Chapter 3

DISSOLVED OXYGEN

3.1 DISSOLVED OXYGEN SATURATION

3.1.1 Introduction

Dissolved oxygen saturation, commonly symbolized as C_s and expressed in mg/l, is a basic parameter used in a great many water quality models. Since dissolved oxygen predictions are often primary reasons for developing water quality models, accurate values for C_s are needed.

Table 3-1 illustrates the equations used to calculate saturation dissolved oxygen values in a number of water quality models. The most frequently used equation is the polynomial equation developed by Elmore and Hayes (1960) for distilled water (Equation (3-1) in Table 3-1). In this equation, neither pressure nor salinity effects are considered (pressure is assumed to be 1 atm and salinity is 0 ppt).

Effects of pressure on saturation values are expressed as a ratio of site pressure to sea level (Equation (3-5)) or as a function of elevation (Equation (3-6)). Effects of salinity (relevant to estuaries and oceanic systems) are considered in the last two model equations (Equations (3-7) and (3-8)). When used in fresh water applications, the sections of the equations in which the saline term appears reduce to zero and have no effect on the dissolved oxygen saturation. Every saturation equation, whether or not modified to include non-standard pressure or salinity, evaluates dissolved oxygen saturation as a function of temperature.

TABLE 3-1. METHODS USED BY SELECTED MODELS TO PREDICT DISSOLVED OXYGEN SATURATION

Equation Number	Model Name (or Description)	Model Reference	Equation or Method for Dissolved Oxygen Saturation C_s (mg/l)
3-1	Limnological Model for Eutrophic Lakes and Impoundments	Baca and Arnett, 1976	$C_s = 14.652 - (0.41022\ T) + (0.007991\ T^2) - (7.7774 \times 10^{-5}\ T^3)$ $T = {}^{\circ}C$
3-1	EXPLORE-1	Battelle, 1973	Same as above
3-1	Level III-Receive	Medina, 1979	Same as above
3-1	Water Quality Model for Large Lakes: Part 2: Lake Erie	Di Toro and Connolly, 1980	Same as above
3-2	WRECEV	Johnson and Duke, 1976	$C_s = 14.62 - 0.3898\ T + 0.006969\ T^2 - 5.897 \times 10^{-5}\ T^3$ $T = {}^{\circ}C$
3-3	QUAL-II	Roesner, *et al.*, 1981	$C_s = 24.89 - 0.4259\ T + 0.003734\ T^2 - 1.328 \times 10^{-5}\ T^3$ $T = {}^{\circ}F$
3-4	CE-QUAL-R1	U.S. Army COE, 1982	$C_s = (14.6)e^{(-(0.027767 - 0.00027\ T + 0.000002\ T^2)\ T)}$ $T = {}^{\circ}C$
3-5	One Dimensional Steady State Stream Water Quality Model	Bauer, *et al.*, 1979	$C_s = (14.652 - .41022\ T + 0.007910\ T^2 - 7.7774 \times 10^{-5}\ T^3)\ (BP/29.92)$ $T = {}^{\circ}C$ BP = Barometric pressure (in.Hg)
3-5	HSPF (Release 7.0)	Imhoff, *et al.*, 1981	Same as above
3-6	DOSAG and DOSAG3	Duke and Masch, 1973	$(14.62 - (0.3898\ T) + (0.006969\ T^2) - (5.897 \times 10^{-5}\ T^3))$ $\left[1.0 - (6.97 \times 10^{-6}\ E)\right]^{5.167}$ $T = {}^{\circ}C$ E = Elevation, ft.
3-7	Pearl Harbor Version of Dynamic Estuary Model (DEM)	Genet *et al.*, 1974	$C_s = 14.5532 - .38217\ T + .0054258\ T^2 - CL(1.665 \times 10^{-4} - 5.866 \times 10^{-6} T + 9.796 \times 10^{-8}\ T^2)$ $T = {}^{\circ}C$ CL = Chloride concentration (ppm)
3-8	RECEIV-II	Raytheon Co., 1974, and Weiss, 1970	$C_s = 1.4277 \exp[-173.492 + 24963.39/T + 143.3483 \ln(T/100.)$ $-0.218492\ T + S(-0.033096 + 0.00014259\ T - 0.00000017\ T^2)]$ $T. = {}^{\circ}K = {}^{\circ}C + 273.15$ S = Salinity (ppt)

3.1.2 Dissolved Oxygen Saturation As Determined by the APHA

The APHA (1985) presents a tabulation of oxygen solubility in water as a function of both chlorinity and water temperature (see Table 3-2). This table is the work of Benson and Krause (1984) who collected the data and developed the equations for conditions in which the water was in contact with water saturated air at standard pressure (1.000 atm).

Since chlorinity is related to salinity, and salinity is more often measured than chlorinity, the relationship between the two quantities is of interest. The relationship, expressed here three ways, is:

$$\text{Salinity (ppt or } {}^{0}/_{00}) = 0.03 + 0.001805 \text{ Chlorinity (mg/l)} \qquad (3\text{-}9a)$$

or

$$\text{Salinity (ppt or } {}^{0}/_{00}) = 5.572 \times 10^{-4}(SC) + 2.02 \times 10^{-9}(SC)^2 \qquad (3\text{-}9b)$$

where SC = specific conductance in micromhos/cm

or

$$\text{Salinity} = 1.80655 \text{ (chlorinity as ppt)} \qquad (3\text{-}9c)$$

where chlorinity and salinity are as defined in the footnote to Table 3-2.

Equation (3-9b) is from USGS (1981) and Equation (3-9c) is from APHA (1985).

The APHA (1985) recommends that the concentration of oxygen in water (at different temperatures and salinity) at equilibrium with water saturated air be calculated using the equation below (Benson and Krause, 1984):

$$\begin{aligned} \ln C_s = & -139.34411 + (1.575701 \times 10^5/T) \\ & -(6.642308 \times 10^7/T^2) + (1.243800 \times 10^{10}/T^3) \\ & -(8.621949 \times 10^{11}/T^4) \\ & -\text{Chl}\big[(3.1929 \times 10^{-2}) - (1.9428 \times 10/T) \\ & \quad +(3.8673 \times 10^3/T^2)\big] \end{aligned} \qquad (3\text{-}10)$$

TABLE 3-2. SOLUBILITY OF OXYGEN IN WATER EXPOSED TO WATER-SATURATED AIR AT 1.000 ATMOSPHERIC PRESSURE (APHA, 1985)

Temp. in °C	Chlorinity, ppt						Difference per 0.1 ppt Chlorinity
	0.0	5.0	10.0	15.0	20.0	25.0	
	Dissolved Oxygen, mg/l						
0.0	14.621	13.728	12.888	12.097	11.355	10.657	0.016
1.0	14.216	13.356	12.545	11.783	11.066	10.392	0.015
2.0	13.829	13.000	12.218	11.483	10.790	10.139	0.015
3.0	13.460	12.660	11.906	11.195	10.526	9.897	0.014
4.0	13.107	12.335	11.607	10.920	10.273	9.664	0.014
5.0	12.770	12.024	11.320	10.656	10.031	9.441	0.013
6.0	12.447	11.727	11.046	10.404	9.799	9.228	0.013
7.0	12.139	11.442	10.783	10.162	9.576	9.023	0.012
8.0	11.843	11.169	10.531	9.930	9.362	8.826	0.012
9.0	11.559	10.907	10.290	9.707	9.156	8.636	0.012
10.0	11.288	10.656	10.058	9.493	8.959	8.454	0.011
11.0	11.027	10.415	9.835	9.287	8.769	8.279	0.011
12.0	10.777	10.183	9.621	9.089	8.586	8.111	0.011
13.0	10.537	9.961	9.416	8.899	8.411	7.949	0.010
14.0	10.306	9.747	9.218	8.716	8.242	7.792	0.010
15.0	10.084	9.541	9.027	8.540	8.079	7.642	0.010
16.0	9.870	9.344	8.844	8.370	7.922	7.496	0.009
17.0	9.665	9.153	8.667	8.207	7.770	7.356	0.009
18.0	9.467	8.969	8.497	8.049	7.624	7.221	0.009
19.0	9.276	8.792	8.333	7.896	7.483	7.090	0.009
20.0	9.092	8.621	8.174	7.749	7.346	6.964	0.009
21.0	8.915	8.456	8.021	7.607	7.214	6.842	0.008
22.0	8.743	8.297	7.873	7.470	7.087	6.723	0.008
23.0	8.578	8.143	7.730	7.337	6.963	6.609	0.008
24.0	8.418	7.994	7.591	7.208	6.844	6.498	0.008
25.0	8.263	7.850	7.457	7.083	6.728	6.390	0.007
26.0	8.113	7.711	7.327	6.962	6.615	6.285	0.007
27.0	7.968	7.575	7.201	6.845	6.506	6.184	0.007
28.0	7.827	7.444	7.079	6.731	6.400	6.085	0.007
29.0	7.691	7.317	6.961	6.621	6.297	5.990	0.007
30.0	7.559	7.194	6.845	6.513	6.197	5.896	0.007
31.0	7.430	7.073	6.733	6.409	6.100	5.806	0.006
32.0	7.305	6.957	6.624	6.307	6.005	5.717	0.006
33.0	7.183	6.843	6.518	6.208	5.912	5.631	0.006
34.0	7.065	6.732	6.415	6.111	5.822	5.546	0.006
35.0	6.950	6.624	6.314	6.017	5.734	5.464	0.006
36.0	6.837	6.519	6.215	5.925	5.648	5.384	0.006
37.0	6.727	6.416	6.119	5.835	5.564	5.305	0.006
38.0	6.620	6.316	6.025	5.747	5.481	5.228	0.006
39.0	6.515	6.217	5.932	5.660	5.400	5.152	0.005
40.0	6.412	6.121	5.842	5.576	5.321	5.078	0.005

DEFINITION OF SALINITY

Although salinity has been traditionally defined as the total solids in water after all carbonates have been converted to oxides, all bromide and iodide have been replaced by chloride, and all organic matter has been oxidized, the new scale used to define salinity is based on the electrical conductivity of seawater relative to a specified solution of KCl and H_2O (UNESCO, 1981). The scale is dimensionless and the traditional dimensions of parts per thousand (i.e., mg/g of solution) no longer applies.

DEFINITION OF CHLORINITY

Chlorinity is now defined in relation to salinity as follows:

Salinity = 1.80655 (Chlorinity)

Although chlorinity is not equivalent to chloride concentration, the factor for translating a chloride determination in seawater to include bromide, for example, is only 1.0045 based on the molecular weights and the relative amounts of the two ions. Therefore, for practical purposes, chloride (in mg/g of solution) is nearly equal to chlorinity in seawater. For wastewater, a knowledge of the ions responsible for the solution's electrical conductivity is necessary to correct for the ions impact on oxygen solubility and use of the tabular value or the equation is inappropriate unless the relative composition of the wastewater is similar to seawater.

where C_s = equilibrium oxygen concentration, mg/l, at 1.000 atm (standard pressure)
T = temperature (^oK) = $^oC+273.150$ and ^oC is within 0.0 to 40.0^oC
Chl = chlorinity within 0.0 to 28.0, ppt

Table 3-2 replaces the older table of previous APHA Standard Methods editions. The USGS (1981) has replaced older tables based on calculations of Whipple and Whipple (1911) with tables generated from an equation by Weiss (1970) (Equation 3-8).

The APHA (1985) recommends that saturation dissolved oxygen concentration at non-standard pressure be calculated using the following equation:

$$C_s' = C_s P \left[\frac{(1-P_{wv}/P)\ (1-\theta P)}{(1-P_{wv})\ (1-\theta)} \right] \qquad (3\text{-}11)$$

where C_s' = equilibrium oxygen concentration at non-standard pressure, mg/l
C_s = equilibrium oxygen concentration at 1.000 atm, mg/l
P = pressure, atm, and is within 0.000 to 2.000 atm
P_{wv} = partial pressure of water vapor, atm, which may be computed
$\ln P_{wv} = 11.8571 - (3840.70/T_k) - (216961/T_k^2)$
T_k = temperature in ^oK
$\theta = 0.000975 - (1.426 \times 10^{-5} T_c) + (6.436 \times 10^{-8} T_c^2)$
T_c = temperature in ^oC

The expressions for P_{wv} and θ are also from APHA (1985).

For altitudes less than approximately 4000 feet the bracketed quantity is very nearly 1 and at these altitudes multiplying C_s by P(atm) alone results in a good approximation of C_s'. A more accurate calculation of C_s'

can be made by using Table 3-3. The quantity in brackets from Equation (3-11) is tabulated for temperatures between 0-40^{o}C and for pressures from 1.1 to 0.5 atm (Benson and Krause, 1980). As an approximation of the influence of altitude, C_s decreases about 7 percent per 2,000 feet of elevation increase.

TABLE 3-3 VALUES FOR THE BRACKETED QUANTITY SHOWN IN EQUATION 3-11 TO BE USED WITH THE CORRESPONDING TEMPERATURES AND PRESSURES (BENSON AND KRAUSE, 1980)

	P atm						
T(^{o}C)	1.1	1.0	0.9	0.8	0.7	0.6	0.5
0.0	1.0005	1.0000	0.9994	0.9987	0.9977	0.9963	0.9944
5.0	1.0007	1.0000	0.9991	0.9980	0.9966	0.9946	0.9918
10.0	1.0010	1.0000	0.9987	0.9971	0.9950	0.9922	0.9882
15.0	1.0015	1.0000	0.9982	0.9959	0.9929	(0.9889)	(0.9833)
20.0	1.0021	1.0000	0.9974	0.9942	(0.9901)	(0.9845)	[0.9767]
25.0	1.0029	1.0000	0.9965	0.9921	(0.9864)	(0.9787)	[0.9680]
30.0	1.0039	1.0000	0.9952	(0.9892)	(0.9814)	[0.9711]	[0.9566]
35.0	1.0053	1.0000	(0.9935)	(0.9854)	(0.9750)	[0.9610]	[0.9415]
40.0	1.0071	1.0000	(0.9913	(0.9805)	[0.9665]	[0.9479]	[0.9217]

Explanation of Interpolation Procedure:

Linear interpolation in P and T will introduce an error ≤0.02% in the upper and left sections of table. Interpolation using numbers in parentheses will lead to errors ≤0.05%. With the numbers in brackets, interpolation errors become larger. Either temperature or presssure may be interpolated first, as illustrated for T = 3.00^{o}C and P = 0.67 atm by the two arrays shown below.

Temperature Interpolated First

	0.7	0.67	0.6
0	0.9977		0.9963
3	0.99704	0.9965 answer	0.99528
5	0.9966		0.9946

Pressure Interpolated First

	0.7	0.67	0.6
0	0.9977	0.99728	0.9963
3		0.9965, answer	
5	0.9966	0.99600	0.9946

Earlier the APHA (1980) calculated the effects of barometric pressure on dissolved oxygen saturation as:

$$C_s' = C_s \left(\frac{P - P_{wv}}{1 - P_{wv}} \right)$$

This is equivalent to Equation (3-11) when $\theta = 0$.

3.1.3 Comparison of Methods

Table 3-4 compares the dissolved oxygen saturation values for Equations (3-1) through (3-8) and APHA (1971) against the values in Table 3-2 from the APHA (1985), Equation (3-10). The comparisons are performed at 0.0 mg/l salinity and sea level. When the values from the equations are compared with the APHA (1985) values within the temperature range 10-30°C* and the maximum differences examined, four "groups" of differences appear. Values from Equation (3-8) are in the group that shows the least difference from APHA (1985): 0.03 mg/l higher than the APHA (1985) predictions. Values from Equations (3-2), (3-4), (3-6) and APHA (1971) are in the second group with differences of .07 to .11 mg/l higher than APHA (1985). Values from Equations (3-1), (3-3) and (3-5) are in the third group with differences of .11 to .13 mg/l lower than APHA (1985). Equation (3-7) produced differences that comprise the fourth group with some values >0.4 mg/l higher than APHA (1985). Generally, the maximum differences with each equation occur at higher temperatures, when dissolved oxygen depletion may contribute to serious water quality problems.

In Table 3-5 Equations (3-7), (3-8), (3-13) and APHA (1971) (those including salinity factors) are evaluated at a chloride concentration of 20,000 mg/l at 1 atm pressure and compared to APHA (1985) values.

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* Typically, the temperature range in which most freshwater water quality analyses take place.

TABLE 3-4. COMPARISON OF DISSOLVED OXYGEN SATURATION VALUES FROM TEN EQUATIONS AT 0.0 mg/l SALINITY AND 1 ATM PRESSURE

Temperature	Equation Number From Table 3-1								APHA	APHA(1985)
°C	(3-1)	(3-2)	(3-3)	(3-4)	(3-5)	(3-6)	(3-7)	(3-8)	(1971)	3-10
0.0	14.652	14.620	14.650	14.600	14.652	14.620	14.553	14.591	14.6	14.621
1.0	14.250	14.237	14.248	14.204	14.250	14.237	14.176	14.188	14.2	14.216
2.0	13.863	13.868	13.861	13.826	13.863	13.868	13.811	13.803	13.8	13.829
3.0	13.491	13.512	13.490	13.465	13.491	13.512	13.456	13.435	13.5	13.460
4.0	13.134	13.169	13.133	13.120	13.134	13.169	13.111	13.084	13.1	13.107
5.0	12.791	12.838	12.790	12.790	12.791	12.838	12.778	12.748	12.8	12.770
6.0	12.462	12.519	12.460	12.475	12.462	12.519	12.456	12.426	12.5	12.447
7.0	12.145	12.213	12.144	12.173	12.145	12.213	12.144	12.118	12.2	12.139
8.0	11.842	11.917	11.841	11.883	11.842	11.917	11.843	11.823	11.9	11.843
9.0	11.551	11.633	11.550	11.606	11.551	11.633	11.553	11.540	11.6	11.559
10.0	11.271	11.360	11.270	11.340	11.271	11.360	11.274	11.268	11.3	11.288
11.0	11.003	11.097	11.002	11.085	11.003	11.097	11.006	11.008	11.1	11.027
12.0	10.746	10.844	10.744	10.840	10.746	10.844	10.748	10.758	10.8	10.777
13.0	10.499	10.601	10.497	10.605	10.499	10.601	10.502	10.517	10.6	10.537
14.0	10.262	10.367	10.260	10.378	10.262	10.367	10.266	10.286	10.4	10.306
15.0	10.034	10.142	10.033	10.161	10.034	10.142	10.041	10.064	10.2	10.084
16.0	9.816	9.926	9.814	9.951	9.816	9.926	9.827	9.850	10.0	9.870
17.0	9.606	9.718	9.604	9.749	9.606	9.718	9.624	9.644	9.7	9.665
18.0	9.404	9.518	9.401	9.555	9.404	9.518	9.432	9.446	9.5	9.467
19.0	9.209	9.325	9.207	9.367	9.209	9.325	9.251	9.254	9.4	9.276
20.0	9.022	9.140	9.019	9.186	9.022	9.140	9.080	9.070	9.2	9.092
21.0	8.841	8.961	8.838	9.011	8.841	8.961	8.920	8.891	9.0	8.915
22.0	8.667	8.789	8.664	8.842	8.667	8.789	8.772	8.720	8.8	8.743
23.0	8.498	8.624	8.495	8.679	8.498	8.624	8.634	8.554	8.7	8.578
24.0	8.334	8.464	8.331	8.521	8.334	8.464	8.506	8.393	8.5	8.418
25.0	8.176	8.309	8.172	8.367	8.176	8.309	8.390	8.238	8.4	8.263
26.0	8.021	8.160	8.017	8.219	8.021	8.160	8.285	8.088	8.2	8.113
27.0	7.871	8.015	7.866	8.075	7.871	8.015	8.190	7.943	8.1	7.968
28.0	7.723	7.875	7.719	7.935	7.723	7.875	8.106	7.802	7.9	7.827
29.0	7.579	7.739	7.574	7.800	7.579	7.739	8.033	7.666	7.8	7.691
30.0	7.437	7.606	7.432	7.668	7.437	7.606	7.971	7.533	7.6	7.559
31.0	7.298	7.477	7.292	7.539	7.298	7.477	7.920	7.405	7.5	7.430
32.0	7.159	7.350	7.154	7.414	7.159	7.350	7.880	7.281	7.4	7.305
33.0	7.022	7.227	7.016	7.293	7.022	7.227	7.850	7.161	7.3	7.183
34.0	6.885	7.105	6.880	7.174	6.885	7.105	7.832	7.043	7.2	7.065
35.0	6.749	6.986	6.743	7.058	6.749	6.986	7.824	6.930	7.1	6.950
36.0	6.612	6.868	6.606	6.945	6.612	6.868	7.827	6.819	-	6.837
37.0	6.474	6.751	6.468	6.834	6.474	6.751	7.841	6.711	-	6.727
38.0	6.335	6.635	6.329	6.726	6.335	6.635	7.866	6.606	-	6.620
39.0	6.194	6.520	6.188	6.620	6.194	6.520	7.901	6.505	-	6.315
40.0	6.051	6.404	6.045	6.517	6.051	6.404	7.948	6.405	-	6.412

Equation (3-13) is based on the data of Green and Carritt (1967). From their data Hyer et al. (1971) developed an expression relating C_s to both temperature and salinity. C_s is given by:

$$C_s = 14.6244 - 0.367134T + 0.0044972T^2$$

$$- 0.0966S + 0.00205ST + 0.0002739S^2 \qquad (3\text{-}13)$$

TABLE 3-5. COMPARISON OF DISSOLVED OXYGEN SATURATION VALUES FROM SELECTED EQUATIONS AT A CHLORIDE CONCENTRATION OF 20,000 mg/l (36.1 ppt SALINITY) AND 1 ATM PRESSURE

Temperature	Equation Number from Table 3-1			APHA	APHA (1985)
°C	(3-7)	(3-8)	(3-13)	(1971)	(3-10)
0.0	11.215	11.400	11.492	11.3	11.354
1.0	10.953	11.105	11.203	11.0	11.067
2.0	10.699	10.823	10.924	10.8	10.790
3.0	10.452	10.553	10.653	10.5	10.527
4.0	10.212	10.295	10.391	10.3	10.273
5.0	9.978	10.048	10.139	10.0	10.031
6.0	9.752	9.811	9.895	9.8	9.801
7.0	9.532	9.585	9.661	9.6	9.575
8.0	9.320	9.367	9.435	9.4	9.362
9.0	9.114	9.158	9.218	9.2	9.156
10.0	8.915	8.958	9.011	9.0	8.957
11.0	8.723	8.765	8.812	8.8	8.769
12.0	8.538	8.580	8.623	8.6	8.586
13.0	8.360	8.402	8.442	8.5	8.411
14.0	8.189	8.231	8.270	8.3	8.241
15.0	8.025	8.067	8.108	8.1	8.077
16.0	7.868	7.908	7.954	8.0	7.922
17.0	7.718	7.755	7.809	7.8	7.770
18.0	7.574	7.607	7.674	7.7	7.624
19.0	7.438	7.465	7.547	7.6	7.482
20.0	7.308	7.327	7.429	7.4	7.347
21.0	7.186	7.194	7.321	7.3	7.215
22.0	7.070	7.066	7.221	7.1	7.087
23.0	6.961	6.942	7.130	7.0	6.964
24.0	6.859	6.822	7.049	6.9	6.844
25.0	6.764	6.594	6.976	6.7	6.727
26.0	6.676	6.594	6.912	6.6	6.616
27.0	6.595	6.485	6.857	6.5	6.507
28.0	6.521	6.379	6.812	6.4	6.401
29.0	6.454	6.277	6.775	6.3	6.297
30.0	6.394	6.177	6.747	6.1	6.197
31.0	6.340	6.081	6.729	-	6.100
32.0	6.294	5.987	6.719	-	6.005
33.0	6.254	5.896	6.718	-	5.912
34.0	6.221	5.808	6.726	-	5.822
35.0	6.196	5.722	6.743	-	5.734
36.0	6.177	5.638	6.770	-	5.648
37.0	6.165	5.557	6.805	-	5.564
38.0	6.160	5.477	6.849	-	5.481
39.0	6.162	5.400	6.902	-	5.400
40.0	6.171	5.325	6.965	-	5.322

where T = temperature, ^{o}C
S = salinity, ppt.

The values were compared over a temperature range of 5-30^{o}C. Equation (3-8), as before, agreed closely with APHA (1985) throughout the 5-30^{o}C temperature range with a maximum difference of .022 mg/l less than APHA (1985). Equation (3-7) had differences of .08 less than and .04 mg/l greater than APHA (1985) from 5-25^{o}C and near .2 mg/l higher than APHA (1985) at 30^{o}C. The values from the APHA (1971) (reported to the nearest tenth mg/l) had a maximum difference range of 0 to .1 mg/l higher than APHA (1985) and the fourth equation, Equation (3-13), varied the most from APHA (1985) with differences in the range of approximately .03 to 0.5 mg/l higher.

3.1.4 Methods of Measurement

Elmore and Hayes (1960) have summarized the work of numerous researchers who have measured dissolved oxygen saturation. According to Elmore and Hayes, Fox in 1909 used a gasometric technique in which a known volume of pure oxygen was exposed to a known volume of water. After equilibrium had been established the volume of oxygen above the water was determined, and the solubility calculated assuming air contained 20.90 percent oxygen.

From Fox's expression, Whipple and Whipple (1911) converted their results from milliliters per liter to parts per million. These results were tabularized, circulated and used as standards by water agencies for years, and are only now being gradually replaced with tables developed from more elaborate equations.

Benson and Krause (1984) determined the solubility of oxygen in fresh and seawater over a temperature of 0-60^{o}C using an equilibrator different from the Jacobsen Worthington-type equilibrator used in previous investigations. They felt the new apparatus minimized the uncertainties associated with methods involving thin films of liquids (Benson, et al., 1979). The dissolved gas values were determined with use of a mercury manometric system. The resulting data and equations were compared to previous sets of values from Carpenter (1966), Green (1965), and Murray and Riley (1969). The APHA (1985) subsequently adopted the Benson and Krause concentrations as tabulated in Table 3-2. In earlier work involving fresh water only (Benson and Krause, 1980) the new concentration values were recommended by Mortimer (1981) for use in fresh water systems.

To date there is no "standard method" recommended by APHA to measure saturated dissolved oxygen. The laboratory methods noted in the preceeding paragraphs are sophisticated methods developed and/or modified for each research effort and are not conducive to simplier laboratory environments nor are they adaptable for field use.

Calibration of popular dissolved oxygen probes is carried out under saturation conditions by methods recommended by the instrument manufacturers in conjunction with a table such as Table 3-2. The values obtained may be verified with one of the several wet chemistry iodometric methods (or "Winkler" titrations) (APHA, 1985).

3.1.5 Summary

Notable differences exist among the results obtained by various methods used to determine saturated dissolved oxygen values under specified conditions of temperature, salinity and pressure. These discrepancies may be as high as 11 percent for high saline conditions (Table 3-5). Under conditions of zero salinity observed differences are generally less than 2 percent (Table 3-4). The accuracy of the Elmore and Hayes expression, one of the most frequently used formulas, rapidly deteriorates at water temperatures exceeding 25°C. The algorithm, Equation (3-8), used in the RECEIV-II model (Weiss, 1970 and USGS, 1981) matches the APHA (1985) data better than any formula reviewed, for both saline and freshwater conditions. The algorithm, Equation (3-10), presented in APHA (1985) and its corresponding table of saturation values, Table 3-2, are based on latest research and provide the most accurate values of C_s to date. Knowing the possible sources of error using any other particular formulation for C_s permits the user to decide whether they are significant in a particular study.

3.2 REAERATION

3.2.1 Introduction

Reaeration is the process of oxygen exchange between the atmosphere and a water body in contact with the atmosphere. Typically, the net transfer of oxygen is from the atmosphere and into the water, since dissolved oxygen levels in most natural waters are below saturation. However, when photosynthesis produces supersaturated dissolved oxygen levels, the net transfer is back into the atmosphere.

The reaeration process is modeled as the product of a mass-transfer coefficient multiplied by the difference between dissolved oxygen saturation and the actual dissolved oxygen concentration, that is:

$$F_c = k_L(C_s-C) \tag{3-14}$$

where F_c = flux of dissolved oxygen across the water surface, mass/area/time

C = dissolved oxygen concentration, mass/volume

C_s = saturation dissolved oxygen concentration, mass/volume

k_L = surface transfer coefficient, length/time

For practically all river modeling applications and for vertically mixed estuaries a depth averaged flux (F_c'), is used:

$$F_c' = \frac{F_c}{H} = \frac{k_L}{H}(C_s-C) \tag{3-15}$$

where H = water depth, length

In Equation 3-15 the surface transfer rate and depth are typically combined into a single term, called the reaeration rate coefficient or reaeration coefficient, denoted in the literature by k_2 or k_a:

$$k_2 = \frac{k_L}{H} \tag{3-16}$$

3.2.2 Reaeration in Rivers

3.2.2.1 Overview

Rivers have been the focus of the majority of reaeration research in natural waters. Some of the equations that have been developed for rivers have been successfully applied to estuaries, and is indicative of the lack of estuarine reaeration research.

Table 3-6 summarizes reaeration coefficient expressions (k_2 values) for rivers. All formulas for reaeration in Table 3-6 are depth averaged values and are in units of 1/day. The table also shows the units required for the parameters in each formula, and when possible the range of conditions used in the development of the formulas. All values of k_2 are base e, and are referenced to 20°C, unless otherwise noted. Although base e values are used directly in most modeling formulations, in the earlier days of reaeration research, k_2 values were often expressed in base 10. The relationship between base e and base 10 reaeration coefficients is:

$$k_{2_{\text{base e}}} = \ln (10) k_{2_{\text{base 10}}} = 2.303\ k_{2_{\text{base 10}}} \tag{3-17}$$

Stream reaeration research began in earnest in the late 1950's, and continues today. The formulas that are shown in Table 3-6 are based on theory, empiricism, or a combination of the two. In the late 1960's the radioactive tracer method was introduced by Tsivoglou and Wallace (1972). The tracer method, or a modification of it, forms the basis for much of the research being conducted on reaeration today.

TABLE 3-6. REAERATION COEFFICIENTS FOR RIVERS AND STREAMS

Author(s)	k_2, base e(1/day at 20°C)	Units	Applicability
O'Connor and Dobbins (1958)	$\frac{12.9U^{0.5}}{H^{1.5}}$	U-fps H-feet	Moderately deep to deep channels; $1' \leq H \leq 30'$, $0.5 \leq U \leq 1.6$ fps; $0.05 \leq k_2 \leq 12.2$/day. O'Connor and Dobbins also developed a second formula for shallow streams but O'Connor (1958) showed the differences between the two formulas was insignificant, and recommended that the first formula be used.
Churchill *et al*. (1962)	$\frac{11.6U^{0.969}}{H^{1.673}}$	U-fps H-feet	Based on observed reaeration rates below dams from which oxygen deficient water was released. $2' \leq H \leq 11'$; $1.8 \leq U \leq 5$ fps. Churchill *et al*. also developed other formulas, but recommended this formula.
Owens *et al*. (1964)	$\frac{21.7U^{0.67}}{H^{1.85}}$	U-fps H-feet	Oxygen recovery monitored for six streams in England following deoxygenation with sodium sulfite. $0.1 \leq U \leq 5$ fps; $0.4' \leq H \leq 11'$.
Owens *et al*. (1964)	$\frac{23.3U^{0.73}}{H^{1.75}}$	U-fps H-feet	This is a second formula developed by Owens *et al*., and applies for $0.1 \leq U \leq 1.8$ fps; $0.4' \leq H$ 11'.
Langbein and Durum (1967)	$\frac{7.6U}{H^{1.33}}$	U-fps H-feet	Based on synthesis of data from O'Connor-Dobbins (1958), Churchill *et al*. (1962), Krenkel and Orlob (1963), and Streeter *et al*. (1936).
Isaacs and Gaudy (1968)	$\frac{8.62U}{H^{1.5}}$	U-fps H-feet	Developed using regression analyses from data collected using a recirculating cylindrical tank. $0.6 \leq U \leq 1.6$ fps; $0.5' \leq H \leq 1.5'$.
Parkhurst and Pomeroy (1972)	$\frac{48.4(1+0.17F^2)(SU)^{3/8}}{H}$	U-m/s S-m/m H-meters	Developed from data collected in 12 sewers and in natural streams.
Negulescu and Rojanski (1969)	$10.9\left(\frac{U}{H}\right)^{0.85}$	U-fps H-feet	Developed from a recirculating flume with depths less than 0.5 feet.
Thackston and Krenkel (1969)	$\frac{24.9\left(1+F^{0.5}\right)u_*}{H}$	u_*-fps H-feet	Based on measurements made in a 2' wide flume with deoxygenated waters. $0.05' \leq H \leq 0.23'$.
Lau (1972b)	$2515\left(\frac{u_*}{u}\right)^3\frac{U}{H}$	u_*-fps U -fps H-feet	Based on reanalysis of the data of Thackston and Krenkel (1969), Krenkel (1960), and Churchill *et al*. (1962).

(continued)

TABLE 3-6. (continued)

Author(s)	k_2, base e(1/day at 20°C)	Units	Applicability
Krenkel and Orlob (1962)	$\frac{234(US)^{.408}}{H^{.66}}$	U-fps S-ft/ft H-feet	Based on 1' wide flume data. 0.08' ≤ H ≤ 0.2'
Krenkel and Orlob (1962)	$\frac{8.4\ D_L^{1.321}}{H^{2.32}}$	D_L-ft²/min H-feet	Experiments performed in a 1' wide flume by deoxygenating the water. Other similar formulas are also reported. The flume dispersion coefficient, D_L, was below the range expected in natural systems.
Padden and Gloyna (1971)	$\frac{6.9U^{0.703}}{H^{1.054}}$	U-fps H-feet	Regression analysis performed on data where $9.8 \le k_2 \le 28.8$/day.
Cadwallader and McDonnell (1969)	$\frac{336(US)^{0.5}}{H}$	U-fps S-ft/ft H-feet	Based on multivariate analysis of reaeration data.
Bansal (1973)	$\frac{4.67U^{0.6}}{H^{1.4}}$	U-fps H-feet	Based on reanalysis of reaeration data in numerous rivers.
Bennett and Rathbun (1972)	$\frac{106U^{0.413}S^{0.273}}{H^{1.408}}$	U-fps S-ft/ft H-feet	These two equations are based on a reanalysis of historical data, with the second equation being al most as good a predictor as the first, but not having the slope term.
	$\frac{20.2U^{0.607}}{H^{1.689}}$		
Dobbins (1964)	$\frac{117\left(1+F^2(US)^{0.375}\right)}{(0.9+F)^{1.5}H} \coth\left[\frac{4.10(US)^{0.125}}{(0.9+F)^{0.5}}\right]$	U-fps H-feet S-ft/ft	Theory combined with measurements in natural streams, and flume data of Krenkel and Orlob (1963).
Ice and Brown (1978)	$\frac{37W^{2/3}S^{1/2}U^{7/6}g^{1/2}}{Q^{2/3}}$	W-feet S-ft/ft U-fps Q-ft³/sec	Based on data collected in several small Oregon streams.
McCutcheon and Jennings (1982)	$\frac{-\ln\left[1-2\left(\frac{\alpha I\ 24}{\pi(30.48H)^2}\right)^{1/2}\right]}{I}$	H-feet $\alpha = 1.42\ (1.1)^{T-20}$ T-°C	Based on the Velz method (1970) and replaces the iterative technique. The expressions for the mix internal I are based on an accumulation of applications of the Velz technique.
	$I = \begin{cases} 0.0016 + 0.0005\ H & H \le 2.26\ \text{ft} \\ 0.0097\ \ln(H) - 0.0052 & H > 2.26\ \text{ft} \end{cases}$		

(continued)

TABLE 3-6. (continued)

Author(s)	k_2, base e(1/day at 20°C)	Units	Applicability
Long (1984)	$\frac{1.923U^{0.273}}{H^{0.894}}$	U-meters/sec H-meters	Known as the "Texas" equation. Based on data collected on streams in Texas.
Foree (1976)	$0.30+0.19S^{1/2}$ at 25°C	S-feet/mile	Radioactive tracer technique used on small streams in Kentucky. $0 \le S \le 42$ feet/mile.
Foree (1977)	• $0.888\ (0.63+0.4S^{1.15})q^{0.25}$ at 25°C for $0.05 \le q \le 1$	S-feet/mile q-cfs/mi^2	Reanalysis of Foree's (1976) data.
	• $0.888\ (0.63+0.4\ S^{1.5})$ at 25°C for $q > 1$		
	• $0.42\ (0.63+0.4S^{1.15})$ at 25°C for $q < 0.05$		
Tsivoglou and Wallace (1972)	$0.054\ \frac{\Delta h}{t}$ at 25°C	Δh-feet t-days	Based on summary of radioactive tracer applications to 5 rivers.
Tsivoglou and Neal (1976)	• $0.11 \left(\frac{\Delta h}{t}\right)$ for $1 \le Q \le 10$ cfs	Δh-feet t-days	Based on data collected on 24 different streams using radioactive tracer method.
	• $0.054 \left(\frac{\Delta h}{t}\right)$ for $25 < Q \le 3000$ cfs		
Grant (1976)	$0.09\left(\frac{\Delta h}{t}\right)$ at 25°C	Δh-feet t-days	Based on data from 10 small streams in Wisconsin using radioactive tracer techniques: $2.1 \le k_2 \le 55$/day $1.2 \le S \le 70$ ft/mile $0.3 \le Q \le 37$ cfs
Grant (1978)	$0.06\left(\frac{\Delta h}{t}\right)$ at 25°C	Δh-feet t-days	Based on radioactive tracer data developed on Rock River, Wisconsin and Illinois: $0.01 \le k_2 \le 0.8$/day $0.25 \le U \le 1.6$ fps $0.2 \le S \le 3.5$ ft/mile $260 \le Q \le 1030$ cfs

(continued)

TABLE 3-6. (continued)

Author(s)	k_2, base e(1/day at 20°C)	Units	Applicability
Shindala and Truax (1980)	• $0.08\left(\frac{\Delta h}{t}\right)$ at 25°C for $Q \le 10$ cfs • $0.06\left(\frac{\Delta h}{t}\right)$ at 25°C for $10 < Q \le 280$ cfs	Δh-feet t-days	Based on statistical analysis of reaeration coefficients for rivers in 7 states, where the radioactive tracer method was used to find the reaeration rates.
Eloubaidy and Plate (1972)	Wind effects analyzed. See text for discussion.		
Mattingly (1977)	Wind effects analyzed. See text for discussion.		
Gulliver and Stefan (1981)	Wind effects analyzed.		
Frexes *et al*. (1984)	Wind effects analyzed.		

Definitions of Symbols:

D_L = longitudinal dispersion coefficient

F = Froude number

$= \frac{U}{(gh)^{0.5}}$

g = acceleration due to gravity

Δh = change in stream bed elevation between two points

q = stream discharge divided by drainage area

R = hydraulic radius

S = slope

t = travel time between two points where Δh measured

U = stream velocity

u_* = shear velocity = $\sqrt{gRS}$

W = width

3.2.2.2 Reviews of Stream Reaeration

Over the past decade, several researchers have reviewed reaeration formulas, and have tried to evaluate the performance of the formulas. One of the earlier reviews, Bennett and Rathbun (1972), is also an excellent source for reaeration theory. They describe the theories behind various conceptual models of reaeration (including film, renewal, penetration, film-penetration, and two-film theory models), semi-empirical models, and empirical models. They also discuss methods to determine the reaeration coefficient that include dissolved oxygen balances in natural streams, dissolved oxygen balances in recirculating flumes, the distributed equilibrium technique (where sodium sulfite is usually added to the water to deoxygenate it), and the radioactive tracer technique.

Table 3-7 summarizes the Bennett and Rathbun review in addition to other studies that have compared reaeration coefficients. The studies conclude that no single formula is best for all rivers. For one set of data one formula may be best, while for another set of data another formula may appear to be best.

Figure 3-1 compares 13 reaeration coefficient expressions for a range of depths (from Bennett and Rathbun, 1972). The figure illustrates the variability between predictions for a velocity of 1.0 fps and slope of 0.0001. The range of differences between predicted values spans one to two orders of magnitude. The formulas agree with each other best within the depth range of 1 to 10 feet, typical of many rivers.

Figure 3-2 compares calculated and observed reaeration coefficients for the formulas of Dobbins (1965) and Parkhurst and Pomeroy (1972). These formula were found by Wilson and MacLeod (1974) to best fit the observed data. Notice that the spread of data is slightly less than one order of magnitude.

The data of Wilson and Macleod also show that the depth - velocity model of Bennet and Rathbun (1972) does not fit the experimental data nearly

TABLE 3-7. SUMMARY OF STUDIES WHICH REVIEWED STREAM REAERATION COEFFICIENTS

Bennett and Rathbun (1972)

- Thirteen equations were evaluated.
- The standard error of the estimate was used as a measure of the difference between predicted values and data.
- The equation which provided the best fit to their original data set was Krenkel (1960).
- The equations which best fit the entire range of data were: O'Connor and Dobbins (1958), Dobbins (1965), Thackston and Krenkel (1969).
- Of the thirteen equations the Churchill et al. (1962) formula provided the best fit to natural stream data.
- The Bennett and Rathbun formula, developed from the data evaluated during their review, provided a smaller standard error for natural streams than the other 13 equations.
- There was a significant difference between predictions from equations derived from flume data and equations derived from natural stream data.
- The expected root-mean-square error from different measurement techniques is: 15 percent using the radioactive tracer technique; 65 percent using the dissolved oxygen mass balance, and 115 percent using the disturbed equilibrium method.

Lau (1972b)

- Both conceptual and empirical models were reviewed.
- Conclusions reported were similar to those of Bennett and Rathbun.
- It was found that no completely satisfactory method exists to predict reaeration.

Wilson and MacLeod (1974)

- Nearly 400 data points were used in the analysis.
- Sixteen equations were reviewed.
- The standard error of estimate and graphical results were both used in error analysis.
- It was concluded that equations which use only depth and velocity are not accurate over the entire range of data investigated.
- The methods of Dobbins (1965) and Parkhurst and Pomeroy (1972) gave the best fits to the data investigated.

Rathbun (1977)

- Nineteen equations were reviewed.
- Equation predictions were compared against radioactive tracer measurements on 5 rivers (Chattahoochee, Jackson, Flint, South, Patuxent).
- The best equations in terms of the smallest standard error estimates was Tsivoglou- Wallace (1972) (0.0528), Parkhurst-Pomeroy (1972) (0.0818), Padden-Gloyna (1971) (0.0712) and Owens et al. (1964), (0.0964).
- No one formula was best for all five rivers.

Rathbun and Grant (1978)

- Compared the radioactive and modified tracer techniques for Black Earth Creek and Madison Effluent Channel in Wisconsin.
- Differences in Black Earth Creek were -9% to 4% in one reach and 16% to 32% on another reach attributable to increased wind during the latter part of the test.
- Unsteady flow during the Madison Effluent Channel tests led to differences of as much as 25 to 58% in one case and -5% to 3% in another.

Shindala and Truax (1980)

- Reaeration measurements for streams in Mississippi, Wisconsin, Texas, Georgia, North Carolina, Kentucky, and New York were made using the radioactive tracer technique.
- The energy dissipation model resulted in the best correlation for reaeration coefficient prediction for small streams. The following escape coefficients (defined as the coefficients of $\frac{\Delta h}{t}$ in energy dissipation models for reaeration coefficients) were recommended:

  0.0802/ft ,for $Q < 10$ cfs
  0.0597/ft ,for $10 \leq Q \leq 280$ cfs

NCASI Bulletin (1982b)

- Six reaeration formulas were compared against measurements made using radioactive tracer techniques and hydrocarbon tracer techniques for a reach of the Ouachita River, Arkansas.
- The hydrocarbon tracer technique produced reaeration rates higher than both the radioactive tracer and empirical formulas.
- The O'Connor - Dobbins (1958) equation was chosen as the best empirical equation.

Kwasnik and Feng (1979)

- Thirteen reaeration formulas were reviewed and compared against values measured using the modified tracer technique for two streams in Massachusetts.
- The equations of Tsivoglou-Wallace (1972) and Bennett-Rathbun (1972) gave the closest predictions to the field values.
- The study indicates that results using the modified tracer technique are reproducible.

Grant and Skavroneck (1980)

- Four modified tracer methods and 20 predictive equations were compared against the radioactive tracer methods for 3 small streams in Wisconsin.
- Compared to the radioactive tracer method the errors in the modified tracer techniques were:

  11% for the propane-area method
  18% for the propane-peak method
  21% for the ethylene-peak method
  26% for the ethylene-area method

- Compared to the radioactive tracer method, the equations with the smallest errors were:

  18% for Tsivoglou-Neal (1976)
  21% for Negulescu-Rojanski (1969)
  23% for Padden-Gloyna (1971)
  29% for Thackston-Krenkel (1969)
  32% for Bansal (1973)

TABLE 3-7. (continued)

House and Skavroneck (1981)

- Reaeration coefficients were determined on two creeks in Wisconsin using the propane - area modified tracer technique and compared against 20 predictive formulas.

- The top five predictive formulas were:

  Tsivoglou - Neal (1976), 34% mean error
  Foree (1977), 35% mean error
  Cadwallader and McDonnell (1969), 45%, mean error
  Isaacs-Gaudy (1968), 45%, mean error
  Langbein-Durum (1967), 49%, mean error.

Zison *et al*. (1978)

- Thirteen reaeration formulas were reviewed, but none were compared against historical data.

- Covar's method (1976) was discussed which shows how stream reaeration can be simulated by using three formulas (O'Connor-Dobbins (1958), Churchill *et al*. (1962), and Owens *et al*. (1964)), each applicable in a different depth and velocity regime.

Yotsukura *et al*. (1983)

- Developed a steady injection method to avoid uncertainty in dispersion corrections.

- Determined reproducibility to be 4%.

- Found negligible effect of wind where stream banks are high.

Ohio Environmental Protection Agency (1983)

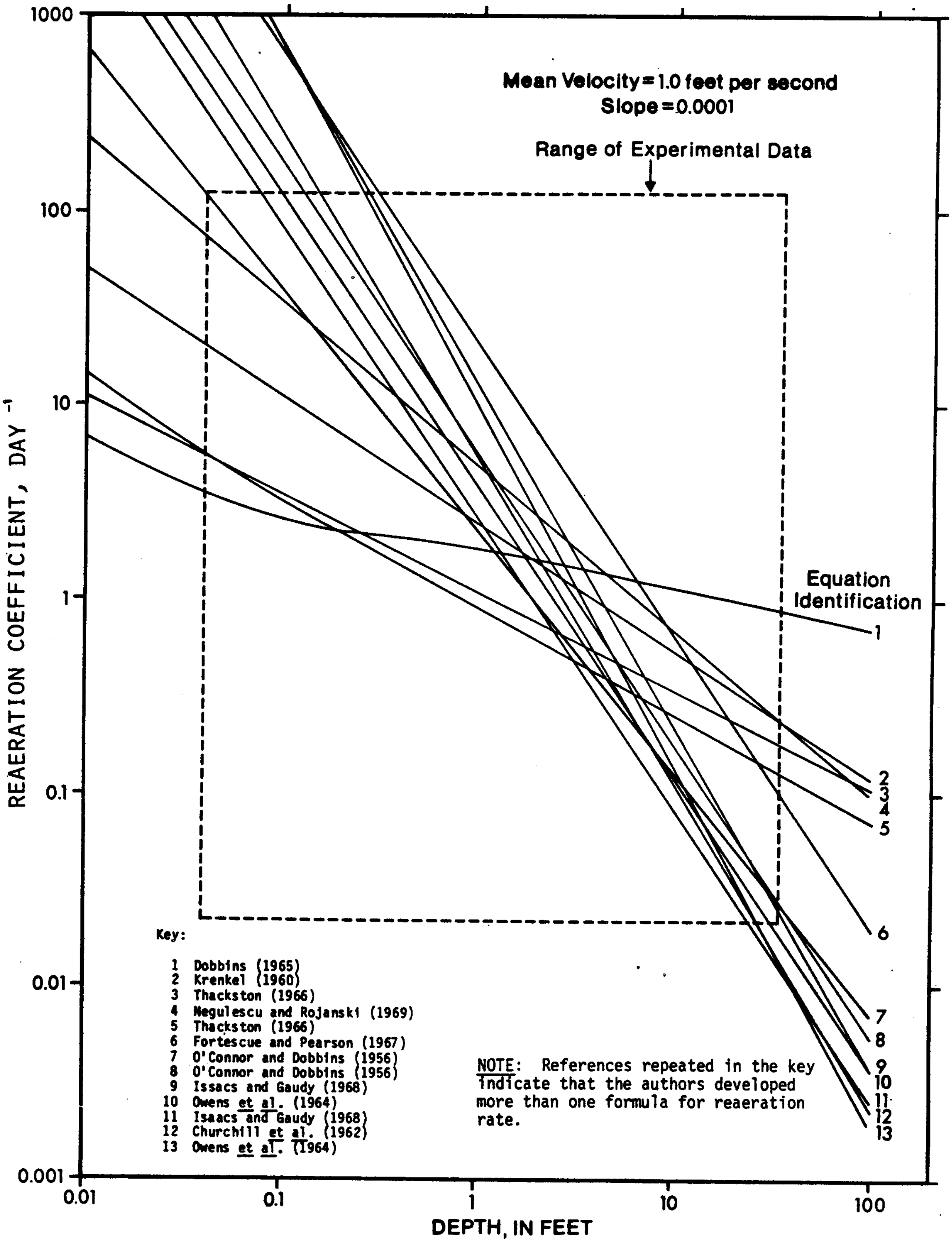
- Eighteen reaeration coefficient equations were compared against data collected in 28 Ohio streams.

- The streams were divided into four groups based on slope and velocity. The best predictive equations for each group are shown below:

Group	Slope (ft/mile)	Flow (cfs)	Preferred Equation
1	<3	All data	Negelescu-Royanski (1969) Krenkel-Orlob (1962)
2	3-10	≤30	Parkhurst-Pomeroy (1972)
3	3-10	>30	Thackston-Krenkel (1969)
4	>10	All data	Parkhurst-Pomeroy (1972) Tsivoglou-Neal (1976)

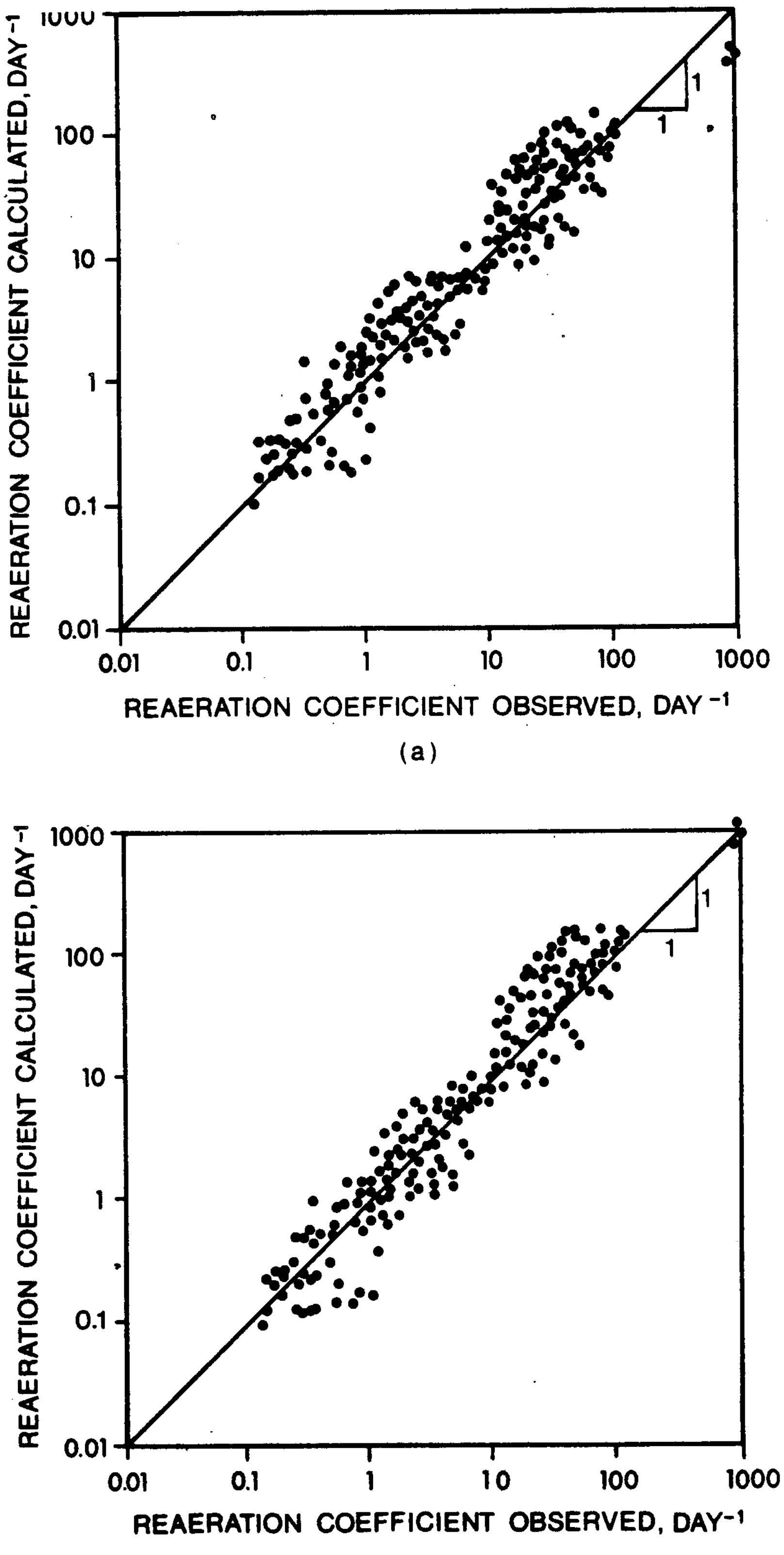
as well (see Figure 3-3). This was the formula which Bennett and Rathbun (1972) found produced the smallest error of the formulas they reviewed.

Figure 3-4 shows the three reaeration formulas found by Rathbun (1977) to best predict observed values for the Chattahoochee, Jackson, Flint,



NOTE: References repeated in the key indicate that the authors developed more than one formula for reaeration rate.

Figure 3-1. Predicted reaeration coefficients as a function of depth from thirteen predictive equations (from Bennett and Rathbun, 1972).



(a)

(b)

Figure 3-2. Comparisons of predicted and observed reaeration coefficients for the formula of Dobbins (1965) (a) and Parkhurst-Pomeroy (1972) (b).

South, and Patuxent Rivers. The range of reaeration coefficients analyzed here is considerably smaller than analyzed by Wilson and Macleod. The Tsivoglou - Wallace method is noticeably better than either the Padden-Gloyna or Parkhurst-Pomeroy methods. However, the Tsivoglou-Wallace method was originally developed using this data set, so it is not surprising that the fit is best.

Figure 3-5 shows the energy dissipation model of Shindala and Truax (1980) applied to streams with flow rates less than 280 cfs. They found that the best fit to the data was achieved when the flow rate was divided into two groups: less than 10 cfs and greater than 10 cfs.

Covar (1976), as discussed by Zison _et al_. (1978) found that the research of O'Connor-Dobbins (1958), Churchill _et al_. (1962), and Owens _et al_. (1964) could be used jointly to predict stream reaeration coefficients

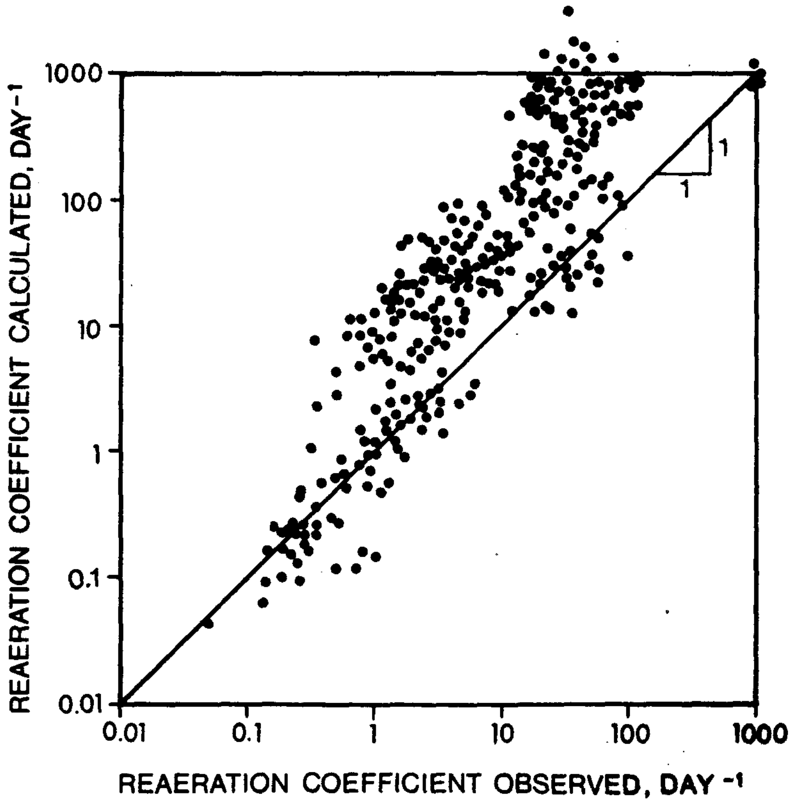
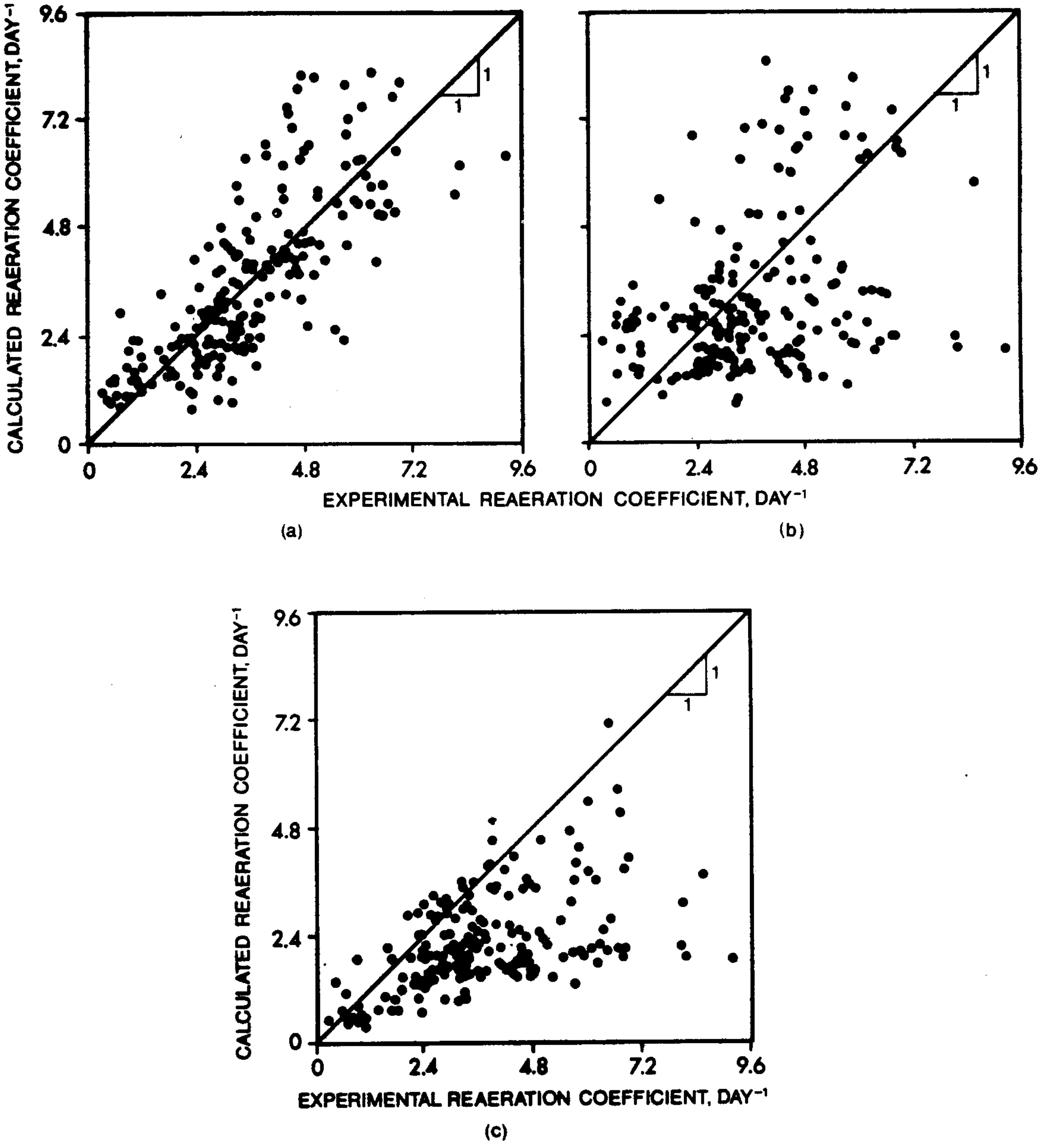


Figure 3-3. Formula of Bennett and Rathbun (1972) compared against observed reaeration coefficients.

for a range of depth and velocity combinations. Figure 3-6 shows the data points collected by each investigator and the regions Covar choose to divide the applicable formulas. Figure 3-7 shows the plots of reaeration prediction. Note that the predictions approximately match at the boundaries of each region.

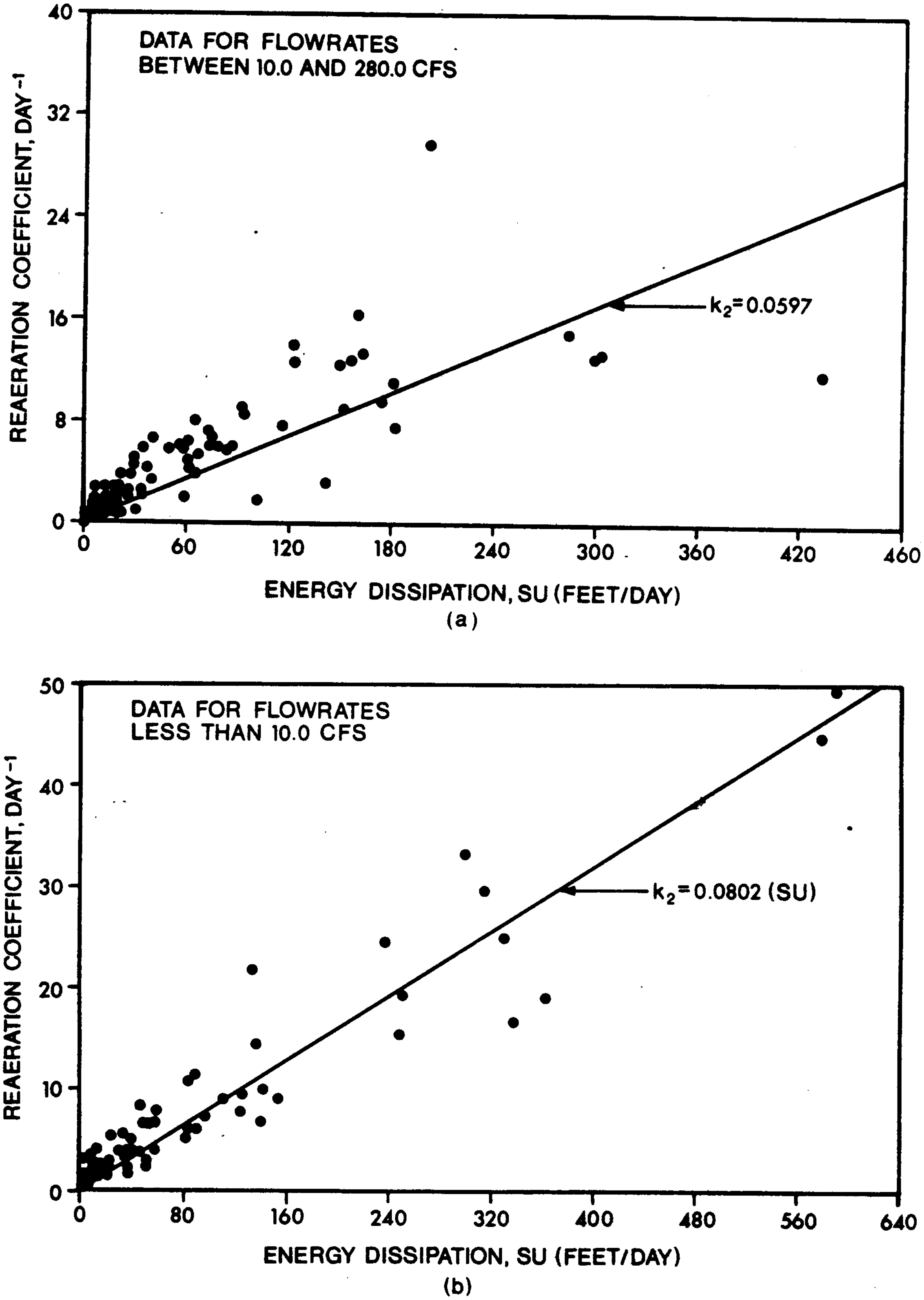


(a)

(b)

(c)

Figure 3-4. Calculated versus experimental reaeration coefficients for equations of (a) Tsivoglou and Wallace (1972), (b) Padden and Gloyna (1971), and (c) Parkhurst and Pomeroy (1972).



(a)

(b)

Figure 3-5. Reaeration coefficient versus energy dissipation (a) for flow rates between 10 and 280 cfs and (b) for flow rates less than 10 cfs. (Note: Curves for predicted reaeation coefficients are forced through the origin).

3.2.2.3 Measurement Techniques

Methods to determine reaeration rates based on instream data include the dissolved oxygen balance, deoxygenation by sodium sulfite, productivity measurements, and tracer techniques (both radioactive tracers and hydocarbon tracers). Today, use of tracers is the most widely accepted method. Productivity measurements are sometimes used, but because of their indirect approach could be subject to considerable error. Some of these methods are discussed in Kelly et al. (1975), Hornberger and Kelly (1975), and Waldon (1983). Only the tracer methods are discussed here.

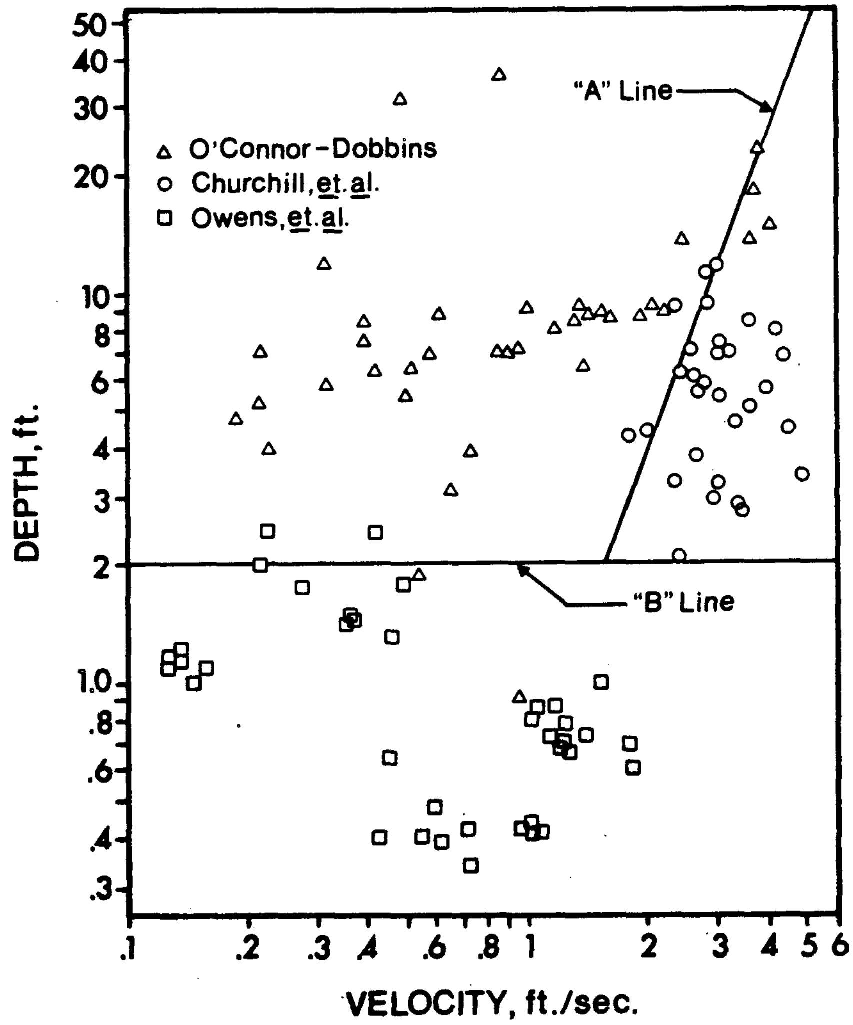


Figure 3-6. Field data considered by three different investigations.

The tracer method which appears to produce the most accurate results is the radioactive tracer technique developed and reported by Tsivoglou et al. (1965), Tsivoglou (1967), Tsivoglou et al. (1968), Tsivoglou and Wallace (1972), and Tsivoglou and Neal (1976). The method involves the instantaneous and simultaneous release of three tracers: krypton-85, tritium, and a fluorescent dye. The fluorescent dye indicates when to sample the invisible radioactive tracers and provides travel time information as well. The tritium acts as a surrogate for dispersion: the

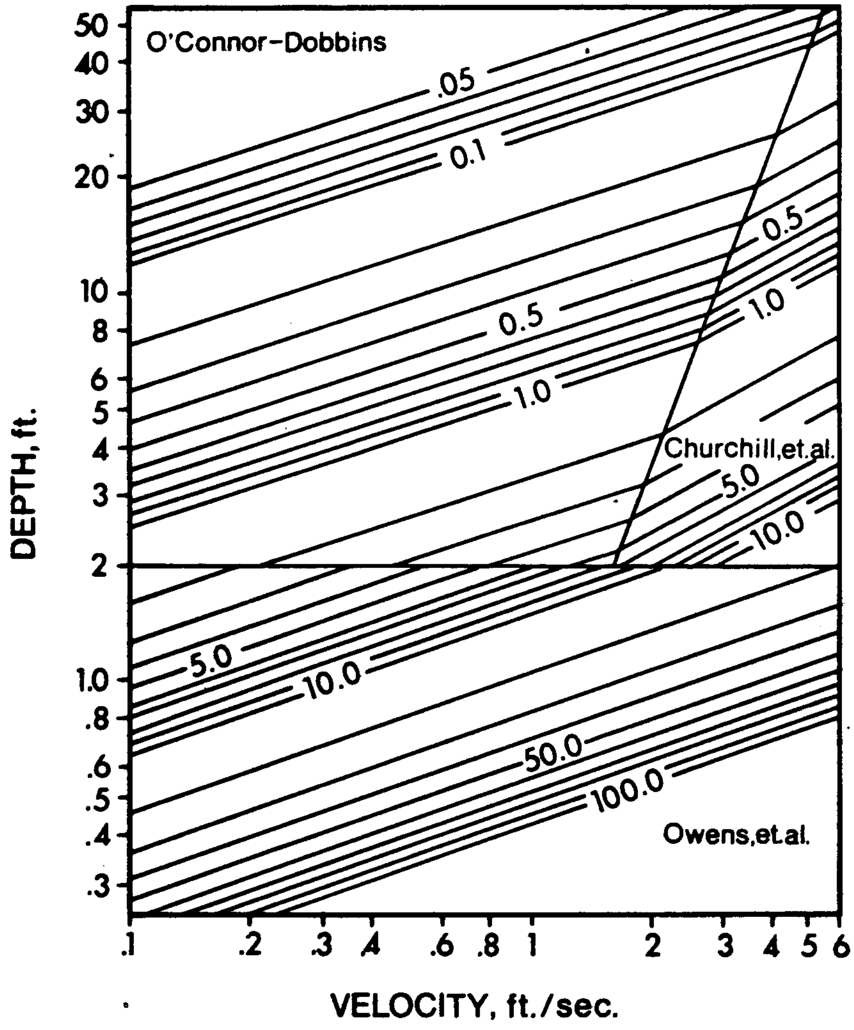


Figure 3-7. Reaeration coefficient (1/day) vs. depth and velocity using the suggested method of Covar (1976).

tritiated water disperses in the same manner as the natural water. The krypton-85 is lost to the atmosphere in a constant, known ratio compared with dissolved oxygen. The formula used is:

$$\frac{\left(\frac{C_{kr}}{C_{tr}}\right)_B}{\left(\frac{C_{kr}}{C_{tr}}\right)_A} = \exp\,(-k_{kr}t) \qquad (3\text{-}18)$$

where $\left(\frac{C_{kr}}{C_{tr}}\right)_{A,B}$ = concentration ratios of krypton and tritium at locations A and B when the dye peaks at each location

t = travel time between A and B

k_{kr} = atmospheric exchange rate of krypton

Since $\frac{k_{kr}}{k_2} = 0.83 \pm 0.04$, the dissolved oxygen reaeration rate, k_2, can be found directly from k_{kr}. The ratio 0.83 was found in the laboratory and has not been proven to be constant for all conditions.

Wilhelms (1980) has applied the radioactive tracer technique to flow through a hydraulic model. The results compared favorably with results from disturbed-equilibrium tests.

Because of the costs and potential hazards of using this method, other tracer techniques have been developed which do not use radioactive tracers. These methods have been discussed by Rathbun _et al_. (1975), Rathbun _et al_. (1978), Rathbun and Grant (1978), Kwasnik and Feng (1979), Bauer _et al_. (1979), Rathbun (1979), Jobson and Rathbun (undated), Grant and Skavroneck (1980), House and Skavroneck (1981), Rainwater and Holley (1984), Wilcock (1984a), and Wilcock (1984b). Not all researchers agree on the accuracy to the modified tracer techniques. Kwasnik and Feng (1979), Grant and Skavroneck (1980), House and Skavroneck (1981) all reported successful applications of the method. However, NCASI (1982b) reported that the hydrocarbon tracer technique produced results higher than both the radioactive tracer and empirical methods. The application was on a large sluggish stream. Rainwater and Holley (1984) have investigated two assumptions of the hydrocarbon tracer technique (constant ratios between mass transfer coefficients and negligible absorptive losses) and found both assumptions to be valid for that particular study.

The modified tracer techniques use a hydrocarbon gas tracer and a fluorescent dye (e.g., rhodamine-WT) as the dispersion-dilution tracer. Sometimes two different tracer gases (e.g., ethylene and propane) can be used simultaneously to yield two estimates of reaeration rate. Two methods can be used: the peak concentration method and the total weight method.

Using the total-weight method the exchange rate of the tracer with the atmosphere, k_T is computed as follows:

$$k_T = \frac{1}{t_d - t_u} \log_e \left(\frac{A_u Q_u}{A_d Q_d} \right) \tag{3-19}$$

where A_u and A_d = areas under the gas concentration-versus-time curves at the upstream and downstream ends of the reach, respectively, and

Q_u and Q_d = stream discharge at each end of the reach.

The reaeration coefficient k_2 is computed as:

$$k_2 = \begin{cases} \dfrac{k_T}{.87} \text{, ethylene} & \text{(3-20a)} \\[2em] \dfrac{k_T}{.72} \text{, propane} & \text{(3-20b)} \end{cases}$$

Recently Wilcox (1984a, b) has proposed methyl chloride as a gas tracer. At 20°C,

$$k_2 = \frac{k_T}{.707} \text{, for methyl chloride} \tag{3-20c}$$

The methyl chloride transfer coefficient k_T was found to exhibit a temperature dependence.

The peak concentration method is similar in form to the radioactive tracer equation:

$$k_T = \frac{1}{t_d - t_u} \log_e \frac{\left(\frac{C_T}{C_D}\right)_u}{\left(\frac{C_T}{C_D}\right)_d} \frac{(A_D)_d}{(A_D)_u} \tag{3-21}$$

where k_T = the base e desorption coefficient for the tracer gas;

t_d-t_u = the time of travel between the peak concentrations;

C_T and C_D = the peak concentrations of the tracer gas and rhodamine-WT dye, respectively

$(A_D)_d$, $(A_D)_u$ = area under dye versus time curve downstream and uptream, respectively

More recently Yotsukura et al. (1983) have conducted tests to assess the feasibility of a steady-state propane gas tracer method as a means of estimating reaeration coefficients. The tests were conducted on Cowaselon Creek, New York. It was concluded that the steady state method, which also includes an instantaneous injection of dye tracer, is feasible and provides a reliable method of determining the reaeration coefficient.

3.2.2.4 Special Influences on Reaeration

In addition to hydraulic variables which typically appear in the expressions in Table 3-6, the reaeration coefficient can be influenced by certain special factors which include:

- surfactants
- suspended particles

- wind
- hydraulic structures, and
- water temperature

While surfactants, suspended solids, and wind can influence reaeration in rivers, in practice the effects of these factors are rarely if ever included in water quality models. Discussion of the influence of surfactants is given in Zison _et al_. (1978), Poon and Campbell (1967), and Tsivoglou and Wallace (1972). The influence of suspended solids is discussed by Holley (1975) and Alonso _et al_. (1975).

3.2.2.4.1 Wind Effects

While wind effects are typically not included in reaeration predictions in rivers, there is evidence that at high wind speeds, the reaeration rates can be significantly increased. These effects are occasionally alluded to in the literature when experimental measurements are abnormally high.

Eloubaidy and Plate (1972) performed experiments in the wind-wave facility at Colorado State University. They arrived at the following expression for the surface transfer coefficient, k_L, in feet per day:

$$k_L = \frac{C U_{*_s} h U_{*_c}}{\upsilon} \tag{3-22}$$

where C = a constant of proportionality

υ = kinematic viscosity of water, m^2/sec

U_{*_s} = surface shear velocity due to wind, m/sec = $0.0185\ V_w^{1.5}$

V_w = wind speed, m/sec

U_* = water shear velocity defined as $\sqrt{ghS_c}$, m/sec

h = normal depth (i.e., depth with uniform flow), m

S_c = pressure-adjusted channel slope, unitless, $S_o + \frac{1dp}{\rho gdx}$

ρ = mass density of water, kg/m^3

g = gravitational constant, m/sec^2

S_o = slope of energy gradient (channel slope for uniform flow), unitless

$\frac{dP}{dx}$ = air pressure gradient in the longitudinal direction, kg/m^2-sec^2

From their experiments Eloubaidy and Plate found that C = .0027.

The variables comprising Equation 3-22 are readily obtainable, with the exception of the pressure gradient. The authors determined that an error on the order of 2 percent was obtained in k_2 (= k_L/h) by neglecting the pressure gradient.

A summary of the conditions under which Equation 3-22 was developed is:

channel slope:	.00043, .001
air velocity:	22, 30, 38 fps for each slope
discharge:	0.79, 0.83, 0.91 cfs at 0.001 slope
	0.58, 0.63, 0.75 cfs at 0.0043 slope
water depth:	0.385 feet

Note the extremely high wind velocities used in the experiments (greater than 22 fps). Hence the validity of the approach to lesser wind speeds typically encountered in the natural environment has not been demonstrated.

Mattingly (1977) also performed laboratory studies of the effects of wind on channel reaeration. He obtained this empirical expression:

$$\frac{k_2}{(k_2)_o} - 1 = 0.2395\ V_w^{1.643} \qquad (3\text{-}23)$$

where k_2 = reaeration coefficient under windy conditions, 1/day
$(k_2)_o$ = reaeration coefficient without wind, 1/day
V_w = wind velocity in meters per second in the free stream above the boundary layer near the water surface

A plot of the experimental data is shown in Figure 3-8. Note the importance of wind induced reaeration at moderate to high wind speeds. Further discussion of the effects of wind are found in Gulliver and Stefan (1981) and Frexes _et_ _al_. (1984).

Because wind effects are typically neglected in river and stream reaeration modeling, this approach is equivalent to assuming a zero wind velocity. For many water quality modeling applications, such as wasteload allocation, this approach is reasonable.

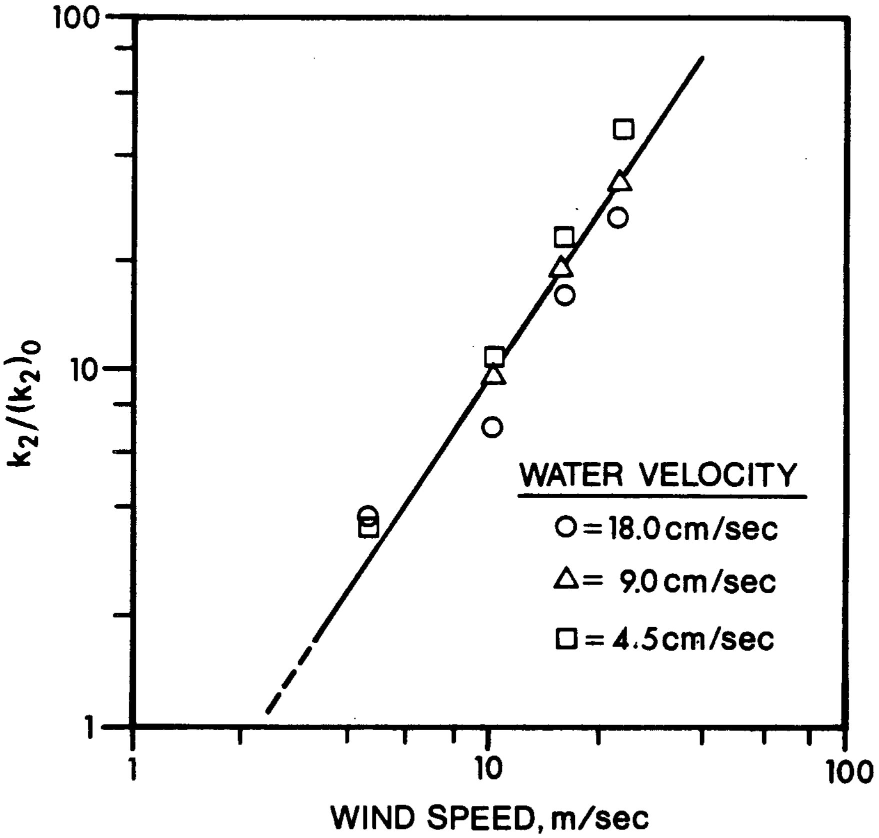


Figure 3-8. Ratio of reaeration coefficient under windy conditions to reaeration coefficient without wind, as a function of wind speed (based on laboratory studies).

3.2.2.4.2 Small Dams

On many rivers and streams small to moderate sized dams are present. Dams can influence reaeration by changing the dissolved oxygen deficit from 1 to 3 mg/l (typically) in a very short reach of the river. Table 3-8 summarizes various predictive equations that have been used to simulate the effects of small dams. Avery and Novak (1978) discuss limitations of these equations and aspects of oxygen transfer at hydraulic structures.

Butts and Evans (1983) have reviewed various approaches that predict the effects of small dams on channel reaeration and further collected field data on 54 small dams located in Illinois to determine their reaeration characteristics. They identified 9 classes of structures, and quantified the aeration coefficient b for use in the following formula:

$$r = \frac{C_s - C_u}{C_s - C_d} = 1 + 0.38abh\ (1 - 0.11h)\ (1 + 0.046T) \qquad (3\text{-}24)$$

where a = water quality factor (0.65 for grossly polluted streams; 1.8 for clean streams)
b = weir dam aeration coefficient
h = static head loss in meters
T = water temperature, ^{o}C

Figure 3-9 shows the general structural classification and the aeration coefficient, b, for each class.

The present review does not include influences of large dams, artificial reaeration, or other hydraulic structures. Cain and Wood (1981) discuss aeration over Aviemore Dam, 40 m (130 ft) in height, Banks et al. (1983) and NCASI (1969) discuss effects of artificial reaeration, and Wilhelms et al. (1981), Wilhelms (1980), and Wilhelms and Smith (1981) further discuss reaeration related to hydraulic structures.

TABLE 3-8. EQUATIONS THAT PREDICT THE EFFECTS OF SMALL DAMS ON STREAM REAERATION

Reference	Predictive Equation	Units	Source
Gameson (1957)	$r = 1+0.5abh$	h, in meters	field survey
Gameson *et al*. (1958)	$r = 1+0.11ab(1+0.046T)h$	h, in feet	model
Jarvis (1970)	$r_{15} = 1.05\ h^{0.434}$	h, in meters	model
Holler (1971)	$r_{20} = 1+0.91h$	h, in meters	model
Holler (1971)	$r_{20} = 1+0.21h$	h, in meters	prototype
Department of the Environment (1973)	$r = 1+0.69h(1-0.11h)(1+0.0464T)$	h, in meters	model
Department of the Environment (1973)	$r = 1+0.38abh(1-0.11h)(1+0.046T)$	h, in meters	model
Nakasome (1975)	$\log_e(r_{20}) = 0.0675h^{1.28}\ q^{0.62}\ d^{0.439}$	d, h, in meters q, in m^2/hr	model
Foree (1976)	$r = \exp(0.1bh)$	h, in feet	field survey

Symbols: $r_T = \frac{C_s-C_u}{C_s-C_d}$

C_s = dissolved oxygen saturation

C_u, C_d = concentration of dissolved oxygen upstream and downstream of dam, respectively

a = measure of water quality (0.65 for grossly polluted; 1.8 for clear)

b = function of weir type

h = water level difference

d = tailwater depth below weir

q = specific discharge.

T = water temperature, °C

3.2.2.4.3 Temperature Effects on Reaeration

The influence of temperature on reaeration is typically simulated using the following type of temperature dependence:

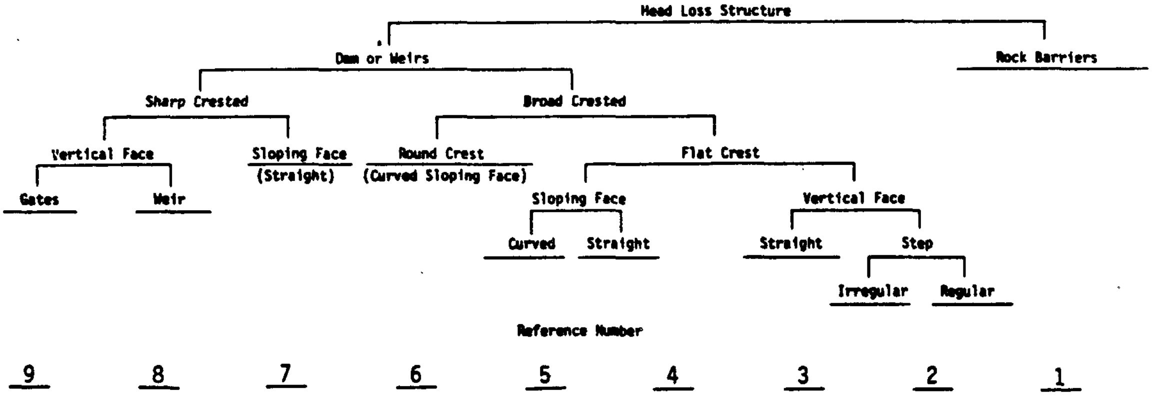
$$k_2(T)=k_2(20^{\circ}C)\theta^{T-20} \tag{3-25}$$

where T = water temperature, °C

θ = temperature adjustment factor

Table 3-9 summarizes values of θ from the literature. Typically values of 1.022 to 1.024 are used in most modeling applications.

Schneiter and Grenney (1983) developed a different approach to simulate temperature corrections over the ranges 4°C to 30°C. Their approach effectively allows θ to vary as a function of temperature. However, the approach is not widely used.



Reference Numbers	Dam Type (2)	b
1	Flat broad-crested regular step	0.70
2	Flat broad-crested irregular step	0.80
3	Flat broad-crested vertical face	0.80
4	Flat broad-crested straight slope face	0.90
5	Flat broad-crested curved face	0.75
6	Round broad-crested curved face	0.60
7	Sharp-crested straight slope face	1.05
8	Sharp-crested vertical face	0.80
9	Sluice gates with submerged discharge	0.05

Figure 3-9. Division of head loss structures by dam type.

TABLE 3-9. REPORTED VALUES OF TEMPERATURE COEFFICIENT

Temperature Coefficient, θ	Reference
1.047	Streeter, et al. (1926)
1.0241	Elmore and West (1961)
1.0226	Elmore and West (1961)
1.020	Downing and Truesdale (1955)
1.024	Downing and Truesdale (1955)
1.016	Dowining and Truesdale (1955)
1.016	Streeter (1926)
1.018	Truesdale and Van Dyke (1958)
1.015	Truesdale and Van Dyke (1958)
1.008	Truesdale and Van Dyke (1958)
1.024	Churchill et al (1962)
1.022	Tsivoglou (1967)
1.024	Committee on Sanitary Engineering Research (1960)

3.2.2.5 Sources of Data

Many sources of stream reaeration rates exist in the literature. Table 3-10 summarizes a number of the major sources. Many state agencies are also repositories of reaeration data.

3.2.3 Reaeration in Lakes

Simulation of reaeration in lakes is normally accomplished using the surface transfer coefficient k_L rather than the depth averaged k_2. Most often in lake simulations the surface transfer coefficient k_L is assumed to be a function of wind speed.

TABLE 3-10. SOURCES OF STREAM REAERATION DATA

Source	Contents
Owens et al., (1964)	Reaeration coefficients using disturbed equilibrium technique for six rivers in England (Ivel, Lark, Derwent, Black Beck, Saint Sunday's Beck, Yewdale Beck), and associated hydraulic data.
O'Connor and Dobbins (1958)	Reaeration data for Clarion River, Brandywine Creek, Illinois River, Ohio River, and Tennessee River.
Churchill et al. (1962)	Reaeration data using dissolved oxygen balance downstream from deep impoundments for Clinch River, Holston River, French Broad River, Watauga River, Hiwassee River.
Tsivoglou and Wallace (1972)	Hydraulic properties and radioactive tracer measured reaeration coefficients for Flint, South, Patuxent, Jackson, and Chattahoochee Rivers.
Bennett and Rathbun (1972)	Summaries of data from Churchill et al., (1962), Owens et al., (1964), Gameson et al. (1958), O'Connor and Dobbins (1958), Tsivoglou et al., (1967,1968), Negulescu and Rojanski (1969), Thackston (1966), Krenkel (1960).
Foree (1976)	Radioactive tracer measurements and reaeration hydraulic characteristics for small streams in Kentucky, and reaeration measurements for small dams in Kentucky.
Grant (1976)	Reaeration measurements and hydraulic characteristics for 10 small streams in Wisconsin.
Grant (1978)	Reaeration measurements and hydraulic characteristics for Rock River, Wisconsin.
Zison et al. (1978)	Summary of reaeration coefficients and hydraulic characteristics for rivers throughout the United States.
Kwasnik and Feng (1979)	Reaeration data using the modified tracer technique on selected streams in Massachusetts.
Grant and Skavroneck (1980)	Reaeration data from three small streams in Wisconsin.
House and Skavroneck (1981)	Reaeration data for two small streams in Wisconsin.
Shindala and Truax (1980)	Radioactive tracer measurements of reaeration rates and escape coefficients, plus hydraulic data, for rivers in Mississippi, Wisconsin, Texas, Georgia, North Carolina, Kentucky and New York.
Terry et al. (1984)	Hydrocarbon tracer measurements of k_2 and hydraulic data for Spring Creek, Osage Creek, and Illinois River, Arkansas. Bennett-Rathbun (1972) best fit all three streams. Eight equations were tested.
Bauer et al. (1979)	Hydrocarbon tracer measurements of k_2 and hydraulic data for the Yampa River, Colorado best matched the Tsivoglou Neal and Thackston and Krenkel energy dissipation type equations. Lau's equation was extremely error prone. Nineteen equations were tested.
Goddard (1980)	Hydrocarbon tracer measurements of k_2 and hydraulic data from the Arkansas River in Colorado were used to test 19 equations. The best fitting equations were those by Dobbins, Padden and Gloyna, Langbein and Durum, and Parkhurst and Pomeroy.

TABLE 3-10. (continued)

Source	Contents
Hren (1983)	Radioactive tracer measurements for the North Fork Licking River, Ohio.
Rathbun et al. (1975)	Hydrocarbon tracer measurements for West Hobolochitts Creek, Mississippi.
NCASI (1982c)	Radioactive tracer measurements for Ouachita River, Arkansas, and Dugdemona River, Louisiana.
Parkhurst and Pomeroy (1972)	Reaeration coefficients were determined by a deoxygenation method in 12 sewers in the Los Angeles County Sanitation District.
Ice and Brown (1978)	Reaeration coefficients were determined using sodium sulfite to deoxygenate the water in small streams in Oregon.
Ohio Environmental Protection Agency (1983)	Reaeration coefficients were determined for 28 different streams in Ohio using predominantly the modified tracer technique, and in one case the radioactive tracer technique.
Long (1984)	Reaeration coefficients, hydraulic data, and time of travel data collected on 18 streams in Texas.

Since many lakes are not vertically well-mixed, multiple layers are often used to simulate dissolved oxygen dynamics. Atmospheric reaeration occurs only through the surface layer, and then dissolved oxygen is dispersed and advected to layers lower in the water body.

Table 3-11 summarizes various methods that have been used to simulate reaeration in lakes. With the exception of the method of Di Toro and Connolly (1980), all formulas include a wind speed term. Di Toro and Connolly applied a constant surface transfer coefficient to Lake Erie. They found that the surface layer of the lake remained near saturation so that the value of k_L used was not important as long as it was sufficiently high to maintain saturated dissolved oxygen levels in the surface layer.

All the surface transfer coefficients shown in Table 3-11 should be viewed as empirical; the researchers have simply hypothesized that the suggested formulas are adequate to simulate reaeration. The coefficients (a and b) are of limited validity, and should be treated as calibration parameters. O'Connor (1983) has analyzed from a more theoretical point of view the effects of wind on the surface transfer coefficient.

TABLE 3-11 . REAERATION COEFFICIENTS FOR LAKES

Author(s)	Surface Transfer Rate, k_L (m/day)
Di Toro and Connolly (1980)	$k_L = 2.0$
Chen _et_ _al_., (1976)	$k_L = \frac{86400D}{(200-60\sqrt{V})10^{-6}}$ D = molecular diffusion coefficient of oxygen in water, m^2/sec V = wind speed, m/sec
Banks (1975)	$k_L = 0.362\ V^{1/2}$ for $0 \leq V \leq 5.5$ m/sec $k_L = 0.0277V^2$ for $V > 5.5$ m/sec
Baca and Arnett (1976)	$k_L = a + bV$ a = 0.005 - 0.01 m/day b = 10^{-6} - 10^{-5} m^{-1} V = wind speed, m/day
Smith (1978)	$k_L = a + bV^2$ a = 0.64 m/day b = 0.128 $sec^2 m^{-1} day^{-1}$ V = wind speed, m/s
Liss (1973)	$k_L = 0.156\ V^{0.63}$ $V \leq 4.1$ m/sec $k_L = 0.0269V^{1.9}$ $V > 4.1$ m/sec V = wind speed, m/sec
Downing and Truesdale (1955)	$k_L = 0.0276V^{2.0}$ V = wind speed, m/sec
Kanwisher (1963)	$k_L = 0.0432V^2$ V = wind speed, m/sec
Broecker _et_ _al_. (1978)	$k_L = 0.864V$ V = wind speed, m/sec
Yu _et_ _al_. (1977)	$k_L = 0.319V$ V = wind speed, m/sec
Broecker and Peng (1974)	$k_L = 0.0449V^2$ V = wind speed, m/sec

TABLE 3-11. (Cont'd)

Author(s)	Surface Transfer Rate, k_L (m/day)
Weiler (1975)	$k_L = 0.398$ $V < 1.6$ m/sec $k_L = 0.155V^2$ $V \geq 1.6$ m/sec V = wind speed, m/sec

Notes:

1. Elevation of wind speed measurements is not always reported.
2. a and b are empirically determined.

Some limited research has addressed the influence of rainfall on reaeration (Banks et al., 1984; Banks and Herrera, 1977). Rainfall effects are more of theoretical interest rather of practical concern.

3.2.4 Reaeration in Estuaries

The present state of reaeration simulation in estuaries combines concepts used in river and lake approaches. Very little original research on estuarine reaeration has been completed to date.

Table 3-12 summarizes different formulations that have been used to predict reaeration in estuaries. The different approaches include both k_L (surface transfer) and k_2 (depth averaged) reaeration terms. In some models, k_2 can be specified (e.g., Genet et al., 1974 and MacDonald and Weismann, 1977). O'Connor et al. (1981) specified the surface transfer rate to be 1 m/day in their two-layered model of the New York Bight. One of the more widely used approaches is the O'Connor (1960) formula, which has subsequently been modified to include wind speed terms (Thomann and Fitzpatrick, 1982).

Few field studies have been performed for the purpose of directly measuring reaeration in estuaries. Baumgartner et al., (1970) used Krypton-85 to measure the range of reaeration in the Yaquina River Estuary. However, no predictive formulas were developed.

TABLE 3-12. REAERATION COEFFICIENTS FOR ESTUARIES

Reference	Reaeration Rate
O'Connor (1960)	$k_2 = \frac{(D_L U_o)^{1/2}}{H^{3/2}}$ (1/day) U_o = mean tidal velocity over a complete cycle, m/day D_L = molecular diffusivity of oxygen, m^2/day H = average depth, m
Genet *et al.*, (1974)	k_2 = user specified
O'Connor *et al.*, (1981)	k_L = 1 m/day
MacDonald and Weisman (1977)	k_2 = user specified
Harleman *et al.*, (1977)	$k_2^a = 10.86 \frac{V^{0.6} H W_T}{H^{1.4} A}$ (1/day) V = tidal velocity, ft/sec H = depth, ft W_T = top width, ft A = cross-sectional area, ft
Thomann and Fitzpatrick (1982)	$k_2 = \frac{13V^{0.5}}{H^{1.5}} + \frac{3.281}{H} (0.728W^{0.5} - 0.317W + 0.0372W^2)$ (1/day) V = depth averaged velocity, fps H = depth, ft W = wind speed, m/sec
Ozturk (1979)	$k_2 = \frac{4.56V^{4/3}}{H}$ (1/day) V = mean tidal velocity, m/sec H = mean depth, m

[a] The coefficient 10.86 is the recommended value, but can be changed as discussed by Harleman *et al.* (1977).

Tsivoglou (1980) has discussed the application of radioactive tracer techniques to small estuaries within the Chesapeake Bay. Special discussion was given to the Ware River Estuary.

3.2.5 Summary

The most common method of simulating reaeration in rivers is to use the depth averaged k_2 approach, while in lakes the surface transfer rate k_L is typically used. In estuaries either k_2 or k_L is used, depending on the importance of stratification. Very little research on reaeration has been done in either lakes or estuaries. In lakes, reaeration is typically specified to be a constant or to be a function of wind speed. Little information is available on how to select parameters in the wind speed functions. Site specific calibration of the parameters may be required.

In contrast to lakes and estuaries much research has been conducted on reaeration in rivers. Thirty-one formulas were shown earlier in Table 3-6. The formulas have been developed based on hydraulic parameters, most often depth and velocity. Consequently, the variables in reaeration expressions are generally not of concern in distinguishing among the utility of the formulas. One exception is formulas that contain longitudinal dispersion coefficients, which are difficult to quantify.

Considerable evidence shows that reaeration formulas are most applicable over the range of variables for which they were developed, and outside of that range, errors might be quite large. This suggests that reaeration rates developed from laboratory flume data may be quite limited for natural stream applications. Some research supports this supposition (Bennett and Rathbun, 1972).

Previous reviews of stream reaeration (see Table 3-7) have shown that no one formula is best under all conditions, and depending on the data set used, the range of the reaeration coefficients in the data set, and the error measurement selected, the "best" formula may change. Some of the reaeration rate expressions which have been judged "best" during past reviews are:

- The O'Connor and Dobbins (1958), Dobbins (1964), and Thackston and Krenkel (1969) formulas best fit the entire range of data reviewed by Bennett and Rathbun (1972).

- The Churchill et al. (1962) formula provided the best fit to natural stream data in the Bennett and Rathbun review.
- The methods of Dobbins (1964) and Parkhurst and Pomeroy (1972) gave the best fits to the data reviewed by Wilson and MacLeod (1974).
- The Tsivoglou-Wallace and Parkhurst-Pomeroy methods were best in the review by Rathbun (1977).
- The energy dissipation model produced the best correlation for small streams based on the study of Shindala and Truax (1980).

From previous reviews, one of the more popular and more accurate methods for reaeration rates prediction is the energy dissipation method of Tsivoglou. The method requires knowledge of the escape coefficient, which appears to depend on streamflow. Typical values of the escape coefficient are 0.08/ft for flow rates less than 10 cfs, and 0.06/ft for flow rates between 10 and 280 cfs.

The method of Covar (1976), which combines the O'Connor-Dobbins, Churchill et al., and Owens et al., formulas, has merit in that it attempts to limit the use of the three formulas to within the depth-velocity range for which they were developed. However, for relatively small and shallow streams, the method of Owens et al., tends to overestimate reaeration, so that the energy dissipation method, which appears to perform well in small streams, could be used to supplement the method.

The radioactive tracer method appears to be the best method for measuring stream reaeration coefficients. Even so, the coefficients that are predicted are valid only for the particular flow condition existing at the time of sampling. Thus to completely characterize the range of values of the reaeration coefficient would require numerous sampling events or use of an acceptable predictive equation.