they control the magnitude of $\Delta O_2$ in the interior ocean. Combination of these effects are largely additive, suggesting that the simulated response of the surface disequilibrium is quasi-linear in this particular model. Relative importance of these processes varies over geographical regions in the model. The effect of the ice cover amplifies the overall undersaturation by reducing the rate of air-sea gas transfer.

4. Discussion

[12] Surface disequilibrium of dissolved oxygen directly affects the distribution of AOU in the interior ocean and decouples AOU from TOU by a significant magnitude. The decoupling is large and systematic, and its spatial pattern is primarily oriented along isopycnals. Estimates of regenerated carbon and nutrients potentially include O(1) error when inferred from AOU. Specific respiration rates inferred from a time rate of change in AOU, so-called Oxygen Utilization Rate (OUR), may be less affected by this error [Jenkins, 1982]. The $\Delta C^*$ method [Gruber et al., 1996; Gruber, 1998], which is a method for estimating anthropogenic CO$_2$ in the oceans, may also be affected by the effect of oxygen disequilibrium. Overestimation of respiration will result in an underestimation of preformed carbon and $\Delta C^*$. However this may be partially accounted in the empirically determined residual term, $\Delta C_{diseq}$. More studies are needed to access its impact on the uncertainty of anthropogenic CO$_2$ inventory estimates. Since the spatial distribution of $\Delta O_2$ is systematic and primarily oriented along isopycnals, it may be possible to develop a method to correct AOU including the effect of surface disequilibrium by combining numerical models and multiple tracer observations such as noble gases [Hamme and Emerson, 2002]. We do not claim that the numerical model used here can adequately capture various physical processes controlling the disequilibrium of surface oxygen, such as bubble formation and mesoscale ocean circulation. Further theoretical and experimental studies are required to understand and quantify these physical processes controlling the saturation state of oxygen in the polar surface oceans.

Figure 4. Surface disequilibrium, $\Delta O_2$, in the Southern Ocean from the control run and the sensitivity experiments (A, B, C). Contour interval is 10 $\mu$M. Dash line represents $\Delta O_2 = 0$. In the experiment (A), the inhibition of air-sea gas transfer due to sea ice is suppressed. In the experiment (B), the effect of respiration on the oxygen is suppressed. In the experiment (C), the temperature dependence of the solubility of oxygen is suppressed. See color version of this figure in the HTML.

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*Simulated $\Delta O_2$ in several geographical regions from each sensitivity experiments are shown in units of $\mu$M. Numbers within the parenthesis represent the fractional changes relative to control, defined as $(\Delta O_2/\Delta O_2(\text{control}) - 1) \cdot 100\%$. 

Acknowledgments. We would like to acknowledge the fruitful collaboration with Stephanie Dutkiewicz and Payal Parekh of MIT for the
development of the physical-biogeochemical model. TI is grateful for the support from the Office of Polar Program at NSF. MJF and EAB are grateful for support from NSF grant OCE-0136609 and OCE-0250672. Two anonymous reviewers provided useful comments for improving the manuscript.

References


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