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VOLUME XV

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OCEANOGRAPHY OF LONG ISLAND SOUND, 1952–1954

II. PHYSICAL OCEANOGRAPHY

By

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ABSTRACT

Temperature and salinity data were obtained at eight stations in the central part of the Sound at weekly or biweekly intervals from March 1952 to March 1954. Seasonal trends and differences from one year to the next are discussed in relation to Weather Bureau data on air temperature and precipitation. Tidal mixing is sufficient to prevent the development of strong stability, although a small thermocline is present from February or March until the end of August; there is a vertical salinity gradient during most of the year. Direct current measurements indicate a weak nontidal drift, but possibly the currents are not continuously present, for observations on the horizontal distribution of salinity and density show that conditions favoring the maintenance of density currents are readily modified by transient winds. The distribution of temperature and salinity in the Sound as a whole is briefly described. Recent and previous current measurements are combined in a generalized estimate of east-west mass transport, and problems of transport exchange and salt balance are discussed. The seasonal temperature cycle in the central basin is used to calculate vertical eddy conductivity coefficients for all of the two-year survey except the periods from mid-August to mid-November. The eddy coefficients are indeterminate during the early autumn, and it is suggested that convection is more important than turbulence in controlling this part of the seasonal cycle. Data on radiation and water transparency are presented, and the latter are analyzed with respect to phytoplankton, winds, currents, and other factors that affect transparency in shallow coastal waters.

TEMPERATURE AND SALINITY DISTRIBUTION

Methods

Temperature measurements at each station included a bathythermograph lowering from surface to bottom as well as a surface temperature reading with a G-M bucket thermometer. BTs were read to the nearest tenth of a degree Fahrenheit at depth intervals of 2.5 m and converted to the nearest 0.05° C. Comparison of average surface readings with the corresponding thermometer temperatures was used to apply a systematic correction to BT readings when the average difference was as much as 0.05°.

Water samples for salinity determinations and other chemical analyses were collected with a Nansen bottle one meter below the surface, one to two meters above the bottom, and at one or two intermediate depths, the number depending on the depth. Salinities were titrated according to the simplified method described by Harvey (1928), using Woods Hole standard sea water to standardize the silver nitrate. The slightly superior accuracy of the Knudsen method is hardly warranted in neritic waters where the salinity range requires the use of two burettes and where local variability often exceeds the titration error.

Seasonal Cycles in the Central Part of the Sound

Fig. 1 shows the temperature and salinity of the inshore waters from March 5, 1952 to March 10, 1954. Surface values are the averages of Sts. 1, 6, 7, and 8 (cf. Introduction, Fig. 1). The bottom temperatures represent depths of about 8 to 12 m and are averages of Sts. 1, 7, and 8. St. 6, in water of only about 4 m, is omitted.

The average temperature and salinity at offshore Sts. 2 to 5 are plotted in similar fashion in Fig. 2. The bottom depths range from

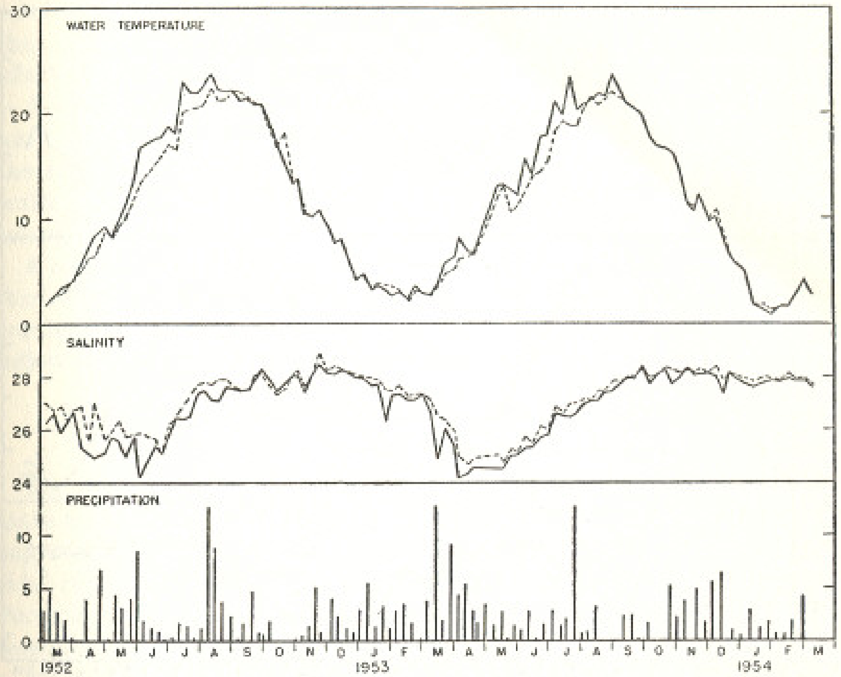

Figure 1. Average temperature (°C) and salinity (°/₀₀) at the surface (solid lines) and bottom (dashed lines) at inshore stations in the central part of Long Island Sound. Total precipitation (cm) between successive dates of oceanographic observation.

19 to 28 m. Included in the figures are data on precipitation (Fig. 1) and air temperature (Fig. 2) obtained through the courtesy of the New Haven office of the U. S. Weather Bureau. Weather data are recorded at 41° 16′ N, 72° 53′ W, half a mile from Long Island Sound and about three miles from Sts. 6 and 7. They are plotted here as average temperature and total precipitation between successive dates of oceanographic observation. Comparison of weather data for 1952–54 with the long-term means of the Weather Bureau shows that in 1952 the summer air temperatures were about 1° C above average, the autumn temperature approximately normal. But the winter of 1952–53 was one of the warmest on record, with individual monthly means ranging from 1.6 to 3.5° C above the long-term average. The spring and autumn of 1953 were also above average, the summer normal. Warm weather persisted through December, but January 1954 was 1.4° below normal.

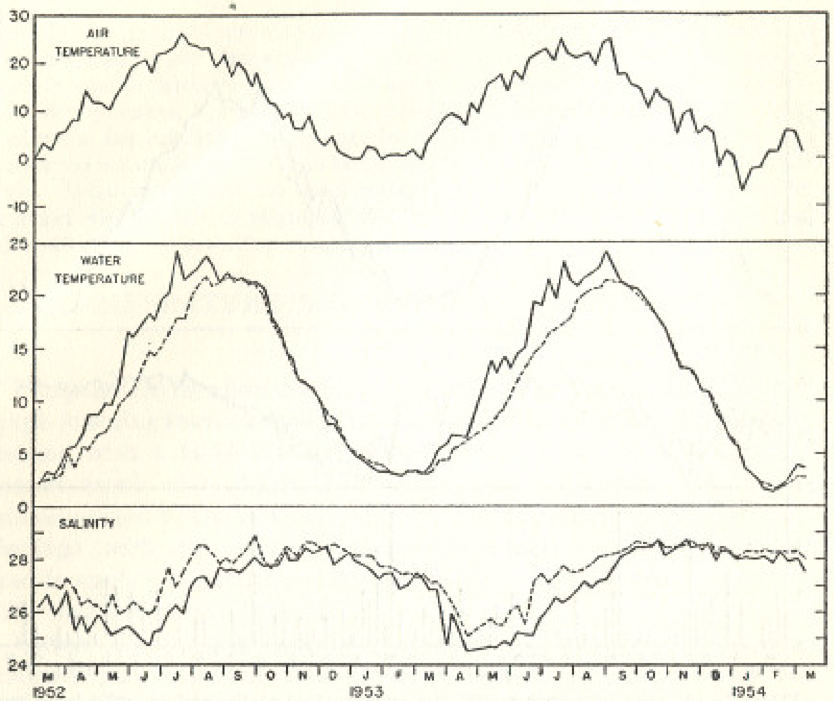

Figure 2. Weekly averages of air temperature (°C) recorded by the New Haven Weather Bureau. Average water temperature (°C) and salinity (°/₀₀) at surface (solid lines) and bottom (dotted lines) at offshore Sts. 2 to 5.

These differences in air temperature are reflected in sea surface temperatures. The latter were slightly warmer in the summer of 1952 than in 1953. Autumn cooling in 1953 lagged two or three weeks behind the preceding year. However, rapid cooling in January 1954 produced a midwinter minimum that was normal or perhaps below average for the area and was nearly 3° colder than the year before.

The total precipitation was 119 cm in 1952 and 137 cm in 1953, as compared with an 81-year mean of 117.5 cm. March and April 1953 were excessively wet, and the salinity dropped off rapidly at that time. Otherwise there were no marked differences in salinity between one year and the next. It has been suggested previously (Riley, 1952) that the volume of freshwater drainage regulates the rate of transport exchange, so that the salinity tends to be held at a

relatively constant level. This theory is strengthened by the fact that a 15% increase in rainfall in 1953 had only a small and transient effect on the general salinity level.

A small thermocline was found each year from February or March until the end of August. Tidal mixing was strong enough so that the surface layer was seldom more than 5° warmer than the bottom water. Positive temperature gradients of as much as 1° have been found in winter.

A small vertical salinity gradient was present most of the time. It tended to be maintained, irrespective of thermal stability, by the combined effects of freshwater dilution of the surface layer and the inflow of saline water along the bottom.

Factors Affecting Salinity and Density Distribution in the Central Basin

Fig. 3 is a chart of the central part of the Sound, reviewing routine station positions and showing measurements of nontidal surface currents. The latter include estimates based on measurements of tidal currents by the U. S. Coast and Geodetic Survey and additional current stations obtained during the present survey, using the method described by Pritchard and Burt (1951).

Many small streams empty into the Sound along the northern shore of the central and western basins, and the Housatonic is a river of moderate size, accounting for approximately one-tenth of the total drainage into the Sound. Previous surveys have commonly shown reduced salinity in the northern inshore waters. This, combined with the evidence from current measurements, led to the hypothesis (Riley, 1952) that a coastwise density current begins in the vicinity of New Haven, flows westward and gives rise to a counterclockwise gyral in the western half of the Sound.

Between New Haven and the mouth of the Housatonic River, but not including the latter, the total freshwater drainage varies seasonally between about 0.3 and 1.5 million m³/day. The major part of this water comes from rivers emptying into New Haven Harbor. Assuming an average coastwise drift of about two miles a day (this is the mean of the available current measurements and is in agreement with the observed movement of oil accidentally dumped into New Haven Harbor), the drainage could reduce the salinity about 0.12 to 0.60 ‰ in a strip of coastal water two miles

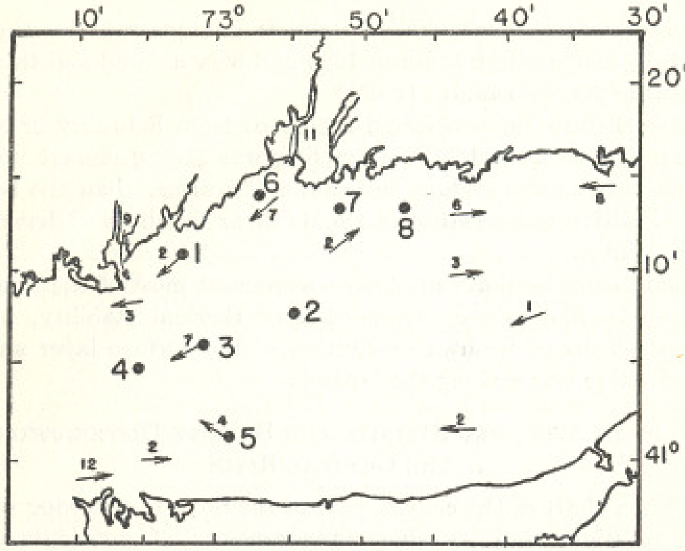

Figure 3. Chart of central Long Island Sound. Dots and large numbers indicate routine stations. Arrows and small numbers show observed direction and speed of nontidal drift in centimeters per second. Values at Sts. 1, 3, and 5 were obtained during the present survey. Others are estimates from Riley (1952), based on U. S. Coast and Geodetic Survey tidal current charts.

wide. The Housatonic River, on the other hand, has a drainage volume roughly an order of magnitude larger than the figures cited above, and wherever its outflow goes, it may be expected to have a correspondingly greater effect on salinity.

During the recent survey, about 5% of the observations at St. 1 have revealed a markedly reduced surface salinity that was almost certainly due to Housatonic River influence. The mouth of the river is three miles west of St. 1 and only slightly beyond the limits of the normal ebb tide excursion. Hence an occasional effect of this kind does not seriously contradict the theory of a general westerly drift.

However, for other reasons the theory needs to be re-examined and somewhat modified. Whatever the actual water movement may be, it has become apparent that conditions tending to generate a density current are not continuously present. Water of relatively low density has been found only about half of the time at St. 1. With northerly and westerly winds it has been common to find water of lower density in the surface layer offshore; concomitantly, the water at St. 1 has temperature and salinity characteristics which are typical of mid-depth

or bottom water from the offshore region. It has also been common at such times to find eddies of fresh water containing debris of terrestrial origin in the vicinity of St. 4. The most likely source of such water is the Housatonic River five miles to the north. In order to present the data as a whole in simple form, differences in surface water density (sigma-t) between Sts. 3 and 1 have been tabulated in the first part of Table I in relation to wind direction. The coastline in this area is oriented approximately SW to NE. Water movement is likely to be somewhat to the right of the wind. Offshore movement is therefore most likely to occur with north to southwest winds.

Table I. Frequency Distribution and Average Magnitude of Density and Salinity Gradients in Relation to Wind Direction During the Day Preceding the Observations

Wind direction		*N to SW*	*S to NE*
Density gradient,	+	11	11
St. 3 minus St. 1	−	22	2
	Mean	−.12	.20
Salinity gradient	+	11	7
St. 4 minus St. 3	−	21	6
	Mean	−.25	.02

Sts. 1 and 3 were ordinarily occupied about 9–10 A.M. Wind data are averaged by the Weather Bureau from midnight to midnight. Thus the most pertinent wind data are for a period of roughly 33 to 8 hours preceding the station observations. This is hardly ideal; nevertheless, the results in Table I clearly show a relation between density distribution and transient winds. Plus signs designate the number of occasions when a positive density gradient was found, a condition favoring the establishment of a coastwise density current. Minus signs indicate the frequency distribution of denser water inshore, which is presumed to be associated with offshore movement and upwelling. The next line of Table I is the mean of all observations, positive and negative, for each group.

The second half of Table I presents an analogous comparison of surface salinities at offshore Sts. 3 and 4. With southerly and easterly winds there is no significant difference and none to be expected. With northerly and westerly winds, about two-thirds of the observations indicate a freshening effect, which is postulated to be due primarily to southward transport of Housatonic River water.

Although it would appear in Table I that the wind effect is transient, it is reasonable to suppose that there might be a prevailing pattern of density distribution over a period longer than a day or a few days, corresponding to patterns of prevailing winds. Table II shows vectorial averages of wind speed in miles per hour and direction by semimonthly periods during the summers of 1952 and 1953. Below these figures are listed all of the observed differences in sigma-t between Sts. 3 and 1 for each period; also listed are differences between Sts. 2 and 1. The latter are regarded as somewhat inferior in quality because the stations are widely separated, but they greatly increase the total amount of data available for examination.

TABLE II. Comparison of Semimonthly Averages of Wind Speed and Direction with Observed Horizontal Density Gradients

	June		*July*		*August*	
1952	*1–15*	*16–30*	*1–15*	*16–31*	*1–15*	*16–31*
Wind direction	252	268	200	254	—	225
Wind speed	2.1	1.7	1.7	3.7	0.0	1.5
Density difference, 3–1	—	.02	−.37	+.17	−.18	−.12
2–1	−.10	−.60	−.34	−.67	−.86	−.38
	—	−.42	−1.26	−.38	.07	.20
	—	—	−.43	—	—	—
1953						
Wind direction	250	197	077	053	067	238
Wind speed	2.3	3.1	0.5	0.1	1.3	2.2
Density difference, 3–1	−.36	.07	.25	.01	.09	−.45
	−.38	—	—	—	—	—
2–1	−.11	−.29	−.29	.00	.08	−.25
	−.18	.38	.03	−.15	.00	−.40
	−.07	—	—	—	—	—

During the summer of 1952, the prevailing winds were westerly, and the preponderance of negative gradients indicated considerable offshore movement and upwelling. The winds were southerly and easterly for two months in the summer of 1953. A majority of the horizontal gradients at that time were positive, suggesting that the inshore water mass was sufficiently stable to become freshened by river drainage, even though the latter is relatively small in summer.

Differences of this type from one year to the next may be important biologically. Many species of bottom invertebrates find a suitable substrate only in the inshore waters, and successful reproduction requires that their planktonic larvae reach the stage of metamorphosis in the coastal zone. Winds that maintain a stable inshore water mass during the period of planktonic existence therefore favor a successful year class. The pattern of prevailing winds may also affect phytoplankton production. However, the latter is a much more complex problem, since it involves the effect of wind on both physical dispersal and the rate of supply of nutrients by upwelling.

The remainder of the two-year period has been examined in the same way. Qualitatively the relationship between winds and density distribution holds throughout, but the quantitative aspects of the relationship vary considerably from one season to another. There are periods in autumn and early winter when strong westerly winds have little effect on the density distribution. This simply means that the Sound is thoroughly mixed and that no amount of upwelling can create a pronounced horizontal gradient. Contrariwise, during the spring peak in freshwater drainage, positive density gradients tend to be maintained despite northerly winds.

With regard to the other inshore sampling positions, St. 6, just inside the New Haven Harbor breakwater, ordinarily had a slightly reduced salinity, as might be expected. St. 7 occasionally showed slight harbor influence on the ebb tide. None was noted at St. 8, which is in an area where previous studies suggested the possibility of a diffuse northerly drift, completing the postulated counterclockwise gyral in the western half of the Sound. No definitive evidence on this subject has since been obtained. However, salinities have been generally similar to these at offshore St. 2 and have exhibited no freshening effect that might be due either to easterly movement from New Haven Harbor or to a coastwise density current originating from the Connecticut River.

Horizontal Distribution in the Sound as a Whole

Previous observations on the general distribution of temperature and salinity in the Sound have been summarized by Riley (1952). Since that time, three additional cruises have been made. Surface observations are shown in Figs. 4 to 6. The east-west salinity gradients and the seasonal change in salinity are more or less similar to what has been described before. Fig. 5 represents a typical autumn distribution, with relatively high salinity throughout the Sound. During the two spring cruises the rivers were in flood, and Figs. 4 and 6 show the effects of freshwater drainage to an extreme degree. The

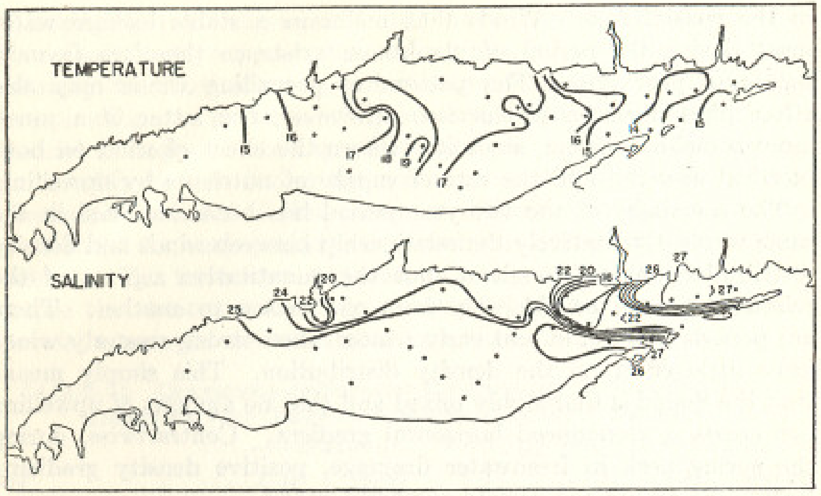

Figure 4. Surface temperature (°C) and salinity (°/oo), June 4 to 11, 1952. Dots indicate positions of observation.

outflow from the Connecticut River was largely responsible for reduced surface salinities in the eastern end of the Sound. This outflow was readily observed at sea as a muddy surface layer with clearly defined boundaries, in one case extending all the way from the mouth of the river to its exit through the eastern passes. The maximum observed westward extension of the muddy water was some five miles west of the mouth of the Connecticut River.

EAST-WEST MASS TRANSPORT

In this Sound, as in many other sounds and estuaries, current profiles of the sort shown in Fig. 7 generally indicate that the ebb is stronger than the flood in the surface layer but weaker than the flood in the bottom water. Averaging through a tidal cycle at the particular current station figured, it was estimated that the surface water moved 6.5 km east, the bottom water 2.2 km west, giving a total divergence of 8.7 km or 4.7 nautical miles during a complete tidal cycle.

LeLacheur and Sammons (1932) described an extensive set of surface current measurements at Bartlett Reef Lightship in the eastern end of the Sound. Their series extended from August 1929 to March 1930, during which period the monthly averages for nontidal

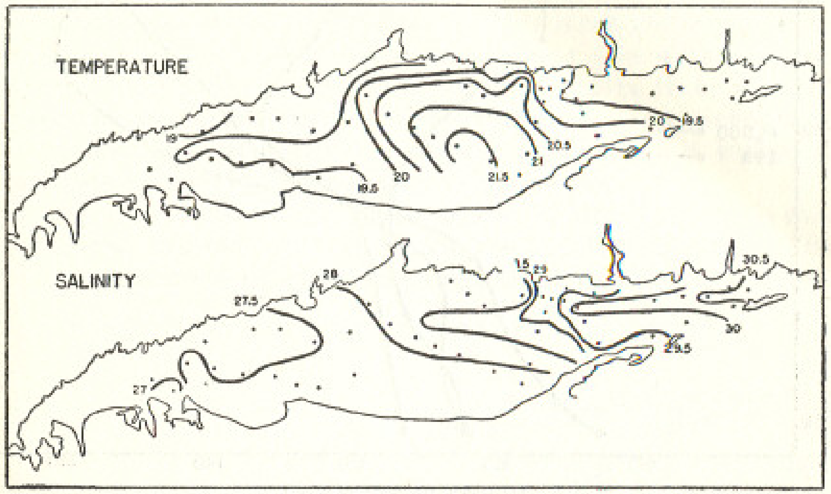

Figure 6. Surface temperature (°C) and salinity (°/oo) September 29 to October 8, 1952.

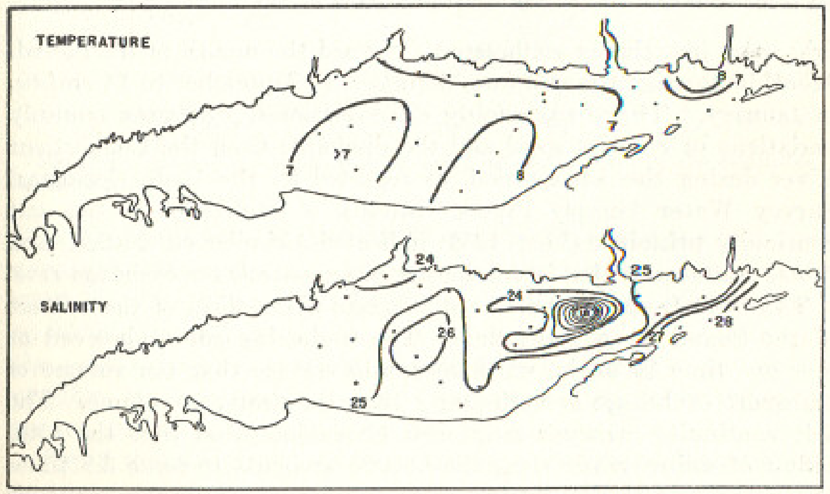

Figure 6. Surface temperature (°C) and salinity (°/oo), April 6 to 15, 1953.

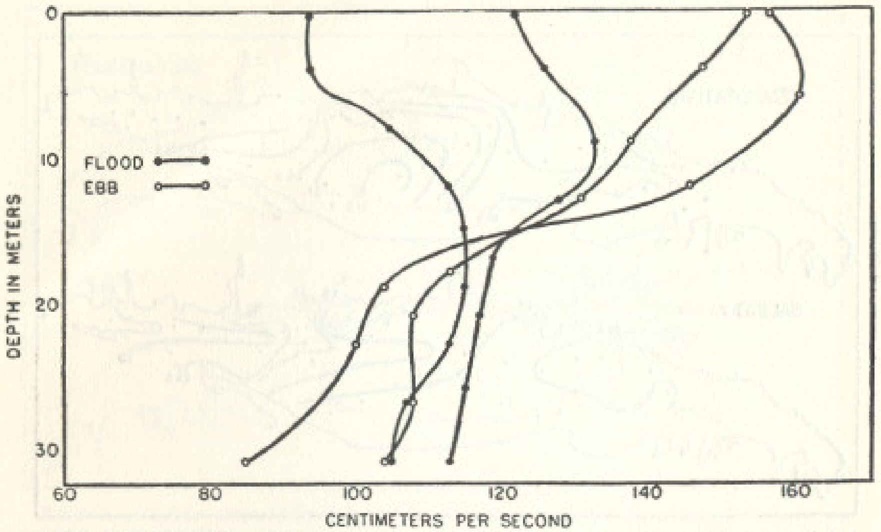

Figure 7. Observations of tidal currents at an anchor station in east-central Long Island Sound (Lat. 41°11.9′N, Long. 72°29.4′W) on April 6 and 7, 1953. Measurements were obtained with Pritchard and Burt (1951) current crosses at one-hour intervals, and the two sets of profiles shown were the ones nearest midebb and midflood.

drift were invariably southeasterly, toward the mouth of the Sound. Monthly averages ranged from 5 cm/sec. in September to 14 cm/sec. in January. There was a fairly close relationship between monthly variations in current speed and the discharge from the Connecticut River during the same period, as reported by the U. S. Geological Survey Water Supply Papers. Studies of exchange by the salt continuity principle (Riley, 1952) indicated a similar correlation, and it was postulated that freshwater drainage controls the exchange rate.

The total freshwater drainage averages about 35% of the volume of the Sound in a year's time. The freshening effect observed at any one time is slight, which means in essence that the volume of transport exchange is much larger than the drainage volume. The salt continuity analysis mentioned above indicated that the total inflow of saline water along the bottom amounts to some 3.8 times the volume of the Sound in a year. However, this gross exchange rate for the Sound as a whole is not necessarily applicable to any particular locality. The matter needs to be examined more fully, with particular emphasis on local variations in transport exchange, because much of the discussion of biological and chemical oceanography will

hinge upon the relative stability or mobility, as the case may be, of populations and chemical elements. Material is not yet available for a definitive treatment of the subject, including the effects of both currents and diffusion. However, diffusion appears to be relatively insignificant in estuaries (Pritchard, 1952), so that an evaluation of mass transport from observed currents should give a sufficiently accurate estimate of exchange for present purposes.

Important as nontidal transport may be, its actual volume is small compared with ordinary tidal oscillations, and it is readily modified by winds and density currents. Variations in residual flow of more

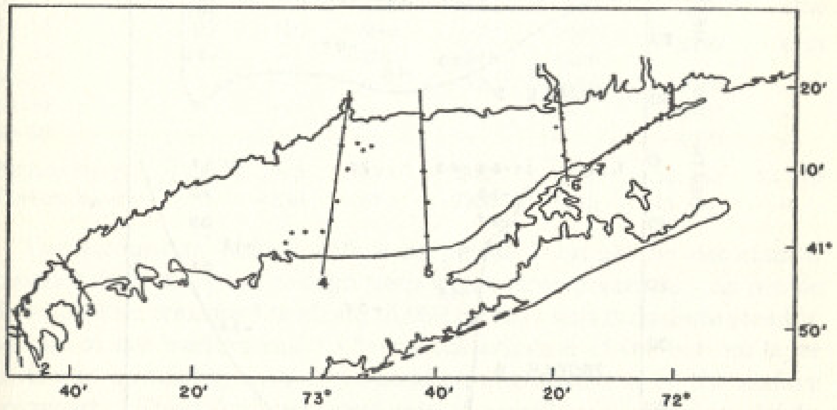

Figure 8. Positions of current stations (dots) and profiles used in the calculation of east-west transport.

than 20% have been noted between one tidal cycle and the next under conditions that seemed to be reasonably comparable. Much of the available station data includes only one tidal cycle. Hence there is no likelihood that all of the observed variation from one station to the next is real. The problem then is to determine whether it is possible, by statistically combining many stations, to derive a consistent pattern of mass transport.

Fig. 8 shows a pattern of stations that will be used for estimating east-west transport across the lines indicated in this figure. The data were obtained in the summers of 1890, 1917, 1929 (LeLacheur and Sammons, 1932) and 1953. The published data listed the duration and maximum velocity of flood and ebb at the surface and usually at two or three subsurface levels. Residual transport was calculated

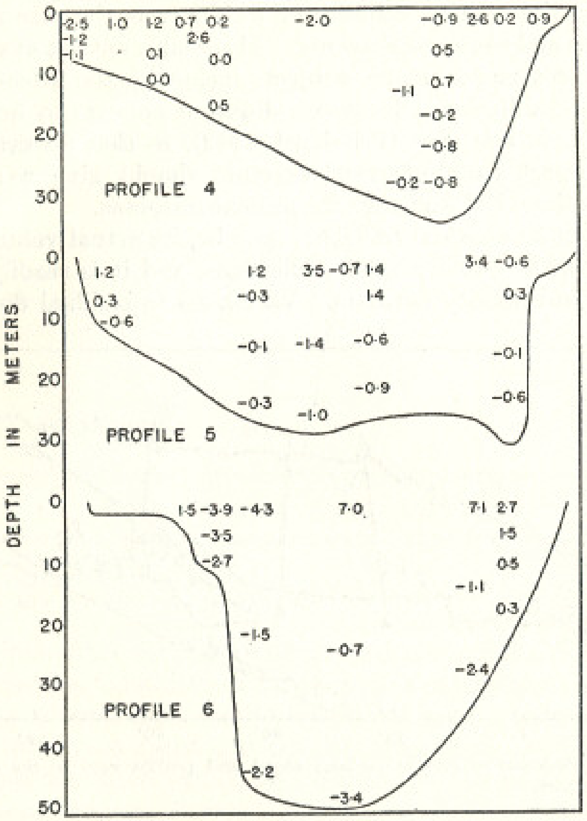

Figure 9. Estimated nontidal drift in nautical miles per tidal cycle through three of the profiles shown in Fig. 8. Positive numbers indicate eastward movement, negative numbers, westward.

by assuming that the mean velocity was two-thirds the maximum, multiplying this figure by the duration, and then taking the difference between flood and ebb. The 1953 data, consisting of half-hourly readings at four to six depths, were numerically integrated through the tidal cycle. Fig. 9 is an example of the type of results obtained. Transport is shown in nautical miles during a tidal cycle through three of the profiles previously identified by lines and numbers in Fig. 8.

The profiles provide a basis for calculating the volume transport through successive cross sections of the Sound. However, the amount

of variation from depth to depth and from one station to another introduces a subjective element into the interpolation. The results of the calculations, which must be regarded as tentative because of the many possible sources of error, are presented in Table III.

TABLE III. VOLUME TRANSPORT IN M³ PER SECOND THROUGH PROFILES 1 TO 7 (FIG. 8). POSITIVE NUMBERS INDICATE EASTWARD MOVEMENT; NEGATIVE ARE WESTERLY

Depth	Profile						
Meters	*1*	*2*	*3*	*4*	*5*	*6*	*7*
0–5	680	740	860	2100	4830	7130	9150
5–10	320	540	480	760	750	4870	4830
10–15	100	−130	−240	−70	−1460	2130	4910
15–20	—	—	−330	−840	−1380		
20–30	—	—	—	−1390	−2850	−4100	—
30–40	—	—	—	—	—	−6840	—
40–50	—	—	—	—	—	−4070	—
Surface layer	1100	1280	1340	2860	5580	14,130	18,890
Bottom layer	0	−130	−570	−2300	−5690	−15,010	—

The amount of bottom inflow at profile 7 cannot be determined because there were not enough deep water measurements. At profile 6 the bottom transport is about 15,000 m³/sec., and it declines steadily to zero at the western end. Clearly the attrition of the bottom layer is due to upwelling, with corresponding augmentation of the surface transport. The calculated mean rate of upwelling required to satisfy the principle of mass continuity is very small. In most of the central and western part of the Sound it is of the order of 5 to 8 cm/month and increases eastward to a maximum of about 45 cm in the vicinity of the passes.

It would appear that about 1100 m³/sec. enter the western end of the Sound and flow eastward as part of the surface layer. The latter is further augmented by freshwater drainage, amounting to about 300 m³/sec. in summer in the whole of the Sound. Thus the surface outflow at the eastern end should exceed the inflow by about 1400 m³/sec., but the transport estimates are not accurate enough to show this.

Fig. 10 is a diagram of transport exchange derived by combining theoretical considerations and calculated transport. The difference between surface and bottom flow, 1100 m³/sec. at the western end of the Sound, gradually increases to 1400 m³ in an easterly direction in

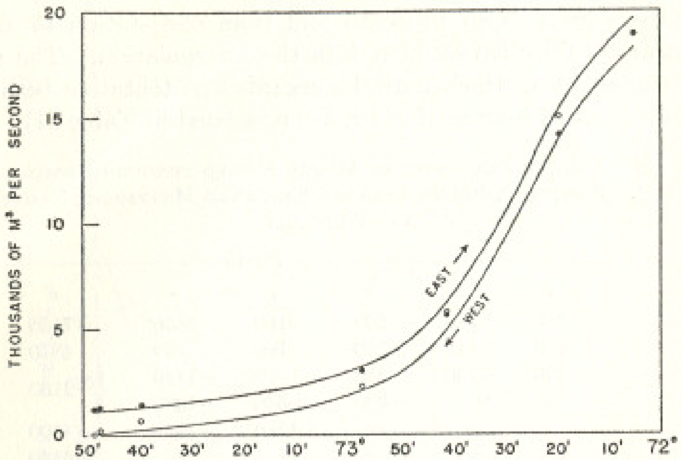

Figure 10. Calculated eastward surface transport (dots) and westward bottom transport (circles), and a schematic transport diagram derived from the calculations.

accordance with the known addition of fresh water. The total exchange increases as a smooth curve that is fitted as well as can be to the calculated transport data. The results, which appear to be internally consistent, agree more or less with previous calculations of inflow by the salt balance method. The latter gave a generalized estimate of 8400 m³/sec. during the summer of 1946, a figure that falls within the limits of variation of observed transport, as might be expected.

Fig. 10 indicates that transport declines rapidly toward the central area, which is particularly concerned in the present survey. Three months appear to be required, on the average, for bottom water to move through the 20 minutes of longitude occupied by the eight routine stations in the central basin, two months for the eastward transport of surface water. Obviously small parcels of water can be expected to move through the area much more rapidly than the general average. Nevertheless, the central water mass appears to be much less mobile than might have been supposed on the basis of salt balance calculations for the Sound as a whole.

There is the further implication in these results that freshwater dilution of the central and western parts of the Sound is largely of local origin. If any considerable fraction of the drainage from the large eastern rivers penetrated into the central basin, the compensating

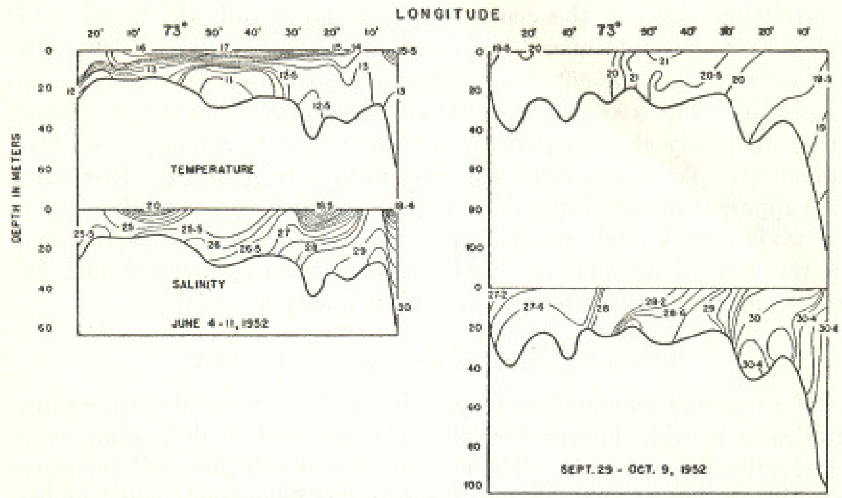

Figure 11. Longitudinal profiles of temperature (°C) and salinity (°/₀₀) in Long Island Sound.

transport would have to be much larger. So far as direct observations are concerned, no evidence has thus far been found either of a coastwise density current originating at the mouth of the Connecticut River or of offshore eddies moving west from the main mass of river outflow.

Fig. 11 shows two sets of longitudinal profiles of temperature and salinity. They are included here rather than in the section on horizontal distribution for convenient examination of the relations between mass transport and the distribution of conservative properties. The salinity diagrams clearly show the westward penetration of saline water along the bottom as well as the resulting slope of the isohaline surfaces. There are a few isolated pockets of water of different salinity that suggest variations in transport of one kind or another. There are also indications in routine station data for the central part of the Sound that the influx of saline bottom water may be somewhat intermittent.

The effects of east-west transport may be detected in the temperature distribution during the autumn cruise (see Fig. 11), but they are not apparent in the spring cruise. In general, local surface heating and cooling are more important than transport in determining the character of the temperature distribution. In the next section, local temperature data in the central part of the Sound will be used to calculate coefficients of vertical eddy conductivity. The calculation will assume that the effects of transport and horizontal diffusion are negligible. The seasonal rate of change in temperature is many times larger than any changes that could be produced by average east-west transport. To that extent the assumption is justified. However, it is apparent in examining Fig. 11 that more serious errors of a degree or two in a week might occur if there were occasional large fluctuations in the rate of movement. Such errors can be eliminated only by averaging the data for periods of several weeks or months.

VERTICAL EDDY CONDUCTIVITY

The seasonal temperature progression and the vertical temperature gradients provide information that can be used to determine eddy conductivity coefficients. Their evaluation at this time will illustrate certain aspects of the thermal cycle in the Sound and will provide data for later analyses of nutrient and production problems. The later applications will presuppose that coefficients of vertical eddy conductivity are the same as the diffusivity coefficients. This is a debatable point. There are certain theoretical reasons for believing that thermal coefficients should be larger. On the other hand, Riley (1951) examined eddy coefficients calculated from both temperature and salinity distribution and detected no essential difference. There is undoubtedly less error in using temperature for this purpose than in attempting the more complicated and unwieldy determination of eddy diffusivity as indicated by salinity distribution.

Ignoring the effects of advection and horizontal mixing, the vertical flux of heat attendant upon surface heating and cooling may be written

$$F = -\frac{A}{\rho} \cdot \frac{\partial T}{\partial z} \cdot \qquad (1)$$

F is the rate of passage of heat through one square centimeter of horizontal surface at any specified depth. A is the coefficient of eddy conductivity, ρ the density, and $\partial T/\partial z$ the vertical temperature gradient.

Under the simplifying assumptions that have been adopted, the heat flux is readily obtained from observed water temperatures. Suppose that, in a total depth of water of 20 m, the lower 10 m increases half a degree in a week's time. The amount of heat that has

passed through 1 cm² of horizontal area at a depth of 10 m is 500 g cal., and

$$F = \frac{500}{7 \times 86400} = 0.83 \times 10^{-3} \text{ g cal/sec.}$$

Further, suppose that during this time the temperature in the vicinity of the 10 m level—say between 8 and 12 m—has shown an average negative gradient of 0.5° C in the 4 m interval. The density is about 1.02. Applying equation (1) in finite difference form,

$$0.83 \times 10^{-3} = \frac{-A}{1.02} \cdot \frac{-0.5}{4 \times 10^2},$$

$$A = 0.68 \text{ g cm}^{-1}\text{sec}^{-1}.$$

In applying this method to the seasonal temperature cycle in the Sound, it is necessary to develop an analysis that takes due account of both the inherent limitations of the data and the further uses for the calculated coefficients. First, it is apparent that relatively small sampling and analytical errors can assume serious proportions in any calculation requiring the use of gradients. The only way to minimize such errors is to combine data. It is therefore advisable to calculate average coefficients over considerable periods of time and also to combine stations when this can be done in a reasonable way. In the present case, Sts. 1, 7, and 8 are of comparable depth and will be combined into a single analysis. Also Sts. 2 and 4 will be combined, as will Sts. 3 and 5. The fact that the stations in each group are of similar depth is important for present purposes, and there is another and more practical reason for combining them in this way. Oxygen and phosphorus were sampled at approximately the same depths within each of the proposed groups. It will be necessary later to calculate the rate of exchange of these elements between the depths sampled, and therefore the eddy coefficients must be calculated for the same depth ranges. This is done by averaging vertical temperature gradients between 0 and 5 m, 5 and 10 m, etc., and determining the heat flux through the midpoints of each stratum, namely 2.5, 7.5, and 15 m. This operation is facilitated by the fact that temperatures were obtained from bathythermograms and could be read easily at any desired depth.

With regard to the seasonal cycle, eddy coefficients are most easily obtained during the spring and summer warming period, when vertical temperature gradients are fairly pronounced. This period may be analyzed as a whole or broken up into smaller units. With later biological needs in mind, it was desirable to divide into three units. The first was relatively short, encompassing the spring diatom flowering or as much of it as occurred within the limits of vernal warming. The remainder was divided into two approximately equal periods, the first constituting post-flowering spring conditions, the second including most of the summer.

About the middle of August or the first of September, the water column began to cool from surface to bottom. Vertical temperature gradients were small and variable, but the average condition for the next three months was a slight negative gradient. Under such conditions vertical eddy conductivity could not be calculated and probably was less important than convection in controlling the seasonal temperature change.

During the remainder of the cooling period, from mid-November until late winter, the common condition was isothermal water or a slight positive gradient. Coefficients of eddy conductivity could be calculated, but they tended to be large and highly variable. What appeared to be a reasonable average could be obtained by combining large quantities of data, although there is some liklihood that convection continued to be important during this period. With reservations because of this possibility of error, generalized averages for winter eddy conductivity are as follows: 0 to 5 m, 24 g cm^{-1}sec.$^{-1}$;

TABLE IV. Coefficients of Vertical Eddy Conductivity (g cm^{-1}sec^{-1}) Calculated from Observed Temperature Distribution and Seasonal Cycles

Stations	*1, 7, 8*		*2, 4*			*3, 5*		
Depth range in meters	*0–5*	*5–9*	*0–5*	*5–10*	*10–20*	*0–5*	*5–15*	*15–25*
1952								
Mar. 5–May 21	1.36	0.51	1.76	1.80	1.16	2.12	2.08	0.96
May 21–Aug. 19	0.85	0.22	0.75	0.68	2.62	1.40	1.33	2.77
1953								
Feb. 10–Mar. 16	2.13	3.55	1.06	1.26	2.70	1.10	1.62	2.70
Mar. 16–May 18	1.10	0.35	0.73	0.79	3.36	1.90	1.75	1.97
May 18–Aug. 25	0.36	0.19	0.79	0.92	1.30	2.16	1.34	1.41
1954								
Jan. 25–Feb. 24	5.5	1.81	1.13	2.26	2.33	6.1	6.1	2.60

5 to 12.5 m, 9.8; 12.5 to 20 or 25 m, 9.0. These figures are based on all available station data from mid-November until the end of the winter cooling period for both years of observation. Calculated values for the spring-summer warming period are shown in Table IV. All of these coefficients are much less than those in winter. Minimum values are found in the surface layer or at mid-depths, depending upon the position of the thermocline. The coefficients are not seriously reduced in the deepest water examined, presumably because of strong tidal mixing.

RADIATION AND TRANSPARENCY

Radiation

Data on total daily radiation in g cal/cm^2 are recorded by the U. S. Weather Bureau at two stations on Long Island. Local variations are important enough so that it was desirable to obtain estimates of radiation nearer the scene of oceanographic operations. The New Haven Weather Bureau is favorably situated but records only the number of hours of sunshine and the amount of cloud cover. However, statistical analysis of a two-year set of data from New York City indicated that measurements of the total amount of sunshine recorded at New York could be converted to estimates of total radiation with sufficient accuracy for present purposes. The analysis was not significantly improved by including a term for percent of cloud cover, presumably because the effect of clouds on radiation depends more upon their quality than upon the total amount.

In estimating total radiation for the New Haven area, the percentages of total possible sunshine were averaged for the periods between successive weekly cruises and applied to a graph constructed on the basis of the New York analysis. The graph contained a systematic correction for the effects of metropolitan haze in New York, which was obtained by adjusting the New York figures to the mean values obtained at the Long Island Stations. The net result is that New Haven estimates may contain a small systematic error, but fluctuations from week to week in the area of oceanographic operations should be relatively accurate. Fig. 12A shows the radiation estimates for New Haven, together with monthly averages of the pyrheliometric measurements at the two Long Island stations.

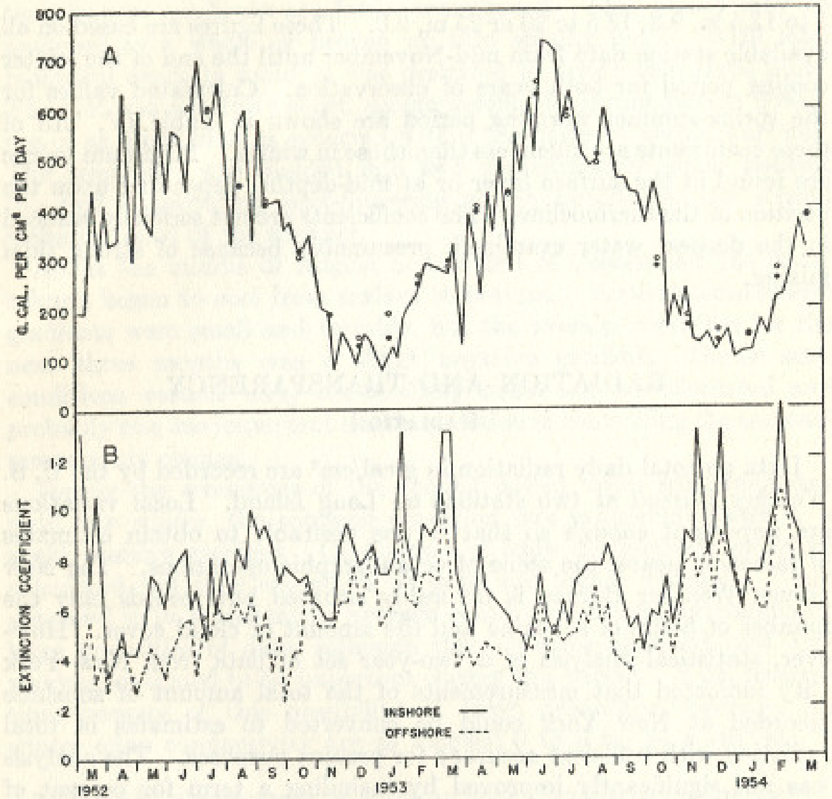

Figure 12. A. Direct plus diffuse solar radiation. Dots and circles are monthly averages of pyrheliometric measurements by the U. S. Weather Bureau at two stations on Long Island. Solid line shows weekly averages of estimates obtained for the New Haven area by methods described in the text. B. Estimates of average extinction coefficients in the inshore and offshore waters based on Secchi disc readings.

Measurement of Transparency

Two methods have been used to measure the transparency of Long Island Sound waters. The first is the so-called Secchi disc reading; a 20 cm white disc is lowered in the water, and the mean depth of disappearance and reappearance is recorded. Results are shown in Fig. 12B as estimated average extinction coefficients at inshore and offshore stations, determined by application of the Poole and Atkins (1929) formula,

$$K = \frac{1.7}{D}, \tag{2}$$

where K is the extinction coefficient per meter and D is the Secchi disc depth in meters. The second measurement was a photoelectric determination of the extinction coefficient of red light in each water sample, using a Klett 66 filter, with ordinary laboratory distilled water as the reference blank. The primary purpose was to obtain a correction factor for turbidity in colorimetric determinations of phosphate. The measurements were made in a 4 cm cell, which is not long enough for highly accurate determination of transparency. Nevertheless, they are useful for detecting gross changes in transparency with depth.

In other papers of this series it will be necessary to calculate the light intensity at various depths. In general, the Secchi disc readings are sufficient. However, in particular cases the extinction coefficients varied so markedly with depth that a correction factor seemed desirable if it could be obtained in a reasonable way. According to Jerlov (1951), the two types of measurements are not strictly comparable. Moreover, a measurement with red light minimizes the effect of dissolved "yellow substance". Notwithstanding these technical objections, there is a moderately satisfactory empirical relation between the two sets of measurements. Statistical comparison of some 460 surface water extinction coefficients with the corresponding Secchi disc readings shows a correlation of 0.74 and yields the equation

$$K = .96K_1 + .246\,, \tag{3}$$

where K is derived from the Secchi disc reading according to equation (2) and K_1 is the extinction coefficient for red light as determined above.

Analysis of Factors Influencing Transparency

It has been recognized for many years that light extinction in sea water is a complex process involving absorption and scattering by the water, dissolved solids, and a variety of suspensoids. The latter include plankton, planktogenic detritus, and—particularly near land—bottom sediments in suspension and particles of terrestrial origin.

In Long Island Sound all of these factors are expected to be operative. In waters of this type, no one has tried very hard to formulate a quantitative expression for the various factors that govern transparency, although it is not only a problem of considerable intrinsic interest but also has an important bearing on several aspects of biological oceanography. Sufficient data have now been amassed for a cursory analysis of transparency in the central part of the Sound.

First to be considered is the general problem of absorption and scattering of light by plankton. Rodhe (1948) and others have measured the extinction coefficients of pure algal cultures and have found, in any one species, a nearly linear relationship between extinction and cell number or chlorophyll concentration. Clarke (1946) compared pigment measurements of natural phytoplankton populations on Georges Bank with Secchi disc estimates of transparency and demonstrated a significant relationship between them. In an attempt to refine this comparison, Fig. 13 presents a group of data that is believed to represent the relation in its simplest possible form, namely (a) observations during the spring flowering on Georges Bank, a period when a rapid reduction in transparency coincided with the increase in phytoplankton, and (b) all available observations in deep oceanic waters, where it may be presumed that particulate matter of nonplankton origin is minimal. Phytoplankton concentrations are expressed as Harvey units of plant pigments per liter, a type of measurement that will be discussed in the next paper of this series. The extinction coefficients were obtained photometrically in a few cases, but most of the values are converted from Secchi disc readings according to equation (2).

The semilogarithmic plot in Fig. 13A is necessary in order to picture the total range of phytoplankton variations, which encompass more than three orders of magnitude. However, the method of presentation obscures the form of the relationship. Hence the smooth curve that has been fitted to the data is repeated in arithmetic form in Fig. 13B. There it is apparent that, unlike algal cultures, the relation between phytoplankton concentrations and extinction coefficients in the sea is nonlinear.

The smooth curve in Fig. 13 corresponds to the equation

$$K = .04 + 2 \times 10^{-6} P + 2 \times 10^{-4} P^{2/3}, \qquad (4)$$

where K is the extinction coefficient and P is Harvey units of plant pigments per cubic meter. The constant in the equation is an average absorption coefficient for visible light in pure water. The terms for plant pigments were derived empirically as a reasonable fit for the data in Fig. 13 and other more recent observations that should be considered in this connection.

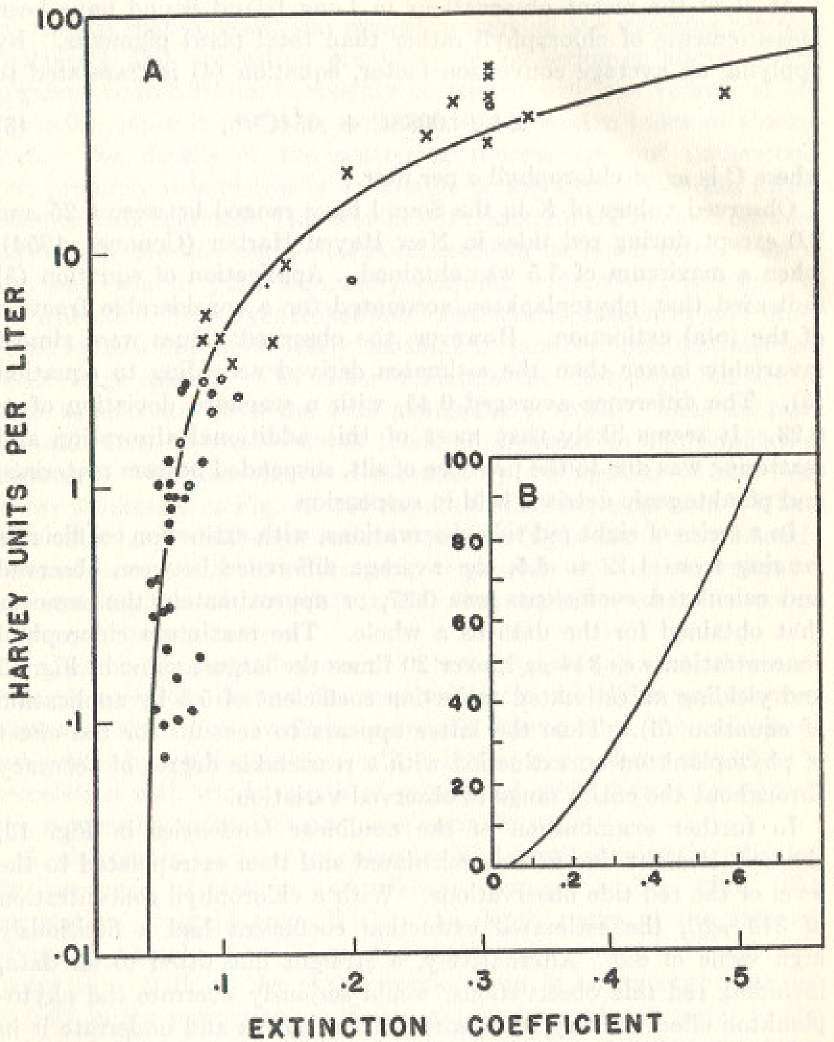

Figure 13. Relationship between extinction coefficients and phytoplankton pigments as determined by the Harvey method (1934). A. Semilogarithmic plot of observations obtained in the Sargasso Sea (dots), continental slope waters off the New England coast (circles), and the spring flowering period of March and April on Georges Bank (x's). B. Arithmetic plot of the smooth curve in Fig. 13A.

Most of the recent observations in Long Island Sound have been measurements of chlorophyll rather than total plant pigments. By applying an average conversion factor, equation (4) is translated to

$$K = .04 + .0088C + .054C^{2/3}, \tag{5}$$

where C is μg of chlorophyll *a* per liter.

Observed values of K in the Sound have ranged between 0.25 and 2.0 except during red tides in New Haven Harbor (Conover, 1954), when a maximum of 5.5 was obtained. Application of equation (5) indicated that phytoplankton accounted for a considerable fraction of the total extinction. However, the observed values were almost invariably larger than the estimates derived according to equation (5). The difference averaged 0.45, with a standard deviation of $\pm$ 0.23. It seems likely that most of this additional absorption and scattering was due to the presence of silt, suspended bottom materials, and planktogenic detritus held in suspension.

In a series of eight red tide observations, with extinction coefficients ranging from 1.12 to 5.5, the average difference between observed and calculated coefficients was 0.37, or approximately the same as that obtained for the data as a whole. The maximum chlorophyll concentration was 314 μg/l, over 20 times the largest value in Fig. 13 and yielding an estimated extinction coefficient of 5.3 by application of equation (5). Thus the latter appears to account for the effect of phytoplankton on extinction with a reasonable degree of accuracy throughout the entire range of observed variation.

In further examination of the nonlinear tendencies in Fig. 13, the best straight line fit was calculated and then extrapolated to the level of the red tide observations. With a chlorophyll concentration of 314 μg/l, the estimated extinction coefficient had a fictitiously high value of 8.5. Alternatively, a straight line fitted to all data, including red tide observations, would seriously overrate the phytoplankton effect in very transparent ocean waters and underrate it in waters with moderately high extinction coefficients of 0.5 to 1.0.

There remains the question of the reason for differences between natural phytoplankton populations and pure algal cultures. In the latter, both scattering and absorption are expected to be proportional to cell number, so that the results approximate Beer's law, although the process is much more complicated than in the case of materials in solution. In a mixed phytoplankton population containing cells

of various shapes, sizes, and optical properties, the relations between scattering and absorption are likely to be seriously altered. The pigment concentration is roughly correlated with the volume of cell material; hence it might be expected to serve as an index of absorption. The details of the scattering process are not understood, but probably it is primarily a function of the area of the cells rather than of their volume. This is sufficient reason for the relation between phytoplankton and extinction coefficients to be much less precise in the sea than in pure laboratory cultures. Moreover, it seems to be true, as a broad generalization, that naked flagellates and other small species are more nearly constant in number than diatoms and other large species. The latter dominate the flowerings but are of little significance when the total population is small. Thus the ratio of total area to total volume tends to decrease as the population increases. Whether this is a satisfactory explanation of the nonlinear tendencies in Fig. 13 will require further investigation. Whatever may be the full explanation, equation (6) presents a sufficiently accurate account of the relationship to provide an entering wedge for the investigation of the other factors that are involved in the transparency of coastal waters.

It has been suggested above that bottom materials and other nonliving particulate matter are responsible for the difference between observed extinction coefficients in the Sound and the calculated extinction by phytoplankton cells. If this is so, there should be a correlation with winds, depth of water, and other factors that affect the suspension and settling rates of such materials.

A relationship between transparency and depth of water is indicated by inspection of inshore and offshore averages in Fig. 12B. The correlation is significant. Within the depth range of the stations occupied, namely 4 to 30 m, the statistically computed correction factor is $-$ 0.01 *d*. In other words, there is an average decrease of 0.01 in the extinction coefficient per increase of one meter in the depth of the station. The computation makes due allowance for differences in phytoplankton concentration and thus it is intended to express the relation between depth and non-living suspended matter.

At certain stations the occurrence of surface eddies of reduced salinity coincided with low Secchi disc readings. There is no doubt that silt effects are appreciable in the Sound. However, the obvious effects in the central basin are of limited area, and general correlations with recent rainfall are too low to warrant inclusion in the analysis.

In considering the effects of wind, it was apparent that transient storms might reduce the transparency materially, while the average wind speed over a longer period might also be important in controlling the suspension of fine materials. In investigating this problem, the mean wind speed was tabulated for the day of observation as well as the preceding day, the preceding week, and month. The means were then correlated with a residual extinction coefficient, obtained by removing the chlorophyll and depth effects from the observed extinction coefficients and averaging the residuals for each of the 96 days of observation. All correlations were statistically significant. The mean wind during the preceding week yielded the highest correlation and was adopted for subsequent analysis.

It is expected that the stirring action of the wind will be materially reduced by vertical stability. Using the difference between bottom and surface density as a rough expression of stability, it was possible to demonstrate a small but significant negative correlation with residual extinction coefficients. However, to be satisfactory from the physical oceanographic point of view, wind and stability should be combined into a single expression which recognizes the existence of a stirring effect in water of indifferent stability and a gradual reduction of that effect with increasing stability. Development of such an expression would be facilitated by a more precise knowledge of the nature of the physical relationship. Alternatively, by trial and error and using statistical methods to test the results, it was possible to formulate an expression that was empirically satisfactory but not necessarily correct from a physical standpoint. It was $.018\ W/.3 + \Delta\sigma_t$, where W is the mean wind speed in miles per hour during the week preceding the observation, $\Delta\sigma_t$ the mean increase in density from surface to bottom at the beginning and end of the week.

Tidal currents also presumably affect transparency in shallow coastal waters. Mean tidal currents have been computed from data on tidal height in the Sound (Riley, 1952). In the central region the current speed is 19 cm/sec., averaged through a complete tidal cycle at a time of average tidal height. The total variation between lowest neaps and highest spring tides is about 13 to 28 cm/sec. Tidal currents, estimated from the predicted height of the tide for the two-year period of the investigation were averaged by weeks. There was a significant correlation between residual transparency and the

mean current speed during the preceding week, and the average effect was computed to be $.013\ v$, where v is the average current in centimeters per second. There was no evidence that the tidal factor was significantly modified by variations in vertical stability, although observations at current stations suggested that stability has a slight effect on the vertical velocity gradient.

Finally, inspection of the data indicated an increase in residual extinction coefficients during the summer months which could not be accounted for by any of the variables thus far examined. Two possible explanations may be offered. First, there is a seasonal variation in the ratio of chlorophyll content to phytoplankton organic matter (see the accompanying paper by Harris and Riley), so that chlorophyll may not be entirely acceptable as an index of light extinction by phytoplankton. In the second place, there is evidence, visual and otherwise, of considerable quantities of organic detritus in the water during the summer. Nets are often clogged with slimy brown material containing little recognizable plant substance. It is apparent (see the accompanying paper on PRODUCTION AND UTILIZATION) that a considerable proportion of the spring phytoplankton growth settles to the bottom and is not immediately decomposed. Summer on the other hand is a period of active bacterial growth, accompanied by reduced oxygen in the lower part of the water column and increasing phosphate concentrations. Bacterial activity appears to exceed production, and presumably the excess spring growth is largely decomposed during this period. The bottom appears to be a more important site of decomposition than the water column, but in both cases the intermediate steps in decomposition can be expected to produce light, finely divided material that is easily carried in suspension.

Both of the phenomena that have been described might be expected to influence transparency, although neither can be rated precisely in quantitative terms. Examination of the data shows that the summer increase in residual extinction coefficients is approximately proportional to the oxygen deficit, defining the latter as the percent undersaturation of the bottom water or the percent difference between surface and bottom when the surface layer is undersaturated. It is statistically justifiable to use the oxygen deficit in the analysis, but with some reservations as to the meaning and validity of the term.

The final form of the equation may now be written

$$K = f(C) - .01d + .018 \frac{W}{.3 + \Delta\sigma_t} + .013v + .0063f(O_2) + .05 . \quad (6)$$

The chlorophyll relationship in equation (5), but not including the constant, is specified $f(C)$, and the percentage deficit of oxygen that has just been described is written $f(O_2)$. Other factors have been sufficiently described above. The final constant is statistically derived. It differs little from the absorption coefficient for pure water, but the form of the equation is such that this fact has little significance.

Discussion

Fig. 14A compares observed and calculated extinction coefficients, and Fig. 14B shows the seasonal variation in the individual components of equation (6). There is a correlation of 0.66 between observed and calculated values. The correlation is highly significant statistically but not very precise by absolute standards. The general seasonal trend of the calculated curve is realistic, but short period fluctuations show considerable error. Perhaps the agreement would have been closer if a more accurate method had been used to measure transparency. Other random errors undoubtedly result from local variability in density, oxygen, etc. Observations at five stations can hardly be expected to give typical, average values for the region on a particular date.

In particular cases there are readily observable reasons for disagreement. For example, on January 27 and December 16, 1953, the observed extinction coefficients exceeded the calculated ones by about 0.50. Both sets of observations were preceded by heavy northwest storms. In both cases it was only the northern inshore stations that were seriously affected. In other words, immediate wind effects can be more important than is indicated by general statistical comparison. However, to fit them into the general equation would require an elaborate treatment of both the speed and direction of the wind in relation to the particular area sampled.

With the possible exception of the phytoplankton treatment, all of the terms in equation (6) are of local significance only. Depth, current, and wind-stability factors presumably are subject to great variation according to the kind of bottom sediments available for suspension. This would introduce an element of empiricism even if the physical oceanographic processes were expressed in a fully

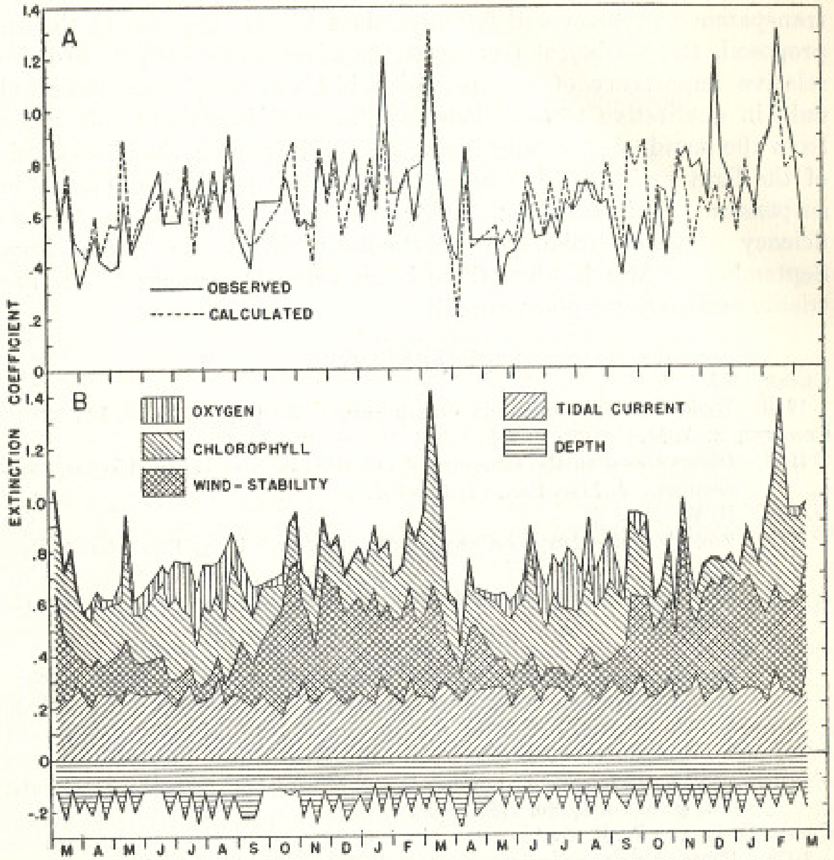

Figure 14. A. Comparison of weekly averages of the observed extinction coefficients in central Long Island Sound with average values calculated from equation (6). B. Relative importance of the individual components in equation (6). Positive factors are plotted one above the other. Thus the uppermost curve represents the sum of all four positive factors, and the calculated extinction coefficient is the uppermost curve minus the depth correction.

acceptable manner. A revision of the analysis may be desirable when the remainder of the survey has been completed. The latter will increase the total depth range to 100 m and hence will provide a better set of data for testing nonlinear characteristics or possible combination with other factors, and will also permit examination of a wider variety of tidal velocities, bottom sediments, and drainage conditions.

Despite the need for a better theoretical understanding of the transparency problem and for more data to test any theory that is proposed, the statistical treatment has given some insight into the relative importance of factors which hitherto have been discussed only in qualitative terms. Perhaps the most important conclusion from the standpoint of later biological applications is that two-thirds of the light extinction is caused by nonliving particulate matter in suspension. This condition must have a serious effect on the efficiency of plant production, particularly during the period from September to March when there is an abundant supply of the nutrients necessary for plant growth.

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