

Appendix A Calculation of Equilibrium Oxygen Concentration

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1 Introduction

In the chemical literature solubility of a gas is generally given as the Bunsen absorption coefficient, α , which is a function of temperature and of the electrolyte content of the solvent. Hydrostatic pressure has little effect on solubility, but with increasing head the escaping tendency of the gas increases. Therefore, at constant concentration, a POS gives a higher reading at greater depth. This effect is negligible ($< 1\%$) at depths less than 100 m.

If α or an equivalent expression of solubility is known, the saturation concentration for any particular set of conditions can readily be calculated. Unfortunately, gas solubility is not a simple function of the controlling variables, and empirical data obtained by physical or chemical methods has to be used. Since the pioneering work by Winkler [21, 22], differing sets of such data have been published, and it has been a matter of much debate which particular set of values is more accurate. Mortimer [14] has reviewed the existing data and has convincingly recommended the solubility data recently published by Benson and Krause [2, 3] as a standard for limnological application. Extensive aids for determining oxygen solubility in the temperature range 0° – 40°C and for atmospheric pressure from 61–111.5 kPa (0.6–1.1 atm) are available in form of nomograms and equations [14]. Essential equations from [3, 14] are also presented here and their use is recommended for POS applications.

2 Solubility Measures

1. *The Bunsen Absorption Coefficient, α* , is the volume of gas (STP) dissolved in a unit volume of solvent at standard partial pressure of the gas $p_{\text{O}} = 1$ atm (101.325 kPa or 760 Torr) at a specific temperature. The dimension is volume per volume per pressure.

2. *The Solubility Coefficient, S_s* , has the dimension of amount of substance per volume per pressure. If moles are taken as units of amount of substance, the relation to α is

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$$S_s = \frac{\alpha}{\text{molar volume} \times K}, \quad (1)$$

where the molar volume is $22.393 \text{ dm}^3 \text{ mol}^{-1}$ and K is a conversion factor depending on the unit of pressure chosen (1 for atm, 101.325 for kPa, and 760 for Torr).

Example: For pure water at 10°C , $\alpha = 0.0381726 \text{ dm}^3 \text{ dm}^{-3} \text{ atm}^{-1}$, expressing pressure in kPa, the solubility of oxygen in the sample is then

$$S_s = \frac{0.0381726}{22.393 \times 101.325} = 16.824 \times 10^{-6} \text{ mol dm}^{-3} \text{ kPa}^{-1} \\ = 16.824 \mu\text{mol dm}^{-3} \text{ kPa}^{-1}. \quad (2)$$

3. Avoiding the term "solubility", Benson and Krause [2] define *unit standard concentration*, C^* , of atmospheric oxygen in pure water as the concentration of dissolved oxygen per unit volume of solution (measured at equilibrium temperature) "when it is in equilibrium with an atmosphere of standard composition and saturated with water vapor at a total pressure (including that of water vapor) of 101.325 kPa (1 atm)". When referred to unit mass of solution, the unit standard concentration is designated C^\dagger . C^* is converted to C^\dagger by dividing by the density of the solution at the temperature concerned, or (to a very close approximation) by the density of pure water at that temperature.

Unit standard concentration C^* and the solubility coefficient S_s are related by

$$C^* = S_s \times (p_o - p_{\text{H}_2\text{O}}) \times \phi_{\text{O}_2}^*, \quad (3)$$

where $p_{\text{H}_2\text{O}}$ is the partial pressure of water vapor and $\phi_{\text{O}_2}^*$ is the volume fraction of oxygen in dry air which is 0.20946.

Example: At 10°C $S_s = 16.824 \mu\text{mol dm}^{-3} \text{ kPa}^{-1}$ [Eq. (2)]

$$p_{\text{H}_2\text{O}} = 1.22763 \text{ kPa, [see Eq. (19)]}$$

then

$$C^* = 16.824 \times (101.325 - 1.22763) \times 0.20946 \\ = 352.739 \mu\text{mol dm}^{-3}. \quad (4)$$

In the biological and limnological literature concentration is generally the preferred measure of solubility, and since very accurate approximation formulas are available we will adhere to this usage. If required, C^* can easily be converted to S_s or α , using the relation of Eqs. (1) and (3). When C^* is given as $\mu\text{mol dm}^{-3}$, S_s in $\mu\text{mol dm}^{-3} \text{ kPa}^{-1}$ is

$$S_s = \frac{C^*}{(101.325 - p_{\text{H}_2\text{O}}) \times 0.20946} \quad (5)$$

and α in $\text{dm}^3 \text{ dm}^{-3} \text{ atm}^{-1}$ is

$$\alpha = \frac{C^*}{(101.325 - p_{\text{H}_2\text{O}}) \times 92.315} \quad (6)$$

3 Calculation of Unit Standard Concentration

In applications involving a POS the accuracy provided by the Benson and Krause solubility tables [3] is hardly necessary and the empirical equation [Eq. (7)] fitted by Mortimer [14] can be used. Over the temperature range 0° – 37.5°C the interpolation error is not more than 0.05%. To be consistent with the S.I. $\mu\text{mol dm}^{-3}$ should be used as units, but the familiar mg dm^{-3} (equivalent to mg l^{-1}) has also been included. To obtain the desired unit the constant A in Eq. (7) is selected from Table 1.

The unit standard concentration (at temperature θ , in $^\circ\text{C}$) is

$$C^* = \exp [A - 1.31403 \ln (\theta + 45.93)]. \quad (7)$$

Table 1

Units O ₂	A
$\mu\text{mol dm}^{-3}$	11.1538
mg dm^{-3}	7.7117
$\text{cm}^3 \text{ dm}^{-3}$ (ideal gas STP)	7.3557
$\text{cm}^3 \text{ dm}^{-3}$ (real gas STP)	7.3547

The calculation can be done quickly on any scientific pocket calculator and the concentration values are more accurate than many solubility tables hitherto in use.

The five term polynomial approximation from Eq. (3) will also be given, because it has been recommended as a standard [14]. The equation is presented in nested form, which is more efficient for programming on computers and calculators. The values for the B -coefficients and the constant A are taken from Table 2A or 2B, depending on whether C^* or C^\dagger is to be calculated.

$$C^* = \exp \{ \{ [(B_1 \times X + B_2) \times X + B_3] \times X + B_4 \} \times X + A \}, \quad (8)$$

$$X = 1/T; \quad 273.15 < T < 313.15 \text{ K} \quad (0^\circ\text{--}40^\circ\text{C}).$$

4 Dependence on Salt Concentration

Due to interactions between the molecular species in a solution, solubility of a non-electrolyte in a salt solution is different from that in pure water. This dependence can be described by a logarithmic relationship proposed by Setschenow [17].

$$\ln \left(\frac{S_s}{S_e} \right) = \kappa_e \times I. \quad (9)$$

S_s and S_e are the solubilities in pure water and in the electrolyte solution respectively, κ_e is the salting coefficient and I the ionic strength. For oxygen dissolved in water, as for most aqueous solutions, κ_e is positive which means that with increasing salt con-

Table 2

(A) C^* = amount of oxygen per unit volume

B coefficients	Units O ₂	A
B_1	-8.621949×10^{11}	$\mu\text{mol dm}^{-3}$
B_2	1.243800×10^{10}	mg dm^{-3}
B_3	-6.642308×10^7	$\text{cm}^3 \text{ dm}^{-3}$ (ideal gas STP)
B_4	1.575701×10^5	$\text{cm}^3 \text{ dm}^{-3}$ (real gas STP)

(B) C^\dagger = amount of oxygen per unit mass

B coefficients	Units O ₂	A
B_1	-8.621061×10^{11}	$\mu\text{mol kg}^{-1}$
B_2	1.243678×10^{10}	mg kg^{-1}
B_3	-6.637149×10^7	$\text{cm}^3 \text{ kg}^{-1}$ (ideal gas STP)
B_4	1.572288×10^5	$\text{cm}^3 \text{ kg}^{-1}$ (real gas STP)

centration solubility decreases. This phenomenon is known as the salting-out effect. The coefficient κ_e varies with temperature, but depends also on the particular ions and the nature of the nonelectrolyte.

For seawater, values of oxygen solubility for a customary range of temperatures and salinities have been tabulated [4, 10, 14, 19]. Equations for calculating a correction factor have also been proposed [8, 9]. In the equations salt concentration is defined as chlorinity $[\text{Cl}^-]_{\text{‰}}$, which is g chlorine per kg of seawater. Since ionic composition of seawater remains highly constant throughout the oceans ($\pm 0.002\%$), the relationship between salinity (g total salt content per kg of seawater) and chlorinity is defined by the Knudsen relationship [13]:

$$\text{Salinity} = 1.805 [\text{Cl}^-]_{\text{‰}} + 0.03. \quad (10)$$

Of the two equations most commonly cited in the literature, that proposed by Green and Carritt [9]

$$S_e = S_s \times \exp \left[-[\text{Cl}^-]_{\text{‰}} \times \left(-0.1288 + \frac{53.44}{T} - 0.04442 \ln T + 7.145 \times 10^{-4} T \right) \right] \quad (11)$$

is usually recommended. It is based on measurements for $273.1 < T < 308.16$ K and $0 < [\text{Cl}^-] < 30\%$ and is more extensive than the equation from Fox [8]. Both equations agree to better than $\pm 1\%$ over the full range of Eq. (11) and with tabulated values [10]. It should be noted that the correction factor in Eq. (11) is for a solubility coefficient (or an absorption coeff.) and the effect of salinity on water vapor pressure is not accounted for. Applying a salinity correction to C^* or C_c , it can be shown, using Eqs. (3) and (5) that an additional correction factor

$$\frac{p_{\text{O}_2} - p_{\text{H}_2\text{O},e}}{p_{\text{O}_2} - p_{\text{H}_2\text{O}}} \quad (12)$$

would be required, where $p_{\text{H}_2\text{O},e}$ is the water vapor pressure in water with dissolved salts. In practice, over the range of salinity and temperature normally encountered, this correction will be near unity and it can be omitted.

5 Correction for Non-Standard Atmospheres

The standard concentrations C^* and C^\dagger obtained from Eqs. (7) and (8) apply for water, in equilibrium with air saturated with water vapor and exerting a total pressure $p_{\text{O}_2} = 101.325$ kPa. Since total pressure includes water vapor pressure

$$p_{\text{O}_2, \text{dry air}} = p_{\text{O}_2} - p_{\text{H}_2\text{O}}. \quad (13)$$

The partial molar volume of oxygen is 0.20946, therefore at standard atmospheric pressure the partial pressure of oxygen is

$$p_{\text{O}_2}^* = (p_{\text{O}_2} - p_{\text{H}_2\text{O}}) \times 0.20946. \quad (14)$$

If air and water are at equilibrium, water vapor is a function of temperature only. At constant temperature and any barometric pressure, ${}_b p$, the partial pressure of oxygen is therefore

$$p_{\text{O}_2} = ({}_b p - p_{\text{H}_2\text{O}}) \times 0.20946. \quad (15)$$

Combining Eqs. (14) and (15) we obtain

$$p_{\text{O}_2} = p_{\text{O}_2}^* \times \frac{{}_b p - p_{\text{H}_2\text{O}}}{p_{\text{O}_2} - p_{\text{H}_2\text{O}}} \quad (16)$$

or separating the water vapor effect

$$p_{\text{O}_2} = p_{\text{O}_2}^* \times \frac{{}_b p}{p_{\text{O}_2}} \times \frac{1 - p_{\text{H}_2\text{O}}/{}_b p}{1 - p_{\text{H}_2\text{O}}/p_{\text{O}_2}}. \quad (17)$$

Since the dissolved oxygen concentration in equilibrium with the gaseous phase is proportional to p_{O_2} , the saturation concentration in solution, C_c (see Table 7), is

$$C_c = C^* \times \frac{{}_b p}{p_{\text{O}_2}} \times \frac{1 - p_{\text{H}_2\text{O}}/{}_b p}{1 - p_{\text{H}_2\text{O}}/p_{\text{O}_2}} \quad (18)$$

Although expressed somewhat differently, Eq. (18) is almost identical with Eq. (4) from [14], only the extremely small correction term for molecular interactions in the atmospheric gas has been omitted.

The water vapor pressure over the range 0° – 40°C can be calculated to within 11 ppm of standard tabulated values by an approximation [14]. Again taking the appropriate constant from Table 3, $p_{\text{H}_2\text{O}}$ can readily be determined in the three most commonly used pressure units,

$$p_{\text{H}_2\text{O}} = \exp [(-216961 \times X - 3840.7) \times X + A] \quad (19)$$

$$X = 1/T; T \text{ in Kelvin; } 0^\circ\text{C} = 273.15 \text{ K.}$$

Table 3

Units p	A
kPa	16.4754
atm	11.8571
mm Hg	18.4904

With regard to the nonstandard atmospheric pressure, it is recommended that the value of ${}_b p$ to be used in Eq. (18) be obtained from a contemporary barometer reading. In a situation where this is not possible ${}_b p$ may be calculated from the pressure-altitude relation predicted by a standard atmospheric model, accepting an error of approximately $\pm 2.5\%$ due to varying local weather conditions. The power law

$${}_b p = p_o \times (1 - h/44.3)^{5.25}, \quad (20)$$

where h is the altitude in km, reproduces the ICAO standard atmosphere to within $\pm 0.004\%$ from -0.5 to 3.0 km, and with a small underestimate at higher levels increasing to -0.03% at 5.0 km [14]. Other approximations widely in use, like $\ln {}_b p = h/7.986$ [16] are much less accurate and Eq. (20) should be used instead. However, considering the actual variations of the real atmosphere, as compared to a standard atmosphere, this is really a moot point.

6 Correction for Hydrostatic Pressure: Absolute Saturation

A water column of 1 m exerts a hydrostatic pressure, ${}_w p$, of 9.80 kPa (0.0967 atm). Since the actual pressure, p_h [kPa], at any depth h [m] is the sum of atmospheric (barometric) and hydrostatic pressure,

$$p_h = {}_b p + {}_w p \quad (21)$$

the amount of gas which can be dissolved in equilibrium with a gas phase is linearly dependent on depth. This equilibrium condition has been termed absolute concentration [12, 15] to distinguish it from the gas solubility corrected for barometric pressure, ${}_b p$, only. The calculation of absolute saturation is of ecological interest in three situations: (1) Air bubbles carried down to moderate depths cause supersaturation with respect to the pressure at the water surface. (2) If air-saturated water warms up at some depth, supersaturation occurs on account of lower solubility at higher temperature. (3) Photosynthetic oxygen production leads to supersaturation (Chap. III.1) which can, above a critical partial pressure of oxygen, $p_{\text{O}_2, c}$, at small depths, even surpass the absolute saturation with the possibility of gas bubble formation.

In cases 1 and 2 of abiogenic supersaturation, $p_{\text{O}_2, c}$ at depth h is simply given by

$$p_{\text{O}_2, c} = (p_h - p_{\text{H}_2\text{O}}) \times \phi_{\text{O}_2}^* \quad (22)$$

According to Eqs. (16) and (18), the corresponding absolute saturation concentration, C_c , is

$$C_c = C^* \times \frac{p_h - p_{\text{H}_2\text{O}}}{p_o - p_{\text{H}_2\text{O}}} \quad (23)$$

Evaluation of $p_{\text{O}_2, c}$ is more complex in case 3, where, at constant partial pressure of other constituent gases, the biogenic increase in p_{O_2} results in the possibility of bubble formation. At this critical point a bubble contains gases in proportion to $p_{\text{O}_2, c}$ and to the partial pressure of all other gases, p_X , since

$$p_h = p_{\text{O}_2, c} + p_{\text{H}_2\text{O}} + p_X \quad (24)$$

If these bubbles rise they will remove the other gases and, while oxygen production continues, $p_{\text{O}_2, c}$ increases to the theoretical maximum, where only oxygen remains in solution

$$p_{\text{O}_2, c} = p_h - p_{\text{H}_2\text{O}} \quad (25)$$

This demonstrates that, for calculating absolute oxygen saturation in general the partial pressures of all dissolved gases must be known. These are hardly ever measured in limnological and oceanographic routine investigations except in studies particularly devoted to the problem of gas bubble disease [20]. Therefore appropriate assumptions have to be made and, at high p_{O_2} in natural waters, only nitrogen (including argon) has to be considered quantitatively whence

$$p_{\text{O}_2, c} = {}_b p - p_{\text{H}_2\text{O}} - p_{\text{N}_2} + {}_w p \quad (26)$$

At equilibrium with air at the water surface

$$p_{\text{N}_2} \approx {}_b p - p_{\text{H}_2\text{O}} - ({}_b p - p_{\text{H}_2\text{O}}) \times \phi_{\text{O}_2}^* \quad (27)$$

If equilibration took place at the temperature still prevailing at depth h , we can insert Eq. (27) into (26)

$$p_{\text{O}_2, c} = ({}_b p - p_{\text{H}_2\text{O}}) \times \phi_{\text{O}_2}^* + {}_w p \quad (28)$$

The first term incorporates the influence of gases other than oxygen (including water vapor) on $p_{\text{O}_2, c}$. This term becomes complicated, however, if temperature does not remain constant. Then the change in solubility of nitrogen and hence the change of p_{N_2} with temperature, and the dependence of $p_{\text{H}_2\text{O}}$ on temperature have to be accounted for. Restricting ourselves to the simplifying assumption of constant temperature, the absolute saturation concentration is calculated as²

² Hutchinson [12] discussed absolute saturation with reference to [15]. His equation for C_c contains a misunderstanding of the role played by water vapor, and hence C_c at 0 m differs erroneously from the surface saturation value

Table 4. Calculation of the critical p_{O_2} at depth h ($p_{O_2,c}$), of absolute saturation concentration (from surface concentration times the multiplication factor), and increase in absolute saturation concentration per metre of depth ($\Delta C_c \times m^{-1}$) as a percentage of air saturation in the range of temperatures between 4° and 25°C and altitudes between 0 and 2000 m

Source of supersaturation	$p_{O_2,c}$ [kPa]	Multiplication factor	$\Delta C_c \times m^{-1}$ [%]
Abiogenic	$(b^p - p_{H_2O}) \times \phi_{O_2}^* + 2.05 \times h$	1	$+\frac{9.80}{b^p - p_{H_2O}} \times h$ 9.8 to 12.9
Biogenic	$(b^p - p_{H_2O}) \times \phi_{O_2}^* + 9.80 \times h$	1	$+\frac{46.79}{b^p - p_{H_2O}} \times h$ 46.7 to 61.8
Pure O ₂	$(b^p - p_{H_2O}) + 9.80 \times h$	$4.774 + \frac{46.79}{b^p - p_{H_2O}} \times h$	

$$C_c = C^* \times \frac{b^p - p_{H_2O} + \frac{p}{w} \phi_{O_2}^*}{p_o - p_{H_2O}} \quad (29)$$

Table 4 compares expressions of the critical p_{O_2} for abiogenic [Eq. (22)] and biogenic sources of supersaturation [Eqs. (26–28)] and for the extreme case of saturation with pure oxygen [Eq. (25)]. It also lists the multiplication factors for obtaining absolute saturation at any depth from surface saturation concentrations [Eq. (18)] and shows the increase in absolute saturation concentration per metre of depth.

7 The Effects of Depth on p_{O_2} and Solubility

The effect of high hydrostatic pressure on p_{O_2} has been investigated by Enns et al. [3]. They found that with increasing hydrostatic head the partial pressure of a gas in solution, equilibrated at some lower reference pressure, increases exponentially. A number of other authors have discussed these results and an explanation for this effect, based on general thermodynamic principles, has been formulated [1, 5, 6]. A summary is given by Hitchman [11]. It has been shown that partial pressure varies according to

$$p_{O_2,h} = p_{O_2} \times \exp \left(\frac{M \times g_n \times h}{R \times T} \right) \quad (30)$$

where p_{O_2} and $p_{O_2,h}$ are the partial pressures at the surface and at depth h respectively, M is the molecular weight of the gas (32 for oxygen), g_n is the gravitational acceleration (9.81 N), h is the depth in metres, with R and T as defined in the gas law. Substituting p_o/p_{O_2} for $M/(RT)$ Eq. (30) becomes identical to the standard pressure/altitude relation. This means: for water saturated with a given gas at the surface the equilibrium partial pressure of that gas at any depth is equal to the partial pressure of

that gas if it were contained in a gas column extending from the surface to that particular depth [5]. Table 5 gives values of $p_{O_2,h}/p_{O_2}$ for several depths, calculated with Eq. (30). The water is assumed to be equilibrated with air ($M = 28.8$) at the surface and the mean water temperature is 5°C. It can be seen that at 500 m a correction of 6% will already be needed. Equation (30) shows that in this situation the increase in partial pressure is independent of the actual hydrostatic pressure and the density of the liquid and is entirely a function of the potential of the gas molecules dissolved in the water column. This effect is different from the case where liquid and gas phase coexist at some higher pressure, for instance, when air bubbles are carried downward. In the latter case equilibrium partial pressure is a linear function of hydrostatic pressure and is dependent on both the depth at which equilibration takes place and the density of the liquid (Table 4).

For gases in general, pressure also influences the equilibrium concentration (solubility). Following the same line of argument as before it has been shown [1, 6, 11] that

$$C_{c,h} = C_c \times \exp \frac{(M - v^\infty \times \rho) \times g_n \times h}{R \times T} \quad (31)$$

C_c and $C_{c,h}$ are the equilibrium concentrations at the surface and at depth, h , v^∞ is the partial molar volume of the gas at infinite dilution and ρ the density of the solvent, with M , g_n , h , R and T as defined in Eq. (30). For oxygen both M and v^∞ have a value of 32. The density of water is always near 1 and $(M - v^\infty \rho)$ is practically zero in the case of freshwater, or very small for seawater ($\rho = 1.023 \text{ g cm}^{-3}$). Therefore the ex-

Table 5. The increase of p_{O_2} with depth in water saturated with air at sea level pressure. Mean temperature of water column 5°C

Depth h in m	$p_{O_2,h}/p_{O_2}$
100	1.01
500	1.06
1000	1.13
5000	1.84
10000	3.40

Table 6. Conversion factors for units of amount of oxygen

	= μmol	mg	cm^3 (ideal gas STP)	cm^3 (real gas STP)
1 μmol	= 1	0.031999	0.022414	0.022392
1 mg	= 31.251	1	0.70046	0.69978
1 cm^3 (ideal gas STP)	= 44.615	1.4276	1	0.99902
1 cm^3 (real gas STP)	= 44.659	1.4290	1.00098	1

Relative molecular mass of oxygen $M_r(O_2) = 31.9988$
 Standard molar volume of ideal gas, STPD $V_o^I = 22.414 \text{ dm}^3 \text{ mol}^{-1}$
 Standard molar volume of real gas, STPD $V_m(O_2) = 22.392 \text{ dm}^3 \text{ mol}^{-1}$

Table 7. Solubility of oxygen, C_C , in air saturated pure water ($\mu\text{mol O}_2 \times \text{dm}^{-3}$)¹

Temp °C	Atmospheric pressure in kilopascal																			
	35	86	87	88	89	90	91	92	93	94	95	96	97	98	99	100	101 ²	102	103	104
0.0	382.9	387.4	392.0	396.5	401.0	405.6	410.1	414.6	419.2	423.7	428.2	432.8	437.3	441.8	446.4	450.9	456.9	460.0	464.5	469.0
1.0	372.3	376.7	381.1	385.5	389.9	394.3	398.8	403.2	407.6	412.0	416.4	420.8	425.2	429.6	434.0	438.4	444.3	447.3	451.7	456.1
2.0	362.1	366.4	370.7	375.0	379.3	383.6	387.9	392.2	396.5	400.8	405.1	409.3	413.6	417.9	422.2	426.5	432.2	435.1	439.4	443.7
3.0	352.4	356.6	360.8	365.0	369.2	373.3	377.5	381.7	385.9	390.1	394.2	398.4	402.6	406.8	411.0	415.1	420.7	423.5	427.7	431.9
4.0	343.2	347.2	351.3	355.4	359.5	363.5	367.6	371.7	375.7	379.8	383.9	388.0	392.0	396.1	400.2	404.2	409.6	412.4	416.5	420.5
5.0	334.3	338.2	342.2	346.2	350.2	354.1	358.1	362.1	366.0	370.0	374.0	378.0	381.9	385.9	389.9	393.8	399.1	401.8	405.7	409.7
6.0	325.8	329.7	333.5	337.4	341.3	345.2	349.0	352.9	356.8	360.7	364.5	368.4	372.3	376.1	380.0	383.9	389.0	391.6	395.5	399.4
7.0	317.7	321.5	325.2	329.0	332.8	336.6	340.3	344.1	347.9	351.7	355.5	359.2	363.0	366.8	370.6	374.4	379.4	381.9	385.7	389.5
8.0	309.9	313.6	317.3	321.0	324.6	328.3	332.0	335.7	339.4	343.1	346.8	350.5	354.2	357.9	361.5	365.2	370.1	372.6	376.3	380.0
9.0	302.4	306.0	309.6	313.2	316.8	320.4	324.1	327.7	331.3	334.9	338.5	342.1	345.7	349.3	352.9	356.5	361.3	363.7	367.3	370.9
10.0	295.3	298.8	302.3	305.8	309.4	312.9	316.4	319.9	323.4	327.0	330.5	334.0	337.5	341.1	344.6	348.1	352.8	355.1	358.7	362.2
11.0	288.4	291.8	295.3	298.7	302.2	305.6	309.1	312.5	315.9	319.4	322.8	326.3	329.7	333.2	336.6	340.0	344.6	346.9	350.4	353.8
12.0	281.8	285.2	288.5	291.9	295.3	298.7	302.0	305.4	308.8	312.1	315.5	318.9	322.2	325.6	329.0	332.3	336.8	339.1	342.4	345.8
13.0	275.5	278.8	282.1	285.4	288.7	291.9	295.2	298.5	301.8	305.1	308.4	311.7	315.0	318.3	321.6	324.9	329.3	331.5	334.8	338.1
14.0	269.4	272.6	275.8	279.1	282.3	285.5	288.7	292.0	295.2	298.4	301.7	304.9	308.1	311.3	314.6	317.8	322.1	324.2	327.5	330.7
15.0	263.5	266.7	269.8	273.0	276.2	279.3	282.5	285.6	288.8	292.0	295.1	298.3	301.5	304.6	307.8	310.9	315.1	317.3	320.4	323.6
16.0	257.9	261.0	264.1	267.2	270.3	273.4	276.5	279.6	282.7	285.8	288.9	292.0	295.1	298.2	301.3	304.4	308.5	310.6	313.7	316.7
17.0	252.5	255.5	258.5	261.6	264.6	267.6	270.7	273.7	276.8	279.8	282.8	285.9	288.9	291.9	295.0	298.0	302.0	304.1	307.1	310.2
18.0	247.2	250.2	253.2	256.2	259.1	262.1	265.1	268.1	271.1	274.0	277.0	280.0	283.0	286.0	288.9	291.9	295.9	297.9	300.8	303.8
19.0	242.2	245.1	248.0	251.0	253.9	256.8	259.7	262.6	265.6	268.5	271.4	274.3	277.3	280.2	283.1	286.0	289.9	291.9	294.8	297.7
20.0	237.3	240.2	243.1	245.9	248.8	251.7	254.5	257.4	260.3	263.1	266.0	268.9	271.7	274.6	277.5	280.4	284.2	286.1	289.0	291.8
21.0	232.6	235.4	238.3	241.1	243.9	246.7	249.5	252.3	255.2	258.0	260.8	263.6	266.4	269.2	272.1	274.9	278.6	280.5	283.3	286.1
22.0	228.1	230.8	233.6	236.4	239.1	241.9	244.7	247.4	250.2	253.0	255.7	258.5	261.3	264.1	266.8	269.6	273.3	275.1	277.9	280.7
23.0	223.7	226.4	229.1	231.8	234.6	237.3	240.0	242.7	245.4	248.2	250.9	253.6	256.3	259.0	261.8	264.5	268.1	269.9	272.6	275.4
24.0	219.4	222.1	224.8	227.5	230.1	232.8	235.5	238.2	240.8	243.5	246.2	248.8	251.5	254.2	256.9	259.5	263.1	264.9	267.6	270.2
25.0	215.3	218.0	220.6	223.2	225.8	228.5	231.1	233.7	236.4	239.0	241.6	244.3	246.9	249.5	252.1	254.8	258.3	260.0	262.7	265.3
26.0	211.3	213.9	216.5	219.1	221.7	224.3	226.9	229.4	232.0	234.6	237.2	239.8	242.4	245.0	247.6	250.1	253.6	255.3	257.9	260.5
27.0	207.5	210.0	212.5	215.1	217.6	220.2	222.7	225.3	227.8	230.4	232.9	235.5	238.0	240.6	243.1	245.6	249.0	250.7	253.3	255.8
28.0	203.7	206.2	208.7	211.2	213.7	216.2	218.8	221.3	223.8	226.3	228.8	231.3	233.8	236.3	238.8	241.3	244.6	246.3	248.8	251.3
29.0	200.1	202.5	205.0	207.5	209.9	212.4	214.9	217.3	219.8	222.3	224.7	227.2	229.7	232.2	234.6	237.1	240.4	242.0	244.5	247.0
30.0	196.5	199.0	201.4	203.8	206.2	208.7	211.1	213.5	216.0	218.4	220.8	223.3	225.7	228.1	230.6	233.0	236.2	237.9	240.3	242.7
31.0	193.1	195.5	197.9	200.3	202.7	205.1	207.5	209.8	212.2	214.6	217.0	219.4	221.8	224.2	226.6	229.0	232.2	233.8	236.2	238.6
32.0	189.7	192.1	194.4	196.8	199.2	201.5	203.9	206.3	208.6	211.0	213.3	215.7	218.1	220.4	222.8	225.2	228.3	229.9	232.2	234.6
33.0	186.4	188.8	191.1	193.4	195.8	198.1	200.4	202.8	205.1	207.4	209.7	212.1	214.4	216.7	219.1	221.4	224.5	226.1	228.4	230.7
34.0	183.3	185.6	187.9	190.2	192.5	194.7	197.0	199.3	201.6	203.9	206.2	208.5	210.8	213.1	215.4	217.7	220.8	222.3	224.6	226.9
35.0	180.2	182.4	184.7	187.0	189.2	191.5	193.8	196.0	198.3	200.6	202.8	205.1	207.4	209.6	211.9	214.2	217.2	218.7	221.0	223.2
36.0	177.1	179.3	181.6	183.8	186.1	188.3	190.5	192.8	195.0	197.3	199.5	201.7	204.0	206.2	208.5	210.7	213.7	215.2	217.4	219.6
37.0	174.1	176.4	178.6	180.8	183.0	185.2	187.4	189.6	191.8	194.0	196.3	198.5	200.7	202.9	205.1	207.3	210.2	211.7	213.9	216.2
38.0	171.2	173.4	175.6	177.8	180.0	182.2	184.3	186.5	188.7	190.9	193.1	195.3	197.4	199.6	201.8	204.0	206.9	208.4	210.5	212.7
39.0	168.4	170.5	172.7	174.9	177.0	179.2	181.3	183.5	185.6	187.8	190.0	192.1	194.3	196.4	198.6	200.7	203.6	205.1	207.2	209.4
40.0	165.6	167.7	169.9	172.0	174.1	176.3	178.4	180.5	182.6	184.8	186.9	189.0	191.2	193.3	195.4	197.6	200.4	201.8	204.0	206.1

¹ This table has been calculated from Eqs. (8) and (19) inserted into Eq. (18)

² This column has been calculated for the standard pressure of 101.325 kPa to give unit standard concentration, C^*

ponential term will always be approximately 1. For seawater and a depth of 10,000 m, $C_{c,h}/C_c$ is 0.97.

This means that oxygen solubility at that depth is much the same as at the surface, while the partial pressure is several times higher [Eq. (30)]. A POS responds to partial pressure; if it is used for measuring oxygen concentration at greater depth, a correction according to Eq. (30) must be made.

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