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**Publication Date**

1981



# Lawrence Berkeley Laboratory

UNIVERSITY OF CALIFORNIA, BERKELEY

## EARTH SCIENCES DIVISION

Published in ESTUARIES, Volume 3, No. 4,  
December 1980, pp. 242-247

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January 1981

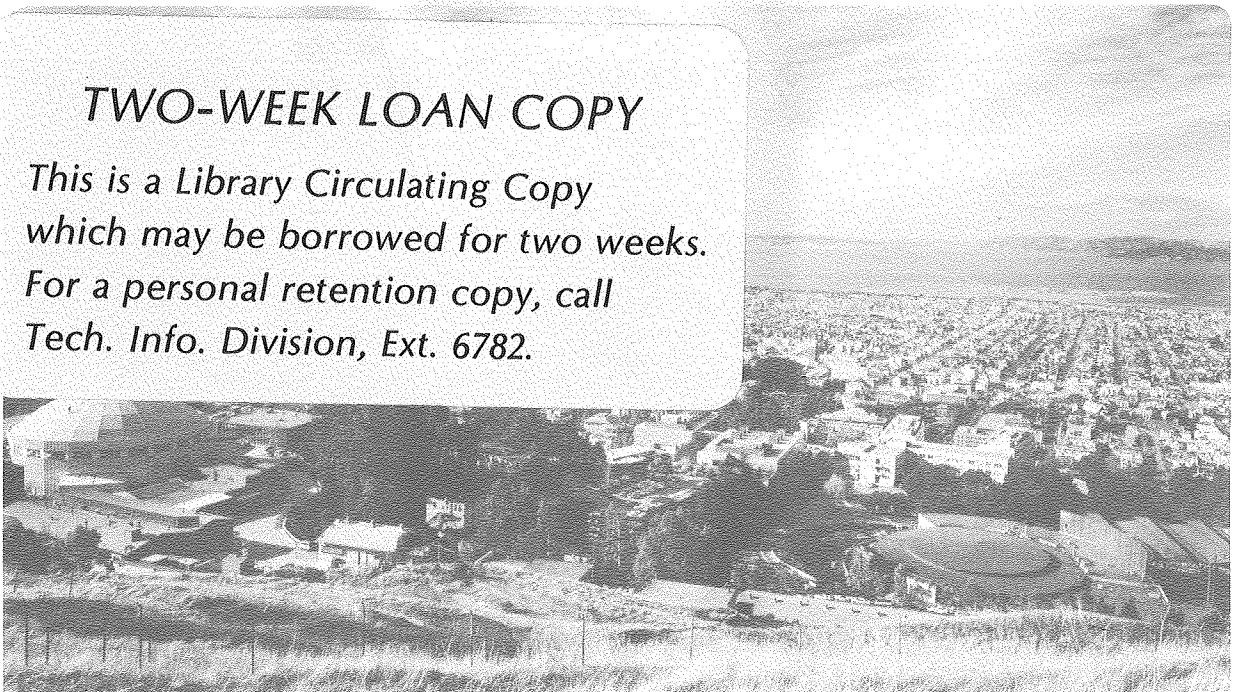
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## Seasonal Oxygen Depletion in Chesapeake Bay

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**ABSTRACT:** The spring freshet increases density stratification in Chesapeake Bay and minimizes oxygen transfer from the surface to the deep layer so that waters below 10 m depth experience oxygen depletion which may lead to anoxia during June to September. Respiration in the water of the deep layer is the major factor contributing to oxygen depletion. Benthic respiration seems secondary. Organic matter from the previous year which has settled into the deep layer during winter provides most of the oxygen demand but some new production in the surface layer may sink and thus supplement the organic matter accumulated in the deep layer.

### Introduction

Oxygen depletion zones which approach anoxia deny use of that portion of the ecosystem to most plankton, nekton and benthos. Thus, oxygen depletion in coastal waters and estuaries impacts upon benthic and demersal populations, including commercially important species, and may adversely influence use of these waters by man for power plant cooling, waste disposal and recreation.

Summer deep-water oxygen depletion is common in fjords, in many lakes (Richards 1965), in some estuaries, including Chesapeake Bay (Newcombe and Horne 1938; Carpenter and Cargo 1957; Whaley, et al. 1966; Hobbie, et al. 1972; Taft and Taylor 1976a, 1976b), among the San Juan Islands (Johnson and Thompson 1929) and in parts of the Puget Sound system (Barnes and Collias 1958; Christensen and Packard 1976). However, large scale coastal ocean deoxygenation, such as observed on the Middle

Atlantic Shelf during the summer of 1976 (IDOE 1976), is not common.

Associated with summer oxygen depletion in Chesapeake Bay are accumulations of soluble reactive phosphorus (SRP) and ammonium. These nutrients are transported vertically to the euphotic zone where they support the annual open-bay primary productivity maximum and phytoplankton biomass maximum. Thus, summer deep-water oxygen depletion is an important facet of the annual nutrient cycle.

There are three processes which could influence deep water oxygen concentration: increased water column stratification during the spring freshet, which minimizes oxygen replenishment from surface to deep waters; increased plankton and benthic respiration rates due to warming in spring; increased organic loading to the deep water due to increased primary productivity in overlying waters during spring. The purpose of this study was to establish the relative impor-

tance of these three processes to seasonal oxygen depletion in Chesapeake Bay.

This study was supported by Department of Energy contract number EY-76-S-02-3279, document number C00-3279-34 and by National Science Foundation grants GA-33445, OCE 76-11399 and OCE 76-01853. Chesapeake Bay Institute contribution number 274. We gratefully acknowledge the assistance of the crews of the R/V MAURY and R/V WARFIELD. We thank Dr. J. Heinbokel for critically reading the manuscript.

### Methods

Data from two surveys of Chesapeake Bay (Whaley, et al. 1966; Taylor and Cronin 1974) were augmented with additional data collected during 1977. R/V MAURY cruises in 1964-66 and R/V WARFIELD cruises in 1969-71 and 1977 were made to established Chesapeake Bay Institute stations (Fig. 1). On all cruises a submersible pump was lowered from surface to near bottom to collect samples for profiling temperature and conductivity (Schiemer and Pritchard 1961) and dissolved oxygen (Carpenter 1965). Respiration measurements were made by Winkler titration (Hartwig and Michael 1978) on water collected with a Niskin sterile bag sampler. Precision estimates for replicate standardizations on board ship were  $\pm 3.2\%$  and  $\pm 0.5\%$  on two cruises. Particulate carbon (PC) was measured by the method of Gordon and Sutcliffe (1973), chlorophyll *a* by Strickland and Parsons (1972) and adenosine triphosphate (ATP) by luciferase assay after Strickland and Parsons (1972). Primary productivity was measured by  $^{14}\text{C}$  uptake during simulated *in situ* incubations (McCarthy, et al. 1974).

### Results

**Volume of Low  $\text{O}_2$  Water.** Data in Table 1 for August 1970 indicate that the south-north extent of low oxygen deep-water, defined here as  $<20\%$  saturated, was about  $37^\circ 44' \text{N}$  to  $39^\circ 04' \text{N}$  (stations 744 to 904N). The upper boundary of water less than  $20\%$  saturated occurred at about 9-10 m depth between  $38^\circ 04' \text{N}$  to  $39^\circ 04' \text{N}$ . Lateral transects along  $38^\circ 04' \text{N}$ ,  $38^\circ 45' \text{N}$  and  $39^\circ 04' \text{N}$  in July 1977 (Fig. 2) also show the upper boundary for low oxygen water at about 10

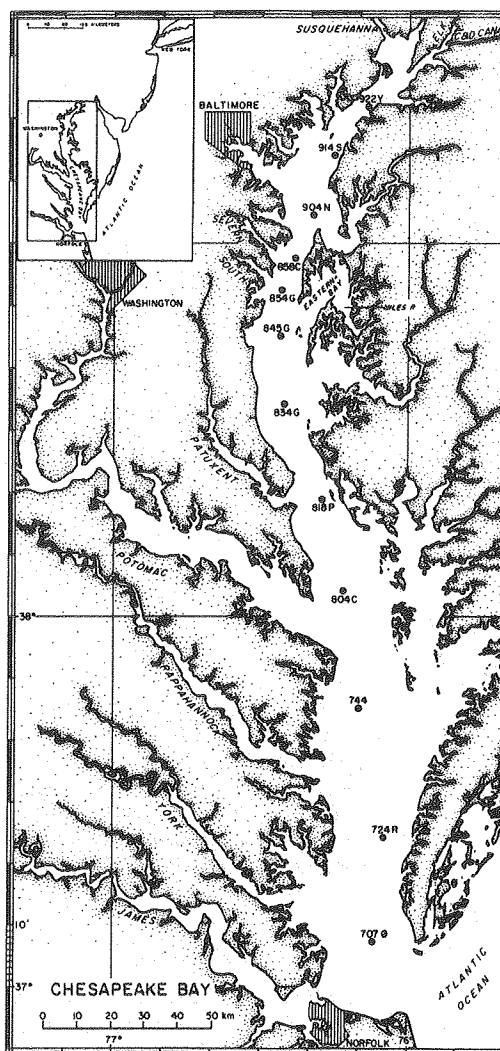


Fig. 1. Station locations.

m or less. Therefore, the minimum volume of low oxygen water in mid-summer is estimated at  $6 \times 10^9 \text{ m}^3$  (Cronin and Pritchard 1975) which is about 8% of the entire Bay volume and 26% of the volume north of  $38^\circ 04' \text{N}$ .

**Rate of Oxygen Depletion.** During the three study periods oxygen concentrations were highest in December-February and generally began to decrease in March or April. Fig. 3 shows data for 1977. Minimum values were observed by June in each year except 1965 when the minimum was observed in early July. The rates of change in deep-water oxygen concentrations from the

TABLE 1. Bottom water oxygen saturation in August 1970. U = Undetectable. - = 20% saturation not observed.

Station	% O <sub>2</sub> Saturation in Bottom Water	Approx. Depth of 20% Saturation (m)
922Y	60.8	—
914S	33.1	—
904N	1.1	9
858C	U	9
845G	U	10
834G	U	9
818P	4.5	10
804C	12.8	10
744	15.5	14
724R	58.2	—
7070	72.9	—

winter-spring maximum to the summer minimum ranged from  $-0.03 \text{ ml}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$  in 1971 to  $-0.11$  in 1969 (Table 2). Six of the seven values were between  $-0.06$  and  $-0.11 \text{ ml}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$ . Thus, when averaged over the same seasons for seven years, oxygen consumption in the deep water exceeded the replenishment rate by  $0.08 \text{ ml}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$ .

Deep water oxygen depletion rates for Chesapeake Bay are considerably more rapid than for the Puget Sound basins examined by Barnes and Collias (1958) in which average oxygen utilization from winter to summer ranged  $0.005\text{--}0.020 \text{ ml}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$ . Oxygen utilization in Dabob Bay, Washington, was  $1.1 \mu\text{g}\cdot\text{atom}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$  (Barnes and Collias 1958) and  $1.25 \mu\text{g}\cdot\text{atom}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$  (Christen-

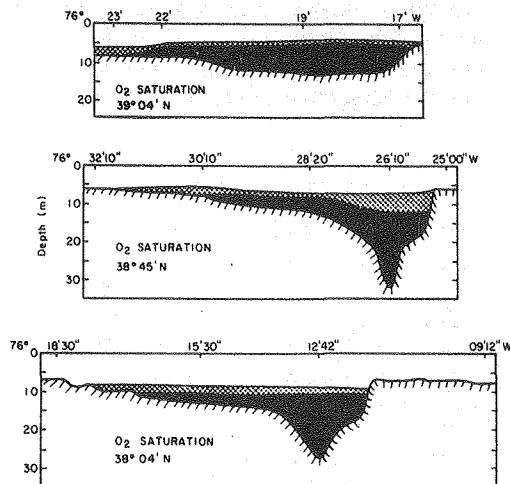


Fig. 2. Summer oxygen saturation on three cross bay transects. White = greater than 20% saturation, cross hatch = 0–20% saturation, black = H<sub>2</sub>S present.

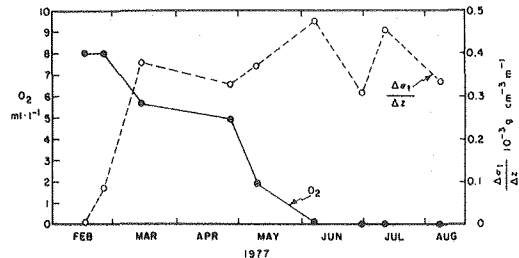


Fig. 3. Oxygen concentration and change in sigma-t with depth for station 845G.

sen and Packard 1976) in two studies as compared to an average of  $7.3 \mu\text{g}\cdot\text{atom}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$  for Chesapeake Bay. Differences in the availability of readily oxidizable organic matter and in the rates of vertical exchange in Dabob and Chesapeake bays were probably the major factors contributing to these observations.

**Spring Freshet.** Water column density structure in partially mixed estuaries is largely controlled by salinity. Maximum salinity at all depths occurred in January–February in the years studied. During the spring the upper layer freshened from 12–15‰ to 5–10‰ at latitude  $38^{\circ}45' \text{N}$  for example, due to runoff from the Susquehanna River. Data for the seven years considered here show decreases in oxygen concentration immediately following the dilution of surface layer salinity by spring river runoff. The spring freshet was characterized by sustained surface salinity decrease averaging  $0.07\text{‰}\cdot\text{d}^{-1}$  at station 845G during spring of 1965 and 1966, and  $0.06\text{‰}\cdot\text{d}^{-1}$  in spring of 1970 and 1971. In some of the data, salt was diluted in two steps due to downstream displacement of water at all depths by the fresh water pulse entering the head of the Bay

TABLE 2. Deep water oxygen change at Station 845G measured from winter-spring concentration maximum to summer minimum.

	$\Delta \text{O}_2$ $\text{ml}\cdot\text{l}^{-1}\cdot\text{d}^{-1}$
1964	-0.10
1965	-0.08
1966	-0.06
1969	-0.11
1970	-0.06
1971	-0.03
1977	-0.09

TABLE 3. Respiration of whole plankton samples measured as oxygen loss expressed in  $\text{mg-at O}_2 \cdot \text{d}^{-1} \cdot \text{column heading}$ . Incubation temperatures were 5 °C in February, 15 °C in April and 25 °C in August.

Station	February 1976		April 1976			August 1976		
	mg-at C <sup>-1</sup>	mg Chl a <sup>-1</sup>	mg-at C <sup>-1</sup>	mg Chl a <sup>-1</sup>	mg ATP <sup>-1</sup>	mg-at C <sup>-1</sup>	mg Chl a <sup>-1</sup>	mg ATP <sup>-1</sup>
904N	0.45	4.4	0.54	12	54	0.52	6.4	38
854G						0.75	8.9	33
834G	0.77	9.1	0.43	16		0.49	8.4	61
818P			0.36	7	24	0.49	9.8	28
804C	0.71	15.7				1.0	32	64

(Schubel 1972). Following initial displacement, seaward flowing fresh water diluted the surface salinity thereby increasing density differences between upper and lower layers of the estuary. This was evident in April 1964, May 1965, March 1966, May 1969 and 1977.

**Respiration.** Winter respiration rates (February 1976) ranged  $0.23\text{--}0.89 \text{ ml} \cdot \text{l}^{-1} \cdot \text{d}^{-1}$ , spring rates (April 1976 and 1977) were  $0.26\text{--}1.35 \text{ ml} \cdot \text{l}^{-1} \cdot \text{d}^{-1}$  and summer rates (August 1975 and 1976) were  $0.54\text{--}1.91 \text{ ml} \cdot \text{l}^{-1} \cdot \text{d}^{-1}$ . These broad ranges reflect differences in temperature, plankton biomass and composition, substrate availability and ambient oxygen concentration. The effects of containment on the natural populations could also contribute to the wide ranges (Venrick, et al. 1977). But, since the worst analytical precision expected is  $\pm 3.2\%$  (Hartwig and Michael 1978) the variability in the respiration rate measurements was not due primarily to analytical technique.

Respiration rates normalized to biomass expressed as particulate carbon (PC), chlorophyll *a* and ATP are shown in Table 3. When normalized to PC, respiration rates were comparable in February, April, and August 1976 at incubation temperatures of 5, 15 and 25 °C respectively. Respiration normalized chlorophyll *a* ranged 4.4 to 15.7  $\text{mg-at O}_2 \cdot \text{mg Chl a}^{-1} \cdot \text{d}^{-1}$  in February, 7.0 to 12  $\text{mg-at O}_2 \cdot \text{mg Chl a}^{-1} \cdot \text{d}^{-1}$  in April and 6.4 to 32  $\text{mg-at O}_2 \cdot \text{mg Chl a}^{-1} \cdot \text{d}^{-1}$  in August. Respiration normalized to ATP was 24 and 54  $\text{mg-at O}_2 \cdot \text{mg ATP}^{-1} \cdot \text{d}^{-1}$  in April and 28 to 64  $\text{mg-at O}_2 \cdot \text{mg ATP}^{-1} \cdot \text{d}^{-1}$  in August. No clear evidence for elevation in biomass specific oxygen consumption with temperature increase from February to August was observed.

Comparing water column respiration rates with average deep water oxygen change of  $-0.08 \text{ ml} \cdot \text{l}^{-1} \cdot \text{d}^{-1}$  reveals that the

seasonal net loss of oxygen from the deep water is about 10% of the maximum winter respiration rate and 4% of the maximum summer rate. Therefore, in Chesapeake Bay, the imbalance between oxygen flux to the deep water and water column respiration need be sustained at only 10% or less of the respiration rate in order to produce the observed seasonal depletion.

**Primary Productivity.** Monthly primary productivity data for April 1969 to April 1971 for station 845G and 834G are shown in Fig. 4. Productivity during February to May, the period of most rapid oxygen depletion, ranged 15 to 427  $\text{mg C} \cdot \text{m}^{-2} \cdot \text{h}^{-1}$  at 845G and 53 to 142  $\text{mg C} \cdot \text{m}^{-2} \cdot \text{h}^{-1}$  at 834G. The April 1971 value of 427  $\text{mg C} \cdot \text{m}^{-2} \cdot \text{h}^{-1}$  for 845G was unusually high in comparison to the more commonly observed rates of about 50 to 200  $\text{mg C} \cdot \text{m}^{-2} \cdot \text{h}^{-1}$  in February to May. Productivity increased about three-fold from February to May. However, the annual maximum occurred in summer, after oxygen depletion was well underway, or in fall when mixing and reoxygenation were occurring. Therefore, oxygen depletion is not specifically a response to organic loading resulting from phytoplankton productiv-

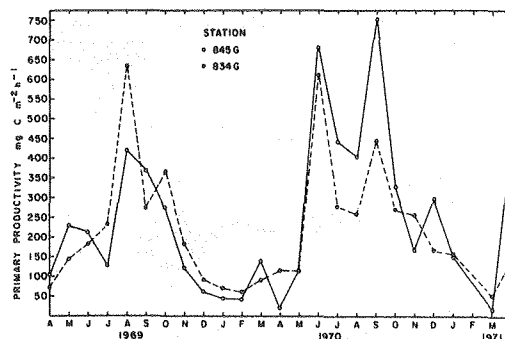


Fig. 4. Primary productivity at two stations in upper Chesapeake Bay.

ity maxima but may be primarily supported by organic material accumulated in the deep water during the previous fall and winter. Hargrave (1972) found that benthic oxygen demand in Lake Esrom was related to epilimnetic productivity and Smith (1974) suggested that productive surface waters increase metabolism of underlying deep-sea sediments. Deep anoxia layer formation in Tokyo Bay has been related to the sinking of fecal material resulting from copepod grazing on phytoplankton in the euphotic zone (Seki, et al. 1974). Also, oxygen consumption may be closely coupled to primary productivity on the Atlantic Continental Shelf (Rowe, et al. 1975).

#### Discussion

Chesapeake Bay is a partially mixed estuary with thermohaline stratification through much of the year (Pritchard 1968). The water column density structure regulates advective exchange from the upper to lower layers. The primary factor controlling density structure is salinity which is graded vertically and longitudinally in the Bay. Rates of salinity change with depth and distance are influenced primarily by fresh water inflow. With low river flow during late summer and winter, sea water penetrates along the bottom of the estuary past Baltimore. River flow usually increases sharply between February and May, freshening the surface waters and increasing the vertical salinity gradient. Increased water column stability reduces advective transport of oxygen to deep water. As oxygen renewal declines, upsetting the balance between oxygen consumption and its input rate, biological and chemical utilization continues to reduce dissolved oxygen to very low concentrations. The water temperature increase in spring compounds the imbalance by reducing oxygen solubility, and further stabilizing the water column.

The data summarized here support this scenario but, as yet, a quantitative cause and effect relation between river flow and oxygen depletion rate cannot be established because the relation between river flow and salinity is itself complex. Boicourt (1969) modelled in one dimension the relation between upper Chesapeake Bay salinity distribution and Susquehanna River flow. Even

this relatively simple approach required the use of a separate predictor model to describe changes in the seaward boundary condition with time. Chesapeake Bay's response to a freshet is a function of the Bay's recent history and cannot be linearized or easily predicted. Two dimensional numerical models describing salinity response to river flow have been proposed (Elliott 1976; Wang and Kravits 1980) and may lead to acceptable quantitation of the salinity-river flow relation for Chesapeake Bay.

The wide ranges of biomass specific plankton respiration observed in winter, spring, and summer overlapped and the oxygen concentration declined at low temperatures during February to March 1966, suggesting that temperature alone has little effect on respiration. There needs to be only a 10% imbalance between respiration and replenishment for the daily oxygen loss from deep water to account for the observed oxygen depletion.

The existing data do not permit us to ascertain whether oxygen utilization in the water column or the sediments dominates the loss process, but an estimate can be made. Hargrave (1969) reported a regression of fresh water and marine sediment respiration data from the literature which estimated oxygen consumption at  $360 \text{ ml O}_2 \cdot \text{m}^{-2} \cdot \text{d}^{-1}$  at  $10^\circ \text{C}$  which is within the range of observed values for Puget Sound sediments (Pamatmat 1971) and Atlantic Shelf sediments (Rowe, et al. 1975). Distributing this respiration through  $1 \text{ m}^3$  of overlying water gives  $0.36 \text{ ml O}_2 \cdot \text{l}^{-1} \cdot \text{d}^{-1}$  which is comparable to the minimum April water column consumption we observed. However, the minimum April water column consumption we measured ( $0.26 \text{ ml O}_2 \cdot \text{l}^{-1} \cdot \text{d}^{-1}$ ) integrated over a 10 m deep lower layer is  $2,600 \text{ ml O}_2 \cdot \text{m}^{-2} \cdot \text{d}^{-1}$  or about 7 times the calculated sediment utilization. Thus it appears water column respiration dominates seasonal oxygen depletion in Chesapeake Bay deep water, at least until the oxygen concentration falls below the minimum required for healthy aerobic plankters. Christensen and Packard (1976) similarly found that plankton metabolism accounted for 70-85% of the oxygen depletion in Dabob Bay. Increased primary production in spring contributes oxidizable or-



ganic matter to the system, but is not the major factor directly driving oxygen depletion in Chesapeake Bay. Organic material from the previous summer and fall which collects in the deep water during winter is more likely providing the oxygen demand.

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This work was partially supported by the U.S. Department of Energy under Contract W-7405-ENG048.