

CHAPTER III

Physical Properties of Sea Water

The properties of pure water are unique in comparison with those of other liquids, and the nature of our physical environment—that is, the characteristics of the oceans, the atmosphere, and the land—is in many ways dependent upon the peculiar properties of water. The “fitness” of water for the manifold needs of living organisms has been pointed out by physiologists and ecologists (for example, Henderson, 1913, Bayliss, 1927). Table 7 shows some of the characteristics which are important in this respect for both fresh water and sea water.

The unique character of water is further emphasized by the fact that, according to studies on related compounds, pure water should freeze at about -150°C and boil at -100°C . The chemical constitution of water offers no explanation for these anomalies and certain other deviations from the theoretical properties. However, it has been found that liquid water is not made up of individual H_2O molecules but that it is polymerized—that is, multiple groups containing one, two, or three elementary H_2O molecules may exist. These are referred to as monohydrol, dihydrol, and trihydrol. The relative proportions of the three forms depend upon the temperature, the immediate past history of the water, and other factors. The degree of polymerization decreases with increasing temperature. The existence of the water polymers is helpful in explaining certain of the peculiar properties of water such as the high melting and boiling points. The degree of polymerization has been thought to have certain physiological effects (Barnes and Jahn, 1934).

The discovery of isotopes of hydrogen and oxygen has modified our conception of “pure” water. All naturally occurring water contains small but variable amounts of heavy hydrogen (deuterium) and heavy oxygen. These modify the density and other properties, but, since their concentrations are extremely small, the effect is slight. As the understanding of this problem increases, it may become desirable to define more exactly certain physical units that are based on the properties of water, such as the liter and calorie, by taking into account the isotopic composition of water. The amounts of the heavy isotopes vary somewhat, depending upon the source of the water. Wirth, Thompson, and

TABLE 7
CERTAIN PHYSICAL PROPERTIES OF WATER
(In part after Fleming and Revelle, 1939)

Property	Comparison with other substances	Importance in physical-biological environment
Heat capacity	Highest of all solids and liquids except liquid NH ₃	Prevents extreme ranges in temperature Heat transfer by water movements is very large Tends to maintain uniform body temperatures Thermostatic effect at freezing point owing to absorption or release of latent heat
Latent heat of fusion	Highest except NH ₃	Large latent heat of evaporation extremely important in heat and water transfer of atmosphere
Latent heat of evaporation	Highest of all substances	Fresh water and dilute sea water have their maximum density at temperatures above the freezing point. This property plays an important part in controlling temperature distribution and vertical circulation in lakes
Thermal expansion	Temperature of maximum density decreases with increasing salinity. For pure water it is at 4°C	Important in physiology of the cell Controls certain surface phenomena and drop formation and behavior
Surface tension	Highest of all liquids	Obvious implications in both physical and biological phenomena
Dissolving power	In general dissolves more substances and in greater quantities than any other liquid	Of utmost importance in behavior of inorganic dissolved substances because of resulting high dissociation
Dielectric constant	Pure water has the highest of all liquids	A neutral substance, yet contains both H ⁺ and OH ⁻ ions
Electrolytic dissociation	Very small	Absorption of radiant energy is large in infra red and ultraviolet. In visible portion of energy spectrum there is relatively little selective absorption, hence is "colorless." Characteristic absorption important in physical and biological phenomena
Transparency	Relatively great	Although important on small scale, as in living cells, the molecular processes are far outweighed by eddy conduction (see text)
Conduction of heat	Highest of all liquids	

Utterback (1935) found that distilled water prepared from sea-water samples was, on the average, 1.4×10^{-6} greater in density than distilled tap water. Sea water from areas of great dilution showed a somewhat smaller anomaly—namely, 0.2 to 0.5×10^{-6} in density. Samples from the depths of the ocean were higher than the average. Swartout and Dole (1939) found that the density of water distilled from a sea-water sample was 1.7×10^{-6} greater than that prepared from Lake Michigan water. They also found that the ratio between hydrogen and deuterium in sea water was about 7000:1. Rain water, and consequently fresh water, has a lower proportion of the heavy isotopes than sea water, since a certain fractionation takes place in the process of evaporation. The field of isotope study is relatively new, and as yet little is known about the distribution or significance of the isotopes. Their variability is apparently rather small and consequently will not significantly affect the magnitude of the properties to be discussed.

Information as to the physical properties of distilled water is found in comprehensive handbooks (for example, Dorsey, 1940) or in the International Critical Tables. A number of these physical properties depend upon two variables, temperature and pressure, but in the case of sea water a third variable has to be considered; namely, the salinity of the water, which will be defined and discussed below. Some of the properties, such as compressibility, thermal expansion, and refractive index, are only slightly altered by the presence of dissolved salts, but other properties that are constant in the case of distilled water, such as freezing point and temperature of maximum density, are dependent on salinity in the case of sea water. Furthermore, the presence of dissolved salts adds a few new characteristics to sea water, such as osmotic pressure. General surveys of the physical properties of sea water have been given by Krümmel (1907), Matthews (1923), Harvey (1928), Johnstone (1928), and Thompson (1932).

Another important aspect of the physical characteristics that has to be considered when dealing with water as it occurs in nature, regardless of whether fresh water or sea water is concerned, is that several important processes are greatly modified by the presence of minute suspended particles or by the state of motion. Thus, the absorption of light in lakes or in the sea is entirely different from the absorption of light in distilled water or in "pure" sea water, because the waters encountered in nature always contain suspended matter that causes increased scattering of the light and consequently increased absorption in layers of similar thickness. The processes of heat conduction, chemical diffusion, and transfer of momentum from one layer to another are so completely altered in moving water that, for water under natural conditions, the coefficients which have been determined under laboratory conditions must be replaced by corresponding "eddy" coefficients that depend

upon the presence of eddies. Some of the physical properties of sea water, therefore, depend only upon the three variables, temperature, salinity, and pressure, which can all be determined with great accuracy, whereas others depend upon such variables as amount of suspended matter or character of motion, which at present cannot be accurately determined. Before turning to a discussion of the physical properties and their relations to these variables, we shall discuss the salinity of the sea water.

Salinity and Chlorinity

In the chapter on chemical oceanography the composition of the dissolved constituents in sea water is considered in detail. Our present interest is only in the more abundant substances whose concentration will affect the physical properties. It has been found that, *regardless of the absolute concentration, the relative proportions of the different major constituents are virtually constant*, except in regions of high dilution (low salinity), where minor deviations may occur. From this rule it follows that any one of the major constituents may be used as a measure of the others and of the total amount of salt, and that water samples having the same total salt content, regardless of their source, are virtually identical in their physical properties.

Owing to the complexity of sea water, it is impossible by direct chemical analysis to determine the total quantity of dissolved solids in a given sample. Furthermore, it is impossible to obtain reproducible results by evaporating sea water to dryness and weighing the residue, because certain of the materials present, chiefly chloride, are lost in the last stages of drying. These difficulties can be avoided by following a technique yielding reproducible results which, although they do not represent the total quantity of dissolved solids, do represent a quantity of slightly smaller numerical value that is closely related and by definition is called the *salinity* of the water. This technique was established by an International Commission (Forch, Knudsen, and Sørensen, 1902), and on the basis of its work the salinity is defined as *the total amount of solid material in grams contained in one kilogram of sea water when all the carbonate has been converted to oxide, the bromine and iodine replaced by chlorine, and all organic matter completely oxidized*.

The determination of salinity by the method of the International Commission is rarely if ever carried out at the present time because it is too difficult and slow, but, owing to the constant composition of the dissolved solids, the determination of any of the elements present in relatively large quantity can be used as a measure of the other elements and of the salinity. Chloride ions make up approximately 55 per cent of the dissolved solids and can be determined with ease and accuracy by titration with silver nitrate, using potassium chromate as indicator.

The empirical relationship between salinity and chlorinity, as established by the International Commission, is

$$\text{Salinity} = 0.03 + 1.805 \times \text{Chlorinity}.$$

The chlorinity that appears in this equation is also a *defined* quantity and does not represent the actual amount of chlorine in a sample of sea water. Both salinity and chlorinity are always expressed in grams per kilogram of sea water—that is, in parts per thousand, or per mille, for which the symbol ‰ is used.

CHLORINITY. In the titration with silver nitrate, bromides and iodides are precipitated together with the chlorides, but in the computation it is assumed that they are chlorides. Chlorinity was therefore originally defined as *the total amount of chlorine, bromine, and iodine in grams contained in one kilogram of sea water, assuming that the bromine and the iodine had been replaced by chlorine*. This definition introduces a chlorine-equivalent that is dependent upon the atomic weights used in preparing the standard solutions. Since the time of the work of the International Commission, there have been changes in the atomic weights, and the relation between salinity and chlorinity as defined above is no longer strictly true. In order to retain this relationship and to avoid apparent changes in the chlorinity of sea water it has fortunately been possible to redefine chlorinity so that it is independent of changes in atomic weights.

The primary standard used in the determination of chlorinity is so-called "Normal Water" (*Eau de mer normale*), prepared by the Hydrographical Laboratories in Copenhagen, Denmark, and distributed to all oceanographic institutions. Some of these institutions made their own secondary standards by means of the Normal Water. As a result of world conditions the preparation of Normal Water has temporarily been taken over by the Woods Hole Oceanographic Institution.

Normal Water is sea water whose chlorinity has been adjusted to about 19.4 ‰ and accurately determined by either direct or indirect comparisons with the original standard prepared in 1902. Hence, the chlorinities of all batches have been independent of changes in the atomic weights. A new primary standard (Urnornormal-1937), prepared in 1937, will be used to establish the chlorinity of future batches of Normal Water for general distribution (Jacobsen and Knudsen, 1940). Comparison with earlier series of Normal Water showed the chlorinity of the new standard to be 19.381 ‰ . As a more absolute method should be available with which to check future Normal Water preparations, the 1937 primary standard was analyzed, using extremely pure "atomic weight silver." The equivalent amount of silver necessary to precipitate the halides was determined and the ratio of chlorinity to silver was found to be 0.3285233. On this basis a new definition of chlorinity was

introduced: *The number giving the chlorinity in grams per kilogram of a sea-water sample is identical with the number giving the mass in grams of "atomic weight silver" just necessary to precipitate the halogens in 0.3285233 kilogram of the sea-water sample.* By this redefinition the chlorinity has been made independent of changes in atomic weights, whereas the chlorine-equivalent, for which the original definition is retained, may vary slightly as the atomic weights are modified. The empirically established relationship between chlorinity and salinity (p. 51) remains valid and, according to the new definition of chlorinity, will never be affected by modification of the atomic weights.

The ratio of chlorine to silver, using the 1940 atomic weights ($\text{Ag} = 107.880$ and $\text{Cl} = 35.457$), is 0.3286707. Hence the ratio of chlorine-equivalent to the chlorinity is:

$$\frac{0.3286707}{0.3285233} = 1.00045.$$

This ratio is important when computing the chemical composition of sea water of given chlorinity, as the chlorine-equivalent will be greater than the indicated chlorinity. It must also be taken into account when preparing standard chloride solutions or when direct gravimetric analyses are made to determine the halide content of sea water. It is of interest to see how closely this ratio corresponds to the change which could be expected from the modification of the atomic weights. According to Jacobsen and Knudsen (1940) the indicated chlorinity of the Normal Water should be increased by 0.0094 ‰ to take this modification into account. This procedure gives a ratio between chlorine-equivalent and chlorinity of 1.000485, which is in good agreement with that given above.

When dealing with the chemistry of sea water, other substances are generally determined and reported on a volume basis, wherefore it is convenient to introduce *chlorosity* (Cl) (Intern. Assn. Phys. Oceanogr., 1939), which is the property corresponding to the chlorinity expressed as grams per 20°-liter (p. 169). Chlorosity is obtained by multiplying the chlorinity of a water sample by its density at 20°. Table 8 gives the corresponding values of chlorosity for chlorinities between 15.00 and 21.00 ‰.

METHODS FOR OBTAINING SALINITY, OTHER THAN TITRATION WITH SILVER NITRATE. The salinity can also be determined from the density of a water sample at a given temperature or by measuring either the electrical conductivity or the refractive index, both of which depend upon the salinity. The character of these properties will be dealt with below, but their application to salinity determinations will be briefly discussed here.

Determinations of *density* are appropriately included under methods for obtaining salinity, because such determinations are generally made at

atmospheric pressure and at room temperature and will thus differ from the density of the water sample at the locality where it was collected, but from the density thus observed the density at 0°C is computed, and from the latter the salinity can be found by means of Knudsen's Hydrographical Tables (p. 56). Determinations of density are rarely made, as it is difficult and time consuming to obtain an accuracy comparable to that obtained from the chlorinity titration, which is about ± 0.00001 .

TABLE 8
CORRESPONDING VALUES OF CHLORINITIES AND CHLOROSITIES

	15.00	16.00	17.00	18.00	19.00	20.00	21.00
Chlorinity, ‰.....	15.00	16.00	17.00	18.00	19.00	20.00	21.00
Chlorosity, g/L.....	15.28	16.32	17.37	18.41	19.46	20.51	21.57
Difference.....	.28	.32	.37	.41	.46	.51	.57

The methods used can be classified in two groups. In one, the mass of an accurately known volume of water at a definite temperature is determined, for example, by using a pycnometer bottle. In the second group, some form of hydrometer or float is used, and the density is computed from the weight of the hydrometer and the volume of the displaced water. The common form of stem hydrometer is generally not sufficiently accurate except when dealing with coastal waters, where great differences in density are found in short distances. The chain hydrometer of Hans Pettersson (1929) gives greater accuracy, but not as high as desirable. Nansen (1900) developed a hydrometer of total immersion which is very sensitive but which requires a water sample of at least 300 ml. The buoyancy is adjusted by the addition or removal of small weights until the hydrometer neither sinks nor rises in the sample, the temperature of which must be known within $\pm 0.02^\circ$. Another type of instrument makes use of a sinker suspended in the water sample from one arm of a delicate chemical balance (Cummings, 1932). The "weight" of the sinker in a sample at a known temperature is determined, and from this the density may be computed. In all instances various corrections must be applied and the original sources consulted before any such determinations are attempted. Thompson (1932) has described the methods in some detail and gives many references.

Measurements of the electrical conductivity have been employed on board the *Carnegie* and by the U. S. Coast Guard as the routine method for obtaining salinities. Owing to the relatively high concentration of the ions and the effect that temperature has upon the conductivity, the apparatus and the technique employed are rather complicated (Wenner, Smith, and Soule, 1930; Soule, 1932). These instruments are standardized empirically, using sea-water samples of *known salinity* (determined by silver nitrate titrations against Normal Water), and the values for the unknown samples are obtained by interpolation. In order to obtain

results of adequate accuracy, extreme care must be taken to control the temperature of the conductivity cells, and the resistance must be measured very accurately. A simplified diagram of the type of circuit used in electrical conductivity measurements on sea water is shown in fig. 12. *A* and *B* represent two similar electrolytic cells. *R* is a variable resistance in series with *B*, the cell used for the unknown sample. *A* contains sea water of known chlorinity or a standard potassium chloride solution of approximately the same conductance. *C* and *D* are fixed resistances and *S* is a slide-wire resistance. *H* is a source of alternating current of frequency between 600 to 1000 cycles per second at a potential of about 0.5 to 1.0 volt. *T* represents the telephone receiver used to establish the balance of the bridge. Two cells are used to eliminate small temperature effects. For details concerning the circuit and the various instruments the original references cited above should be consulted.

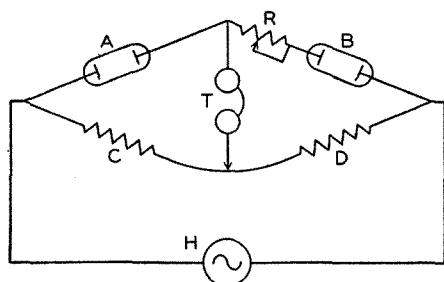


Fig. 12. Circuit used in measurements of the electrical conductivity as a means of obtaining the salinity of sea water. Symbols are explained in the text.

According to Thomas, Thompson, and Utterback (1934) the Grinnel Jones conductivity bridge may be used with Washburn pipette-type cells, the constants of which are determined with standard potassium chloride solutions.

The refractive index of sea water, which will be discussed later

(p. 70) in greater detail, varies only slightly within the ranges of temperature and salinity encountered in the sea, but, in an interferometer, differences in refractive index can be measured with extreme accuracy. At a given temperature such differences depend only upon the salinity, and special types of interferometers have therefore been developed for indirect determinations of salinity, using water of *known salinity* as a standard. Monochromatic light must be used because the refractive index varies with the wave length. This method of determining salinity has not been widely used.

Units of Temperature, Salinity, and Pressure, and Their Ranges in the Sea

In oceanography the *temperature* is measured in degrees centigrade. The thermometers used are described on p. 347. The accuracy of the measurements is about $\pm 0.02^{\circ}\text{C}$. *Salinity* is given as grams per kilogram of sea water; that is, in parts per thousand, or per mille, for which the symbol ‰ is used. An accuracy of $\pm 0.02 \text{‰}$ is required. *Pressure* (p. 170) is measured in atmospheres or in units of the c.g.s. system. An atmosphere is defined as the pressure exerted per square centimeter by a

column of mercury 760 mm high at a temperature of 0°C, where the acceleration of gravity is 980.665 cm/sec². In chemical oceanography a related unit, the Torr, is used which equals the pressure exerted per square centimeter by a column of mercury 1 mm high at a temperature of 0°C and at the above-mentioned acceleration of gravity. The c.g.s. unit of pressure is dyne/cm², and 1 atmosphere = 1.0133×10^6 dynes/cm². One million dynes/cm² was designated as 1 bar by V. Bjerknes. The corresponding practical unit used in physical oceanography is 1 decibar, which equals 0.1 bar. The pressure exerted per square centimeter by 1 m of sea water very nearly equals 1 decibar; that is, the hydrostatic pressure in the sea increases by 1 decibar for approximately every meter of depth. *Therefore, the depth in meters and the pressure in decibars are expressed by nearly the same numerical value.* This rule is sufficiently accurate for determining the effect of pressure on the physical properties of the water, but details of the pressure distribution must be computed from the density distribution (p. 408).

In the oceans the temperature ranges from about -2° to +30°C. The lower limit is determined by the formation of ice, and the upper limit is determined by processes of radiation and exchange of heat with the atmosphere (p. 127). In landlocked areas the surface temperature may be higher, but in the open ocean it rarely exceeds 30°C.

The salinity in the oceans is generally between 33 ‰ and 37 ‰. The surface salinity in high latitudes, in regions of high rainfall, or where there is dilution by rivers may be considerably less, and in certain semi-enclosed areas, such as the Gulf of Bothnia, the salinity may approach zero. In isolated seas in intermediate latitudes, such as the Red Sea, where evaporation is excessive, salinities may reach 40 ‰ or more. As the range in the open oceans is rather small, it is sometimes convenient to use a salinity of 35 ‰ as an average for all oceans. In the chapter on the chemistry of sea water the tabulations are made for water of 19.00 ‰ chlorinity; that is, 34.325 ‰ salinity.

In dealing with the pressure in the oceans, the atmospheric pressure is always neglected and the pressure at the sea surface is entered as zero. Since the pressure is essentially a function of depth and the numerical value in decibars nearly equals the depth in meters, the range in pressure will be from zero at the sea surface to over 10,000 decibars in the deepest part of the ocean.

Owing to the character of the distribution of temperature and salinity in the oceans, some relationships exist between these conditions and the pressure. The temperature of the deep and bottom water of the oceans is always low, varying between 4° and -1°C, so that high pressures are associated with low temperatures. Similarly, the salinity of deep and bottom water varies within narrow limits, 34.6 ‰ to 35 ‰, and high pressures are therefore associated with salinities between these limits.

Exceptions are found in isolated seas in intermediate latitudes, such as the Mediterranean and Red Seas, where water of high temperature and high salinity is found at great depths, and hence under great pressure.

Density of Sea Water

The density of any substance is defined as the mass per unit volume. Thus, in the c.g.s. system, density is stated in grams per cubic centimeter. The specific gravity is defined as the ratio of the density to that of distilled water at a given temperature and under atmospheric pressure. In the c.g.s. system the density of distilled water at 4°C is equal to unity. In oceanography, specific gravities are now always referred to distilled water at 4°C and are therefore numerically identical with densities. In oceanography the term density is generally used, although, strictly speaking, specific gravity is always considered.

The density of sea water depends upon three variables: temperature, salinity, and pressure. These are indicated by designating the density by the symbol $\rho_{s,\vartheta,p}$, but, when dealing with numerical values, space is saved by introducing $\sigma_{s,\vartheta,p}$, which is defined in the following manner:

$$\sigma_{s,\vartheta,p} = (\rho_{s,\vartheta,p} - 1)1000.$$

Thus, if $\rho_{s,\vartheta,p} = 1.02575$, $\sigma_{s,\vartheta,p} = 25.75$.

The density of a sea-water sample at the temperature and pressure at which it was collected, $\rho_{s,\vartheta,p}$, is called the density *in situ*, and is generally expressed as $\sigma_{s,\vartheta,p}$. At atmospheric pressure and temperature $\vartheta^{\circ}\text{C}$, the corresponding quantity is simply written σ_t , and at 0° it is written σ_0 . The symbol ϑ will be used for temperature except when writing σ_t , where, following common practice, t stands for temperature.

At atmospheric pressure and at temperature of 0°C the density is a function of the salinity only, or, as a simple relationship exists between salinity and chlorinity, the density can be considered a function of chlorinity. The International Commission, which determined the relation between salinity and chlorinity and developed the standard technique for determinations of chlorinity by titration, also determined the density of sea water at 0° with a high degree of accuracy, using pycnometers. From these determinations the following relation between σ_0 and chlorinity was derived:

$$\sigma_0 = -0.069 + 1.4708 \text{ Cl} - 0.001570 \text{ Cl}^2 + 0.0000398 \text{ Cl}^3.$$

Corresponding values of σ_0 , chlorinity, and salinity are given in Knudsen's Hydrographical Tables for each 0.01 ‰ Cl.

In order to find the density of sea water at other temperatures and pressures, the effects of thermal expansion and compressibility on the density must be known. The coefficient of thermal expansion has been determined in the laboratory under atmospheric pressure, and

according to these determinations the density under atmospheric pressure and at temperature ϑ° can be written in the form

$$\sigma_t = \sigma_0 - D.$$

The quantity D is expressed as a complicated function of σ_0 and temperature, and is tabulated in Knudsen's Hydrographical Tables. Since the values of σ_t are widely used in dynamical oceanography, tables for computing σ_t directly from temperature and salinity have been prepared by McEwen (1929) and Matthews (1932). A special slide rule for the same purpose has been devised by Sund (1929). Knudsen's tables also contain a tabulation of D as a function of σ_t and temperature, by means of which σ_0 can be found if σ_t is known ($\sigma_0 = \sigma_t + D$). This table is useful for obtaining the salinity of a water sample the density of which has been directly determined at some known temperature (p. 53).

The effect on the density of the compressibility of sea water of different salinities and at different temperatures and pressures was examined by Ekman (1908), who established a complicated empirical formula for the *mean* compressibility between pressures 0 and p decibars (quoted in V. Bjerknes and Sandström, 1910). From this formula, correction terms have been computed which, added to the value of σ_t , give the corresponding value $\sigma_{s,\vartheta,p}$ for any value of pressure.

COMPUTATION OF DENSITY AND SPECIFIC VOLUME IN SITU. Tables from which the density *in situ*, $\rho_{s,\vartheta,p}$, could be obtained directly from the temperature, salinity, and pressure with sufficiently close intervals in the three variables would fill many large volumes, but by means of various artifices convenient tables have been prepared. Following the procedure of Bjerknes and Sandström (1910), one can write

$$\rho_{s,\vartheta,p} = \rho_{35,0,0} + \epsilon_s + \epsilon_\vartheta + \epsilon_{s,\vartheta} + \epsilon_p + \epsilon_{s,p} + \epsilon_{\vartheta,p} + \epsilon_{s,\vartheta,p}.$$

The first four terms can be expressed by σ_t , which can readily be determined by the methods outlined above, and the remaining terms represent the effects of the compressibility. When dealing with density it is desirable, for reasons that will be explained later (p. 402), to introduce the dynamic depth, D , as the independent variable instead of the pressure, p , and to write

$$\rho_{s,\vartheta,D} = 1 + 10^{-3} \sigma_t + \epsilon_D + \epsilon_{s,D} + \epsilon_{\vartheta,D} + \epsilon_{s,\vartheta,D}.$$

The ϵ terms in this equation have been tabulated by Bjerknes and Sandström (1910) and by Hesselberg and Sverdrup (1914).

Instead of the density, $\rho_{s,\vartheta,p}$, its reciprocal value, the specific volume *in situ*, $\alpha_{s,\vartheta,p}$, is generally used in dynamic oceanography. In order to avoid writing a large number of decimals, the specific volume is commonly expressed as an anomaly, δ , defined in the following way:

$$\delta = \alpha_{s,\vartheta,p} - \alpha_{35,0,p},$$

where $\alpha_{35,0,p}$ is the specific volume of water of salinity 35 ‰, at 0°C, and at pressure p in decibars. The anomaly depends on the temperature, salinity, and pressure, and hence can be expressed as

$$\delta = \delta_s + \delta_\theta + \delta_{s,\theta} + \delta_{s,p} + \delta_{\theta,p} + \delta_{s,\theta,p}.$$

It should be observed that the anomaly, by definition, does not contain a term δ_p , which would represent the effect of pressure at temperature 0° and salinity 35 ‰. The reason for this is explained on page 409. Of the above terms the last one, $\delta_{s,\theta,p}$, is so small that it can always be neglected. Thus, five terms are needed for obtaining δ , and these were tabulated by Bjerknes and Sandström. If σ_t has already been computed, the terms that are independent of pressure can be combined as $\Delta_{s,\theta}$ (Sverdrup, 1933).

The value of $\Delta_{s,\theta} = \delta_s + \delta_\theta + \delta_{s,\theta}$ is easily obtained from σ_t because

$$\alpha_{s,\theta,0} = \frac{1}{\rho_{s,\theta,0}} = 1 - \frac{10^{-3}\sigma_t}{1 + 10^{-3}\sigma_t}$$

and

$$\alpha_{s,\theta,0} = \alpha_{35,0,0} + \Delta_{s,\theta} = 0.97264 + \Delta_{s,\theta}.$$

Hence

$$\Delta_{s,\theta} = 0.02736 - \frac{10^{-3}\sigma_t}{1 + 10^{-3}\sigma_t}.$$

Thus, in practice,

$$\delta = \Delta_{s,\theta} + \delta_{s,p} + \delta_{\theta,p}.$$

The values for these three terms are given in the appendix in small tables from which one can obtain the specific volume anomaly *in situ* of any water sample when its temperature, salinity, σ_t , and the pressure are known. In these tables the terms are entered with one extra decimal place in order to avoid any accumulation of errors due to rounding-off of figures, and also in order to facilitate preparation of exact graphs that may be used instead of the tabulation, or for the preparation of tables in which the arguments are entered at such close intervals that interpolation becomes easy or unnecessary.

The procedure that is followed in calculating the density or specific volume *in situ* can be summarized as follows. For a given water sample the temperature, salinity, and depth at which it was collected must be known. For reasons stated elsewhere it can be assumed that the numerical value of the pressure in decibars is the same as that of the depth in meters. From the temperature and salinity the value of σ_t is obtained from Knudsen's Tables or from graphs or tables prepared from this source (McEwen, 1929; Matthews, 1932). With the values of σ_t , temperature, salinity, and pressure the specific volume anomaly is computed by means of the tables given in the appendix. If the absolute value of the specific volume is required, the anomaly must be added to

the appropriate value of $\alpha_{35.0,p}$ given in the appendix. In this table are given the specific volume of water of 35 ‰ and 0° at various pressures in decibars. The absolute density *in situ* can then be obtained as the reciprocal of the specific volume.

Another set of tables for computing the specific volume *in situ* has been prepared by Matthews (1938), who, in our notations, defines the anomaly as $\delta' = \alpha_{s,\sigma,p} - \alpha_{34.85,0,p}$. Thus, he refers the anomalies to water of salinity 34.85 ‰, for which $\sigma_0 = 28.00$. The difference, $\delta - \delta' = \alpha_{34.85,0,p} - \alpha_{35.0,p}$, depends upon the pressure:

Pressure, decibars.....	0	2000	4000	6000	8000	10,000
$(\delta - \delta') \times 10^5$	12.0	11.5	11.1	10.7	10.4	10.1

Before comparing numerical values of the specific volume anomalies published in different reports, it is necessary to know on what tables the reported values are based.

USE OF KNUDSEN'S HYDROGRAPHICAL TABLES. A certain point concerning the use of Knudsen's Hydrographical Tables should be kept in mind. Although they have been shown to hold very well over the normal range of the concentration of sea water, they are not necessarily valid for highly diluted or concentrated sea water. The tables are based on the careful examination of a series of samples collected from various regions. The dilute samples used were taken in the Baltic Sea, where dilution sometimes reduces the chlorinity to about 1 ‰, and where the river water that is mainly responsible for the dilution contains relatively large quantities of dissolved solids. This is shown by the fact that the equation relating salinity to chlorinity shows a salinity of 0.03 ‰ for zero chlorinity, and according to Lyman and Fleming (1940) the total dissolved solids corresponding to this figure are probably of the order of 0.07 ‰. Thus, empirically, the salinity of sea water can be expressed by an equation of the type

$$S = a + b \text{ Cl},$$

where the numerical value of a depends upon the composition of the diluted samples used for establishing the relation. If 1 kg of water of high salinity, S , is diluted by adding n kilograms of *distilled* water, the salinity of the dilution will be $S_D = S/(n + 1)$, and the chlorinity of the diluted sample will be $\text{Cl}/(n + 1)$. According to Knudsen's Tables this sample, however, has a salinity $S_K = a + b \text{ Cl}/(n + 1)$. The difference between this and the true salinity is $S_K - S_D = a[n/(n + 1)]$, meaning that, if after dilution the chlorinity were determined by titration and the salinity were taken from Knudsen's Tables, it would be too high. Knudsen's Tables would therefore also give too great a density. As an example, let us assume that 1 kg of water of salinity 35 ‰ and chlorinity 19.375 ‰ is diluted by adding 9 kg of distilled water, reducing the

chlorinity to 1.938 ‰. Knudsen's Tables give for this chlorinity a salinity of 3.53 ‰, whereas the "true" salinity would be 3.50 ‰. Similarly, Knudsen's Tables would give a σ_0 equal to 2.78, whereas the true value should be 2.75. At low concentration, chlorinities computed from direct density determinations, and vice versa, may therefore be in error. For example, "chlorinities" of sea ice computed from density measurements made on the melt water were consistently smaller than those determined by titration (p. 219), and in this case the diluting water was essentially distilled water. The restricted application of the Cl:S:density relations to highly diluted water occurring naturally or prepared in the laboratory should always be kept in mind.

Thermal Properties of Sea Water

THERMAL EXPANSION. The coefficient of thermal expansion, e , defined by $e = (1/\alpha_{s,\sigma,p})(\partial\alpha_{s,\sigma,p}/\partial\theta)$, is obtained, at atmospheric pressure, from the terms for D in Knudsen's Hydrographical Tables, and at higher pressures from Ekman's tables or formulae (p. 57). The coef-

TABLE 9
COEFFICIENT OF THERMAL EXPANSION OF SEA WATER AT
DIFFERENT TEMPERATURES, SALINITIES, AND PRESSURES ($e \times 10^6$)

Pressure (decibars)	Salinity ‰	Temperature (°C)							
		-2	0	5	10	15	20	25	30
0.....	0	-105	-67	17	88	151	207	257	303
	10	-65	-30	46	112	170	222	270	315
	20	-27	4	75	135	189	237	282	324
	30	7	36	101	157	206	250	292	332
	35	23	51	114	167	214	257	297	334
2,000.....	35	80	105	157	202	241	278		
	40	94	118	168	210	248	283		
4,000.....	35	132	152	196	233	266			
	40	144	162	204	240	272			
6,000.....	34.85	177	194	230					
8,000.....	34.85		231	246					
10,000.....	34.85		276	287					

ficient for sea water is greater than that for pure water and increases with increasing pressure. A few numerical values are given in table 9, in which negative values indicate contraction with increasing temperature.

THERMAL CONDUCTIVITY. In water in which the temperature varies in space, heat is conducted from regions of higher to regions of lower

temperature. The amount of heat in gram calories per second which is conducted through a surface of area 1 cm² is proportional to the change in temperature per centimeter along a line normal to that surface, and the coefficient of proportionality, γ , is called the coefficient of thermal conductivity ($dQ/dt = -\gamma d\vartheta/dn$). For pure water at 15°C the coefficient is equal to 1.39×10^{-3} . The coefficient is somewhat smaller for sea water and increases with increasing temperature and pressure. This coefficient is valid, however, only if the water is at rest or in laminar motion (p. 89), but in the oceans the water is nearly always in a state of turbulent motion in which the processes of heat transfer are completely altered. In these circumstances the above coefficient of heat conductivity must be replaced by an "eddy" coefficient which is many times larger and which depends so much upon the state of motion that effects of temperature and pressure can be disregarded (p. 91).

SPECIFIC HEAT. The specific heat is the number of calories required to increase the temperature of 1 g of a substance 1°C. When studying liquids, the specific heat at constant pressure, c_p , is the property usually measured, but in certain problems the specific heat at constant volume, c_v , must be known.

The specific heat of sea water of different chlorinities was investigated by Thoulet and Chevallier, whose results have been recalculated and presented in different ways. Krümmel (1907) gives the following values for the specific heat at 17.5°C and atmospheric pressure:

S (%.....	0	5	10	15	20	25	30	35	40
c_p (cals/g)....	1.000	0.982	0.968	0.958	0.951	0.945	0.939	0.932	0.926

Kuwahara (1939) gives an empirical equation for the specific heat at 0°C and atmospheric pressure:

$$c_p = 1.005 - 0.004136 S + 0.0001098 S^2 - 0.000001324 S^3.$$

It will be noted that the specific heat decreases with increasing salinity, but it has been pointed out by Krümmel that the effect is somewhat larger than might be expected from the composition of the solution, and the problem merits reinvestigation. The effects of temperature and pressure have not been measured, but it has been assumed that they are the same as those for pure water. Ekman (1914) gives the following values showing the effect of temperature on the specific heat of water of $S = 34.85 \%$ at atmospheric pressure:

ϑ (°C).....	-2	0	5	10	15	20
c_p	0.942	0.941	0.938	0.935	0.933	0.932

The effect of pressure on the specific heat has been computed by Ekman (1914) from the equation

$$\frac{dc_p}{dp} = -10^5 \frac{T}{\rho J} \left(\frac{de}{d\vartheta} + e^2 \right),$$

where p is the pressure in decibars, T is the absolute temperature, ρ the density, J the mechanical equivalent of heat, and e the coefficient of thermal expansion. The combined effect of temperature and pressure is shown in table 10, where $c_{p,0}$ is the specific heat at atmospheric pressure

TABLE 10
DIFFERENCE BETWEEN SPECIFIC HEAT AT ATMOSPHERIC PRESSURE
AND AT PRESSURE p , ($c_{p,0} - c_{p,p}$)
(Salinity 34.85 % at indicated temperature)

ϑ (°C)	p (decibars)				
	2000	4000	6000	8000	10,000
-2.....	0.0171	0.0315	0.0435		
0.....	0.0159	0.0291	0.0401	0.0492	0.0566
5.....	0.0136	0.0248	0.0340	0.0416	0.0479
10.....	0.0120	0.0220			
15.....	0.0110	0.0203			
20.....	0.0105				

of water of salinity 34.85 %. The specific heat at constant volume, which is somewhat less than c_p , may be computed from the following equation:

$$c_v = c_p - \frac{Te^2}{\rho K J},$$

where K is the true compressibility (p. 68) and where the other symbols are defined above. According to Matthews (1923), the ratio of $c_p:c_v$ for water of S = 34.85 % at atmospheric pressure increases from 1.0004 at 0° to 1.0207 at 30°. The effect of pressure is appreciable; for the same water at 0°, the ratio is 1.0009 at 1000 decibars, and 1.0126 at 10,000 decibars. This ratio is important in the study of the velocity of sound (p. 76).

LATENT HEAT OF EVAPORATION. The latent heat of evaporation of pure water is defined as the amount of heat in gram calories needed for evaporating 1 g of water, or as the amount of heat needed for producing 1 g of water vapor of the same temperature as the water. Only in the latter form is the definition applicable to sea water. The latent heat of evaporation of sea water has not been examined, but it is generally assumed that the difference between that and pure water is insignificant; therefore, between temperatures of 0° and 30°C, the formula

$$L = 596 - 0.52\vartheta$$

can be used.

ADIABATIC TEMPERATURE CHANGES. When a fluid is compressed, without gain or loss of heat to the surroundings, work is performed on the system and there is a rise in temperature. Conversely, if expansion takes place, the liquid itself gives up energy, which is reflected in a drop in temperature. Such *adiabatic* temperature changes are well known and important in the atmosphere. Sea water is compressible, and the effects of adiabatic processes, although small, must be taken into account when studying the vertical distribution of temperature in the great depths of the oceans and in deep isolated basins where the adiabatic heating may

TABLE 11
ADIABATIC TEMPERATURE GRADIENT IN THE SEA, IN °C PER 1000 M
AT SALINITY 34.85 ‰

Depth (m)	Temperature (°C)								
	-2	0	2	4	6	8	10	15	20
0	0.016	0.035	0.053	0.078	0.087	0.103	0.118	0.155	0.190
1,000	0.036	0.054	0.071	0.087	0.103	0.118	0.132	0.166	0.199
2,000	0.056	0.073	0.089	0.104	0.118	0.132	0.146	0.177	0.207
3,000	0.075	0.091	0.106	0.120	0.133	0.146	0.159	0.188	
4,000	0.093	0.108	0.122	0.135	0.147	0.159	0.170	0.197	
5,000	0.110	0.124	0.137	0.149					
6,000	0.120	0.140	0.152	0.163					
7,000	0.155	0.165	0.175					
8,000	0.169	0.178	0.187					
9,000	0.182	0.191	0.198					
10,000	0.194	0.202	0.209					

lead to a temperature increase toward the bottom (for example, p. 739). Adiabatic cooling is of immediate practical concern when water samples are taken with thermally insulated water bottles and the temperature of the water sample is determined after it has been raised to the surface (p. 355).

Any adiabatic effect is related to changes in pressure, but in the sea the pressure can be considered proportional to the depth, and adiabatic temperature changes can be given as changes per unit of depth instead of per unit of pressure. According to Lord Kelvin, the change in temperature for each centimeter of vertical displacement is

$$10^{-5}d\vartheta = \frac{T_e}{J c_p} g \rho,$$

where T is the absolute temperature and g is the acceleration of gravity, and where the other symbols have their previous meanings. This change

is extremely small, and for practical purposes the adiabatic temperature change on a vertical distance of 1000 m, called the *adiabatic temperature gradient*, is used instead. It should be observed that the adiabatic temperature gradient depends mainly upon the coefficient of thermal expansion, ϵ , which varies much more with temperature and pressure than the other quantities involved. Ekman (1914) has computed the adiabatic temperature gradient for different temperatures, salinities, and depths, and some of his values are shown in table 11.

The temperature that a water sample would attain if raised adiabatically to the sea surface has been called the *potential temperature* (Helland-Hansen, 1912b) and has been designated Θ . Thus, $\Theta = \vartheta_m - \Delta\vartheta$, where ϑ_m is the temperature *in situ* and $\Delta\vartheta$ is the amount by which the temperature would decrease adiabatically if the sample were raised to the surface. The potential temperature can be obtained from a table of the adiabatic gradients by step-wise computations. Such computations are long and tedious, but Helland-Hansen (1930) has prepared a convenient table for $\Delta\vartheta$ which has been reproduced in table 12. The table

TABLE 12
ADIABATIC COOLING (IN 0.01°C) WHEN SEA WATER ($\sigma_0 = 28.0$,
 $S = 34.85\text{‰}$) WHICH HAS A TEMPERATURE OF ϑ_m AT THE
DEPTH OF m METERS IS RAISED FROM THAT DEPTH
TO THE SURFACE

Depth (m)	Temperature, ϑ_m (°C)												
	-2	-1	0	1	2	3	4	5	6	7	8	9	10
1,000.....	2.6	3.5	4.4	5.3	6.2	7.0	7.8	8.6	9.5	10.2	11.0	11.7	12.4
2,000.....	7.2	8.9	10.7	12.4	14.1	15.7	17.2	18.8	20.4	21.9	23.3	24.8	26.2
3,000.....	13.6	16.1	18.7	21.2	23.6	25.9	28.2	30.5	32.7	34.9	37.1	39.2	41.2
4,000.....	21.7	25.0	28.4	31.6	34.7	37.7	40.6	43.5	46.3	49.1	51.9	54.6	57.2
5,000.....	31.5	35.5	39.6	43.4	47.2	50.9	54.4						
6,000.....	42.8	47.5	52.2	56.7	61.1	65.3	69.4						
7,000.....			66.2	71.3	76.2	80.9	85.5						
8,000.....			81.5	87.1	92.5	97.7	102.7						
9,000.....			98.1	104.1	109.9	115.6	121.0						
10,000.....			115.7	122.1	128.3	134.4	140.2						

is based on a salinity of 34.85‰ ($\sigma_0 = 28.0$) and is applicable in general to the deep ocean areas, because in these the salinity does not differ much from 34.85‰ and because the effect of salinity on the adiabatic processes is small. It may be seen that if water of 2°C is raised adiabatically from 8000 m to the surface, $\Delta\vartheta = 0.925^\circ$, and therefore the potential temperature of that water is 1.075° . The adiabatic cooling of water of different

salinities that may occur near the surface and in the Mediterranean Sea is given in tables 13 and 14.

Colligative and Other Properties of Sea Water

COLLIGATIVE PROPERTIES. The colligative properties—namely, vapor-pressure lowering, freezing-point depression, boiling-point elevation, and osmotic pressure—are unique properties of solutions. If the

TABLE 13

ADIABATIC COOLING (IN 0.01°C) WHEN WATER OF THE INDICATED TEMPERATURE AND SALINITY IS RAISED FROM 1000 M TO THE SURFACE

Salinity (‰)	Temperature, ϑ_m (°C)											
	0	2	4	6	8	10	12	14	16	18	20	22
30.0.....	3.5	5.3	7.0	8.7	10.3	11.8	13.2	14.7	16.1	17.6	18.9	20.3
32.0.....	3.9	5.7	7.3	9.0	10.6	12.1	13.5	15.0	16.4	17.8	19.1	20.5
34.0.....	4.3	6.0	7.7	9.4	10.9	12.4	13.8	15.3	16.6	18.0	19.3	20.7
36.0.....	4.7	6.4	8.1	9.7	11.2	12.7	14.1	15.5	16.9	18.3	19.6	20.9
38.0.....	5.1	6.8	8.4	10.0	11.6	13.0	14.4	15.8	17.2	18.5	19.8	21.1

TABLE 14

ADIABATIC COOLING (IN 0.01°C) WHEN SEA WATER ($\sigma_0 = 31.0$, $S = 38.57\text{ ‰}$) WHICH HAS A TEMPERATURE OF ϑ_m AT THE DEPTH OF m METERS IS RAISED FROM THAT DEPTH TO THE SURFACE (APPLICABLE TO THE MEDITERRANEAN SEA)

Depth (m)	Temperature, ϑ_m (°C)		
	12	13	14
1000.....	14.4	15.1	15.8
2000.....	30.0	31.4	32.7
3000.....	46.6	48.6	50.6
4000.....	64.2	66.7	69.2

magnitude of any one of them is known for a solution under a given set of conditions, the others may readily be computed. Solutions of the complexity and concentration of sea water do not obey the generalized theories of the colligative properties, but in all cases the departures from the theoretical values are proportional.

Only the depression of the freezing point for sea water of different chlorinities has been determined experimentally (Knudsen, 1903; Miyake,

1939a), and empirical equations for computing the vapor-pressure lowering and osmotic pressure have been based on these observations. Thompson (1932) has shown that the depressions of the freezing point, $\Delta\vartheta_f$, may be calculated from the chlorinity by means of the equation

$$\Delta\vartheta_f = -0.0966 \text{ Cl} - 0.0000052 \text{ Cl}^3.$$

The values of $\Delta\vartheta_f$ for various chlorinities are shown in fig. 13. The freezing point of sea water is the "initial" freezing point; namely, the

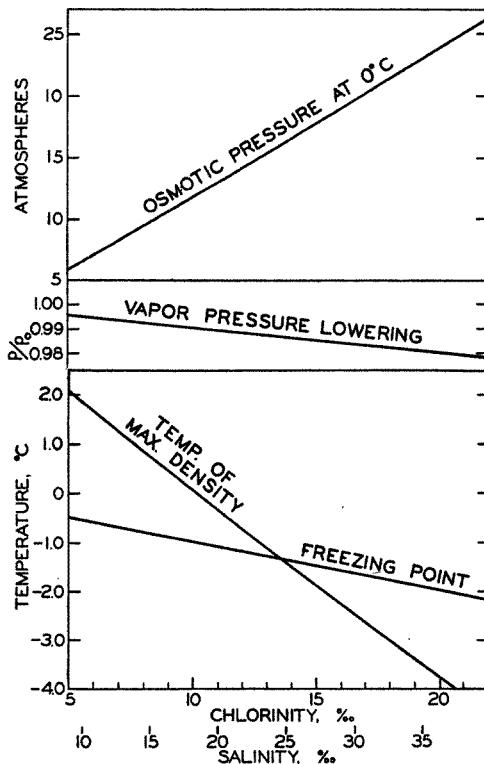


Fig. 13. Osmotic pressure, vapor pressure relative to that of pure water, freezing point, and temperature of maximum density as functions of chlorinity and salinity.

temperature at which an infinitely small amount of ice is in equilibrium with the solution. As soon as any ice has formed, the concentration of the dissolved solids increases, and hence the formation of additional ice can take place only at a lower temperature (p. 216).

The vapor pressure of sea water of any chlorinity referred to distilled water at the same temperature can be computed from the following equation (Witting, 1908):

$$\frac{e}{e_0} = 1 - 0.000969 \text{ Cl},$$

where e is the vapor pressure of the sample and e_0 that of distilled water at the same temperature (fig. 13). Sea water within the normal range of concentration has a vapor pressure about 98 per cent of that of pure water at the same temperature, and in most cases it is not necessary to consider the effect of salinity, since variations in the temperature of the surface waters have a much greater effect upon the vapor pressure (table 29, p. 116).

The osmotic pressure can be calculated from the freezing-point depression by means of the equation derived by Stenius (Thompson, 1932):

$$OP_0 = -12.08 \Delta\vartheta_f.$$

The osmotic pressure at any temperature may then be computed:

$$OP_s = OP_0 \times \frac{273 + \vartheta}{273}.$$

The variations in osmotic pressure over the range in chlorinity from 5 ‰ to 22 ‰ are shown in fig. 13.

It will be noted that the freezing-point depression and, therefore, the other colligative properties are not linear functions of the chlorinity, because chlorinity is reported as grams per kilogram of sea water and not as grams per kilogram of solvent water, in which case a linear relationship should be expected. In agreement with this expectation, Lyman and Fleming (1940) found that the freezing-point depression could be written in the form

$$\Delta\vartheta_f = -0.05241 Z,$$

where Z is the total salt content per kilogram of solvent water.

The magnitude of the colligative properties depends upon the concentration of ions in the solution and upon their activity. According to present concepts the major constituents of sea water exist as ions whose concentrations may be computed from data in table 35 (p. 173). Within the normal range of sea water the gram-ionic concentration per kilogram of solvent water may be obtained from the following expression:

$$I = 0.03183 Z.$$

The gram-ionic concentration for water of 19 ‰ Cl is 1.1368. The gram-molecular lowering of the freezing point is -1.86° . Therefore, the "theoretical" value for the depression of the freezing point of water of chlorinity 19 ‰ should be $-1.86 \times 1.1368 = -2.11^\circ$, but the observed value for water of the same chlorinity is -1.872° . The ratio between the actual and the theoretical value, 0.89, is a measure of the reduced

activity of the ions in sea water of normal concentrations, and, as mentioned above, the other colligative properties bear the same relation to the theoretical values.

MAXIMUM DENSITY. Pure water has its maximum density at a temperature of very nearly 4° , but for sea water the temperature of maximum density decreases with increasing salinity, and at salinities greater than $24.70\%_{\text{oo}}$ is below the freezing point. At a salinity of $24.70\%_{\text{oo}}$, the temperature of maximum density coincides with the freezing point: $\vartheta_f = -1.332^{\circ}$. Consequently, the density of sea water of salinity greater than $24.70\%_{\text{oo}}$ increases continuously when such water

TABLE 15
MEAN COMPRESSIBILITY OF SEA WATER OF SALINITY
 $34.85\%_{\text{oo}}$, ($k \times 10^8$)

p (bars)	Temperature ($^{\circ}\text{C}$)						
	0	5	10	15	20	25	30
0	4659	4531	4427	4345	4281	4233	4197
100	4582	4458	4357	4278			
200	4508	4388	4291				
400	4368	4256					
1000	4009	3916					

is cooled to its freezing point. The temperature of maximum density is shown in fig. 13 as a function of salinity and chlorinity.

COMPRESSIBILITY. Ekman (1908) has derived an empirical equation for the *mean* compressibility of sea water between pressures 0 and p bars (p. 57), as defined by $\alpha_{s,s,p} = \alpha_{s,s,0}(1 - kp)$. Numerical values are given in table 15, where the *bar* has been used as pressure unit.

The *true* compressibility of sea water is described by means of a coefficient that represents the proportional change in specific volume if the hydrostatic pressure is increased by one unit of pressure: $K = (-1/\alpha)(d\alpha/dp)$. The true compressibility can be calculated from the mean compressibility, which was tabulated by Ekman, using the equation

$$K = \frac{\left(k + p \frac{dk}{dp} \right)}{(1 - kp)},$$

where k is the mean compressibility referred to *bar* as pressure unit, and p is the pressure in bars.

VISCOUSITY. When the velocity of moving water varies in space, frictional stresses are present. The frictional stress which is exerted on a

surface of area 1 cm² is proportional to the change of velocity per centimeter along a line normal to that surface ($\tau_s = \mu dv/dn$). The coefficient of proportionality (μ) is called the *dynamic viscosity*. This coefficient decreases rapidly with increasing temperature and increases slowly with increasing salinity (table 16, after Dorsey, 1940). With increasing pressure the coefficient for pure water decreases at low temperature but increases at high temperature (Dorsey, 1940). If the same holds true for sea water, and if the effect is of similar magnitude, the viscosity of water of salinity 35 ‰ and temperature 0° is 18.3×10^{-3} c.g.s. units at a pressure of 10,000 decibars, as against 18.9×10^{-3} at atmospheric pressure. The difference is insignificant, and the effect of pressure on the viscosity can be disregarded in the oceans.

TABLE 16
VISCOSITY OF PURE WATER AND OF SEA WATER AT ATMOSPHERIC PRESSURE ($\mu \times 10^3$ C.G.S. UNITS)
(After Dorsey)

Salinity (‰)	Temperature (°C)						
	0	5	10	15	20	25	30
0.....	17.9	15.2	13.1	11.4	10.1	8.9	8.0
10.....	18.2	15.5	13.4	11.7	10.3	9.1	8.2
20.....	18.5	15.8	13.6	11.9	10.5	9.3	8.4
30.....	18.8	16.0	13.8	12.1	10.7	9.5	8.6
35.....	18.9	16.1	13.9	12.2	10.9	9.6	8.7

The viscosity that has been discussed so far is valid only if the motion is laminar, but, as stated above, turbulent motion prevails in the sea, and an "eddy" coefficient must be introduced which is many times larger (p. 91).

DIFFUSION. In a solution in which the concentration of a dissolved substance varies in space, the amount of that substance which per second diffuses through a surface of area 1 cm² is proportional to the change in concentration per centimeter along a line normal to that surface ($dM/dt = -\delta dc/dn$). The coefficient of proportionality (δ) is called the *coefficient of diffusion*; for water it is equal to about 2×10^{-5} , depending upon the character of the solute, and is nearly independent of temperature. Within the range of concentrations encountered in the sea the coefficient is also nearly independent of the salinity.

What was stated about the coefficient of thermal conductivity in the sea applies also to the coefficient of diffusion. Where turbulent motion prevails, it is necessary to introduce an "eddy" coefficient that is many times larger and that is mainly dependent on the state of motion.

SURFACE TENSION. The surface tension of sea water is slightly greater than that of pure water at the same temperature. Krümmel (1907) carried out experimental observations from which he derived an empirical equation relating the surface tension to the temperature and salt content. This equation was revised by Fleming and Revelle (1939) to take into account the more recent values for pure water. The revised expression has the form

$$\text{Surface tension (dyne/cm}^2\text{)} = 75.64 - 0.144\vartheta + 0.0399 \text{ Cl.}$$

The surface tension is decreased by impurities, and in the sea is mostly smaller than stated.

REFRACTIVE INDEX. The refractive index increases with increasing salinity and decreasing temperature. The problem of determining the relationship between these variables, and the types of equipment to be used, has been discussed by a number of authors (for example, Utterback, Thompson, and Thomas, 1934; Bein, Hirsekorn, and Möller, 1935; Miyake, 1939). Since the index varies with the wave length of light, a standard must be selected, usually the *D* line of sodium.

Utterback, Thompson, and Thomas determined at a number of temperatures the refractive index of ocean-water samples that had been diluted with distilled water. They found that the refractive index could be represented by expressions of the following type:

$$n_{\vartheta} = n_{0,\vartheta} + k_{\vartheta} \text{ Cl},$$

where n_{ϑ} is the refractive index of the sea-water sample at the temperature ϑ° , $n_{0,\vartheta}$ is that of distilled water at the same temperature, and k_{ϑ} is a constant appropriate for that temperature. This equation gives a straight-line relationship between the refractive index and the chlorinity, but it should be remembered that it is valid for ocean water diluted with distilled water and that at low chlorinities the diluted water does not correspond to sea water of the same low chlorinity, according to Knudsen's Hydrographical Tables. In fig. 14 the relationships between n , ϑ , and Cl determined by Utterback, Thompson, and Thomas are shown. Miyake (1939b) determined the refractive index for the sodium *D* line at 25° ($n_{D,25^{\circ}}$) for oceanic water samples that were diluted in the laboratory. He represented his results by the same type of equation, but obtained numerical constants that differ slightly from those of the authors mentioned above.

$$n_{D,25^{\circ}} = 1.332497 + 0.000334 \text{ Cl} \quad (\text{Miyake}),$$

$$n_{D,25^{\circ}} = 1.33250 + 0.000328 \text{ Cl} \quad (\text{U., T., and T.}).$$

Miyake found that the refractive index of sea water could be expressed as

$$n = n_0 + \Sigma(v - n_0),$$

where n_0 is the refractive index of distilled water and v is the refractive index of solutions of single salts having concentrations comparable to those in which these salts occur in the sea water. It is known that individual ions have characteristic ionic refractions. In sea water the salts are completely ionized, and, as the molar refractions are known for each ion, Miyake was able to compute the refractive index with a fair degree of accuracy.

ELECTRICAL CONDUCTIVITY. Thomas, Thompson, and Utterback (1934), and Bein, Hirsekorn, and Möller (1935) have studied conductivity as a function of chlorinity and temperature and have given tables for the specific conductance in reciprocal ohms per cubic centimeter for a wide range in conditions.

The results of the investigations of Thomas, Thompson, and Utterback are expressed at temperatures of 0, 5, 10, 15, 20, and 25°. Their results are shown graphically in fig. 15. The values for the low chlorinities were obtained by diluting ocean water with distilled water, and hence the density and other properties will not correspond exactly to those of water of the same chlorinity, as represented in Knudsen's Hydrographical Tables.

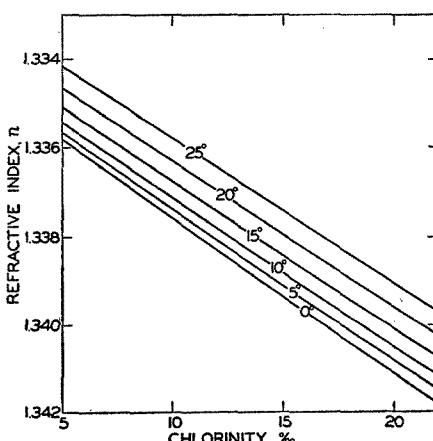


Fig. 14. Refractive index of sea water as a function of temperature and chlorinity.

Properties of Sea Ice

The processes of freezing and the chemical properties of ice formed from sea water in high latitudes are discussed elsewhere. The physical properties of sea ice, like those of the water, depend upon the salt content, which in turn is a function of the rate of freezing, age, thermal history, and so forth. The salts in sea ice (p. 217) do not differ greatly in composition from those in the water, as they are generally present in brine that is enclosed in small cells. Therefore, within practical limits, the terms chlorinity and salinity of sea ice have the same meaning as for the water, although the salts are not uniformly distributed in the ice.

The freezing point of sea water, as was pointed out previously, represents the initial freezing point at which ice is in equilibrium with sea water of the indicated chlorinity. If the ice and sea water are in a closed system, as when brine is enclosed in cells in the ice, a further reduction of temperature is necessary to cause additional ice to separate.

Observations by Ringer (p. 217) make it possible to find the salt content of the brine that will be in equilibrium with ice at different temperatures (fig. 16A). The salt content used in this case is the total amount present, although at low temperatures certain salts crystallize out. The reason

for giving the data in this way is that analyses for the chlorinity or salinity of the ice do not discriminate between that existing in solution or as crystals. Given the salt content of the brine that will be in equilibrium with the ice at any temperature, it is possible to calculate the amount of enclosed brine per kilogram of ice of unit salinity at any temperature. The amount of brine in any sample of ice at a given temperature can be obtained by multiplying the value at that temperature, shown by the curve in fig. 16B, by the salinity of the

Fig. 15. Specific conductance, reciprocal ohms/cm³, of sea water as a function of temperature and chlorinity.

ice. Thus, sea ice of salinity 10 %/oo at -3°C is essentially a mush containing 200 g of brine per kilogram. From the data in fig. 16 it is also possible to compute the amount of ice that is formed or melted when ice of a known salinity is cooled or heated.

Another variable which markedly affects certain physical properties of sea ice is the gas content. The gases normally occur as small "bubbles" in the ice, and the quantity is generally large in ice that has frozen rapidly, in which case bubbles represent gases originally in solution in the water, or in old ice that has undergone partial thawing and been refrozen, in which case atmospheric air is trapped in the ice.

In the following discussion the numerical values relating to the properties of sea ice are quoted from the work of Malmgren (1927), unless otherwise noted. The corresponding values for pure ice represent physical constants for gas- and salt-free ice, and are taken from Barnes (1928).

Pure ice at 0° has a density of 0.91676, whereas pure water at the same temperature has a density of 0.9998674. The density of sea ice

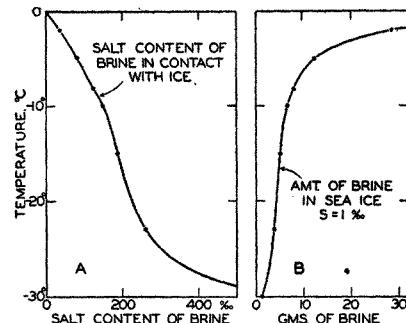
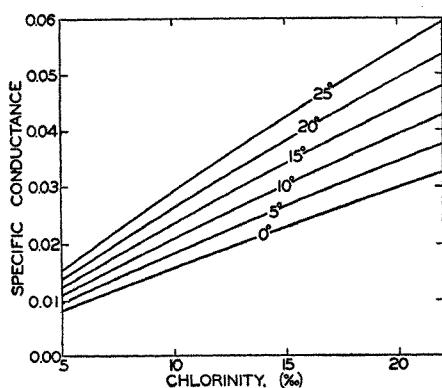


Fig. 16. (A) Salt content of the enclosed brine in equilibrium with sea ice at different temperatures. (B) Amount of brine in 1 kg of sea ice of salinity 1 %/oo at different temperatures.

varies both above and below that of pure ice, depending upon its content of water or brine and the gas content. Malmgren reports a range between 0.857, in old surface ice, and 0.92.

The specific heat of pure ice depends upon its temperature and varies within narrow limits, but that of sea ice is a much more variable property, depending upon the salt or brine content and the temperature. Changing the temperature of sea ice will generally involve either melting or freezing, and the amount of heat required will depend upon the salinity of the ice, as shown in table 17. It should be noted that the specific heat of pure ice is less than half that of pure water. Near the initial

TABLE 17
SPECIFIC HEAT OF SEA ICE
(From Malmgren)

Salinity ‰	Temperature (°C)										
	-2°	-4°	-6°	-8°	-10°	-12°	-14°	-16°	-18°	-20°	-22°
0.....	0.48	0.48	0.48	0.48	0.48	0.47	0.47	0.47	0.47	0.47	0.46
2.....	2.47	1.00	0.73	0.63	0.57	0.55	0.54	0.53	0.53	0.52	0.52
4.....	4.63	1.50	0.96	0.76	0.64	0.59	0.57	0.57	0.56	0.55	0.54
6.....	6.70	1.99	1.20	0.88	0.71	0.64	0.61	0.60	0.58	0.57	0.56
8.....	8.76	2.49	1.43	1.01	0.78	0.68	0.64	0.64	0.61	0.60	0.58
10.....	10.83	2.99	1.66	1.14	0.85	0.73	0.68	0.67	0.64	0.62	0.60
15.....	16.01	4.24	2.24	1.46	1.02	0.85	0.77	0.76	0.71	0.68	0.65

freezing point, the extremely high specific heat of ice of high salinity is, of course, due to the formation of ice from the enclosed brine or its melting.

The latent heat of fusion of pure ice at 0°C and at atmospheric pressure is 79.67 calories per gram. As the melting of sea ice does not occur at a fixed temperature on account of the presence of the salts, it is impossible to designate the latent heat in the usual way. Malmgren gives the heat required to melt 1 g of ice of given salinity that was initially at the indicated temperature (table 18). It may readily be seen that the presence of salts reduces the apparent latent heat.

The vapor pressure of sea ice has not been investigated, but it cannot depart very much from that of pure ice, which has the following values:

$$\begin{array}{ll} 0^\circ, e = 6.11 \text{ mb} & -20^\circ, e = 1.04 \text{ mb} \\ -10^\circ, e = 2.61 \text{ mb} & -30^\circ, e = 0.39 \text{ mb} \end{array}$$

The latent heat of evaporation of pure ice is variable. It has been found that under certain conditions the ice can volatilize directly to vapor without going through the liquid stage, in which case the latent heat of evaporation is about 600 calories per gram. If the evaporation

proceeds more slowly, the ice melts before vaporizing and 700 calories per gram are required. The latter process seems to prevail in nature.

TABLE 18
LATENT HEAT OF MELTING OF SEA ICE
(From Malmgren)

Temperature (°C)	Salinity (‰)							
	0	2	4	6	8	10	15	
-1.....	80	72	63	55	46	37	16	
-2.....	81	77	72	68	63	59	48	

For pure ice the coefficient of thermal expansion (e) per one degree is about 1.7×10^{-4} , where $e = (1/\alpha)(d\alpha/d\vartheta)$. The coefficient is independent of temperature. The thermal expansion of sea ice is a function of its temperature and salinity and shows a considerable range over both positive and negative values, as shown in table 19, where negative values indicate expansion on cooling, positive values contraction on cooling. This anomalous behavior is again related to the salt or brine content because any change in temperature leads to the formation or melting of a certain amount of ice. Thus, the processes in sea ice are a combination of the sudden change in volume associated with the ice \rightleftharpoons water transformation and the thermal expansion of the brine and the ice. According to table 19, sea ice of high salinity expands rapidly as it is cooled below the initial freezing point. The coefficient decreases at lower temperatures but is always negative. On the other hand, ice of low salinity first expands and then contracts as its temperature is lowered.

The coefficient of thermal conductivity of pure ice is about 0.005, which is approximately three times as great as that of pure water at 0°. Malmgren carried out a number of measurements in the Arctic ice fields and found that the conductivity was greatly affected by the character of the ice, particularly by the gas content (that is, the porosity) of the ice. Under natural conditions, porosity is greater near the surface than in the deeper portions of the ice. On the average, the coefficient of thermal conductivity near the surface was about 1.5×10^{-3} , at 0.5 m it was 4.0×10^{-3} , and below 1.0 m it approached the value of pure ice given above—namely, 5.0×10^{-3} .

Transmission of Sound

Water is a very efficient medium for the transmission of sound, which travels more rapidly and with much less absorption of energy through

TABLE 19
COEFFICIENT OF THERMAL EXPANSION PER 1°C FOR SEA ICE ($e \times 10^4$)
(From Malmgren)

Salinity (‰)	Temperature (°C)						
	-2	-4	-6	-8	-10	-12	-14
2.....	22.10	-4.12	-1.06	0.16	0.83	1.13	1.23
4.....	45.89	-9.92	-3.81	-1.37	-0.92	0.57	0.78
6.....	69.67	-15.73	-6.55	-2.90	-0.88	0.00	0.33
8.....	93.46	-21.53	-9.30	-4.43	-1.73	-0.57	-0.13
10.....	117.25	-27.34	-12.05	-5.95	-2.59	-1.13	-0.59
15.....	176.72	-41.85	-18.92	-9.78	-4.73	-2.54	-1.72

water than through air. This characteristic has made possible the development of submarine acoustic methods that are of tremendous value in navigation. The most familiar use is in echo sounding, where the time required for an impulse to travel to the sea floor and back to the vessel is used as a measure of the depth. Horizontal sound transmission is used in radio-acoustic range finding, which is employed by surveying vessels in order to determine their location accurately when carrying on sounding operations out of sight of land. The position of the vessel is determined by measuring the time intervals required for the impulse of a bomb explosion to travel to two or more submarine sound detectors (hydrophones) at known locations, usually near shore. Sound transmission from subsurface bells has also been used as a navigational aid near lighthouses. Ultrasonic impulses (frequency above the range audible to the human ear) are largely directional, and many attempts have been made to develop instruments for locating icebergs and other navigational hazards in the path of a vessel. The practical aspects of this problem and the types of equipment used are too numerous to consider in this discussion, and reference is made to the Hydrographic Review (Monaco) as an excellent source of material on these problems.

The velocity of sound in sea water is independent of the wave length except for impulses resulting from the detonation of relatively large amounts of explosives. Initially the impulse from such explosions may travel about 30 per cent faster than normal, apparently because of the tremendous energy involved.

The velocity of sound in a liquid may be computed from the elasticity and density:

$$v = \sqrt{\frac{\text{elasticity}}{\text{density}}},$$

but in practice it is more convenient to use the expression

$$v = \sqrt{\frac{\gamma}{\rho K}},$$

where γ is the ratio of the specific heats, c_p/c_v (p. 62), ρ is the density, and K is the true compressibility. If ρ and K are in c.g.s. units, the velocity is in centimeters per second. The ratio γ is introduced because the sound impulse is a wave of compression and, hence, heats the water it passes through. The three variables, γ , ρ , and K , all change with temperature, salinity, and pressure, and therefore must be evaluated for any given set of conditions. For example, water of salinity 34.85 $^{\circ}/_{\text{o}}$ at 30° has a density of 1.021637 at atmospheric pressure ($p = 0$), and for these conditions $\gamma = 1.0207$ and $K = 4.196 \times 10^{-11}$. Hence,

$$v = 1.543 \times 10^5 \text{ cm/sec} = 1543 \text{ m/sec.}$$

By means of the above formula, suitable tables have been prepared that give the velocity of sound as a function of temperature, salinity, and pressure. The first practical tables were those of Heck and Service (1924) of the U. S. Coast and Geodetic Survey, but these have been superseded by the British Admiralty Tables prepared by Matthews (1927), which, although based on the same original data, are slightly more consistent. The variations in velocity as a function of temperature and salinity are shown in fig. 17. The effect of pressure, and hence depth, is shown in fig. 18. This effect is almost independent of the temperature and salinity, but the curve shown in the figure is actually for salinity 34.85 ‰ and 0°C. This salinity is the standard reference salinity used in these tables, and corrections due to salinity variations are given in the form of anomalies to be added to or subtracted from the standard values.

If the velocity of the sound is known, it is possible to determine the

wavelength of sound of different frequencies from the equation $\lambda = v/n$, where λ is the wave length, v is the velocity, and n is the frequency (table 20). Frequencies between about 25 and 10,000 vibrations per second are easily detected by the human ear, and the maximum sensitivity is at about 1000 per second. Ultrasonic waves of frequencies above those audible to the human ear have certain desirable properties that make them valuable in submarine acoustics, but, as will

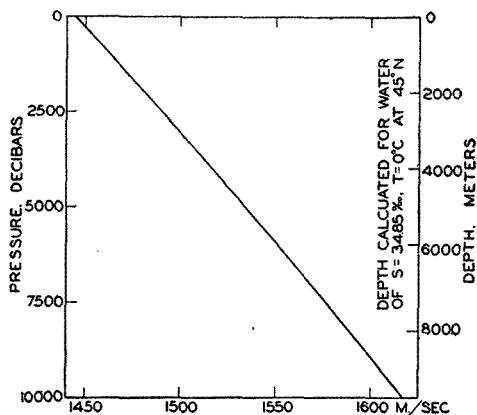


Fig. 18. Effect of pressure upon the velocity of sound in sea water of salinity 34.85 ‰ at 0°.

be shown later, their effective range is much less because their absorption is greater. The effective range has been increased by constructing sound emitters which for small wave lengths give a directional beam, somewhat

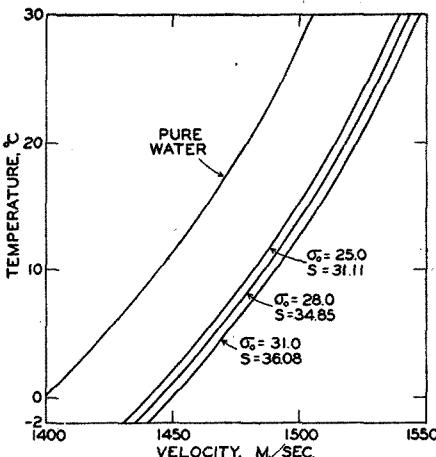


Fig. 17. Velocity of sound in pure water and in sea water at atmospheric pressure as a function of temperature and salinity.

analogous to that of a searchlight. In practice the spread is about 10° to 15° . A directional beam serves not only to measure the distance to some reflecting body, but also to determine the direction to the body.

In the absence of appreciable absorption and refraction the intensity of sound varies inversely as the square of the distance from the source. Owing to the viscosity of the water, a certain amount of the kinetic energy of the sound waves is converted into heat, and hence there is an absorption of sound somewhat analogous to that of light. The problem of absorption of sound in sea water has been discussed by Langevin (1924). The sound intensity of a plane sound wave decreases exponentially from I_0 to I_x by passage through a distance x . Therefore

$$I_x = I_0 e^{-2\nu x},$$

where $\nu = 8\pi^2\mu/3\lambda^2\rho v$ and corresponds to the absorption coefficient for radiant energy. The distance d over which the intensity is reduced to $1/e$ (approximately $\frac{1}{3}$) is therefore

$$d(\text{cm}) = \frac{3\lambda^2\rho v}{16\pi^2\mu},$$

TABLE 20
WAVE LENGTHS IN AIR AND WATER FOR SOUND OF DIFFERENT FREQUENCIES

Frequency (n)	Air	Sea water	Sea water
	$\vartheta = 20^{\circ}$, $p = 1 \text{ atm}$ $v = 346 \text{ m/sec}$	$S = 34.85 \%_{\text{oo}}$ $\vartheta = 0^{\circ}$, $p = 1 \text{ atm}$ $v = 1445.2 \text{ m/sec}$	$S = 34.85 \%_{\text{oo}}$ $\vartheta = 20^{\circ}$, $p = 1 \text{ atm}$ $v = 1518.5 \text{ m/sec}$
	wave length, λ (m)	wave length, λ (m)	wave length, λ (m)
10.....	35.6	144.5	151.8
100.....	3.56	14.45	15.18
1,000.....	0.36	1.44	1.52
10,000.....	0.036	0.144	0.152
40,000.....	0.009	0.036	0.038
100,000.....	0.0036	0.0144	0.0152

where all values are in c.g.s. units. The quantities λ , ρ , and v have already been defined, and μ is the dynamic viscosity. The ranges of ρ and v are small, and therefore d varies as λ^2/μ . Hence the absorption increases rapidly with increasing frequency and somewhat with increasing viscosity, and is significant only for ultrasonic waves. According to Bergmann

(1939) the absorption in water is much greater than is indicated by the equations above. Hartmann and Focke (1940) have obtained experimental data which indicate that the absorption is approximately a thousandfold larger. Whether absorption in the sea is as great or greater than that shown by these laboratory tests must await investigation in the field.

The velocity with which a sound wave travels through the water varies considerably with depth. Hence a "beam" of ultrasonic waves may be refracted when transmitted in a quasi-horizontal direction. Generally the velocity decreases with increasing depth in the upper layers and the beam is bent downward. Studies made by the U. S. Coast and Geodetic Survey (Swainson, 1936) have shown that a sound impulse may travel directly to the hydrophone, or may reach the hydrophone after having been reflected one or more times at the surface and the bottom. In many cases it was possible to distinguish several different "rays" which were received after different intervals. The direct transmission could be obtained only at distances less than 20 km and when the depth to the bottom was about 2000 m. The velocity of the direct wave impulse corresponded to that computed from the temperature and salinity, but those wave impulses which were reflected showed an apparent "horizontal" velocity less than the theoretical because of the longer paths traveled. This empirical horizontal velocity depends upon the distance between the vessel and the hydrophone, the depth to the bottom, the bottom profile, the physical properties of the water, and so forth.

As stated previously, the vertical velocity is a function of the depth and the distribution of temperature and salinity. Most sonic depth-finding instruments are adjusted for a constant "sounding velocity," usually 800 to 820 fathoms per second (1463 to 1500 m/sec). In certain cases it is desirable to correct the readings to true depths. This can be done if the distributions of temperature and salinity are known. For different areas of the oceans and for different depths, the British Admiralty Tables contain "mean sounding velocities"; that is, mean velocities from the surface to the stated depth. In general, these first decrease somewhat with depth, because the effect of decreasing temperature dominates, but at greater depths they increase again as the pressure

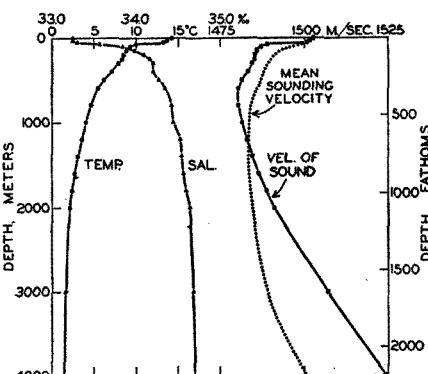


Fig. 19. Vertical distributions of temperature and salinity off southern California at $32^{\circ}57'N$ and $122^{\circ}07'W$, the computed velocity of sound, and the mean sounding velocity.

effect becomes dominant. In fig. 19 are shown the vertical temperature and salinity distributions at a station off the coast of southern California together with the corresponding velocity of sound at all depths as computed from the British Admiralty Tables. The mean sounding velocity is also shown. The latter decreased from 1503 m/sec at the surface to a minimum of 1484 m/sec for depths between 800 and 1800 m, and then increased again to 1496 m/sec at 4000 m.

Absorption of Radiation

ABSORPTION COEFFICIENTS OF DISTILLED WATER AND OF PURE SEA WATER. In water the intensity of parallel beams of radiation of wave length λ decreases in the direction of the beams, the decrease in a layer of infinitesimal thickness being proportional to the energy, I , and to the thickness of the layer:

$$dI_\lambda = -\kappa_\lambda I_\lambda dx.$$

The coefficient of proportionality, κ , is called the *absorption coefficient*. Integrating this equation between the limits $x = h$ and $x = h + L$, one obtains

$$\kappa_\lambda = \frac{2.30}{L} (\log I_{\lambda,h} - \log I_{\lambda,(h+L)}),$$

where the factor 2.30 enters because base-10 logarithms are used instead of natural logarithms, and where L is the thickness of the layer within which the energy of the radiation is reduced from $I_{\lambda,h}$ to $I_{\lambda,(h+L)}$. The latter equation also serves as a definition of the absorption coefficient. The numerical value of the absorption coefficient depends upon the unit of length in which L is expressed. In physics the unit is 1 cm, but in oceanography it has become common practice to use 1 m as the unit of length. Therefore the numerical values of the coefficients that will be given here are one hundred times larger than the corresponding values given in textbooks of physics.

The decrease of intensity of radiation passing through a layer of water depends not only upon the amount of radiation that is truly absorbed—that is, converted into another form of energy—but also upon the amount that is scattered laterally. In “pure” water the scattering takes place against the water molecules, and the effect of scattering is related to the molecular structure of the water (p. 47). However, when the absorption in pure water is measured, the effect of scattering is not separated but is included in the absorption coefficient, which varies greatly with wave length.

A great number of measurements of the absorption coefficients in pure water have been conducted, but the results by different investigators do not agree (Dorsey, 1940). Thus, at a wave length of .48 μ

($1 \mu = 0.0001$ cm), at which the absorption is very small, the following values have been reported:

Hüfner and Albrecht, 1891.....	0.048
Ewan, 1895.....	0.030
Aschkinass, 1895.....	0.020
Sawyer, 1931.....	0.015

Owing to such discrepancies the absorption in pure water is not exactly known, but, as a basis for comparison, table 21 contains values of the absorption coefficients according to determinations by W. R. Sawyer in the range 0.35μ to 0.65μ , and by J. R. Collins for wave lengths greater than 0.65μ (Dietrich, 1939). Sawyer's results have been selected because Clarke and James (1939) have obtained similar values in their examination of filtered sea water.

From the table it is evident that water is very transparent for radiation of wave lengths between 0.4μ and 0.6μ ; that is, for visible rays in the violet, blue, green, and yellow parts of the spectrum. It is less transparent for orange and red light, and in the infrared the transparency is practically nil (fig. 21), because, if the absorption coefficient per meter equals 100, 99.5 per cent of the radiation is absorbed in a layer of thickness 5.3 cm.

TABLE 21
ABSORPTION COEFFICIENTS PER METER OF PURE WATER AT WAVE LENGTHS BETWEEN $.32 \mu$ AND $.65 \mu$, ACCORDING TO W. R. SAWYER,
AND AT WAVE LENGTHS GREATER THAN $.65 \mu$, ACCORDING TO
J. R. COLLINS

Wave length in μ	Absorption coefficient per meter	Wave length in μ	Absorption coefficient per meter	Wave length in μ	Absorption coefficient per meter	Wave length in μ	Absorption coefficient per meter
.32.....	0.58	.52....	0.019	.85....	4.12	1.60....	800
.34.....	0.38	.54....	0.024	.90....	6.55	1.70....	730
.36.....	0.28	.56....	0.030	.95....	28.8	1.80....	1700
.38.....	0.148	.58....	0.055	1.00....	39.7	1.90....	7300
.40.....	0.072	.60....	0.125	1.05....	17.7	2.00....	8500
.42.....	0.041	.62....	0.178	1.10....	20.3	2.10....	3900
.44.....	0.023	.65....	0.210	1.20....	123.2	2.20....	2100
.46.....	0.015	.70....	0.84	1.30....	150	2.30....	2400
.48.....	0.015	.75....	2.72	1.40....	1600	2.40....	4200
.50.....	0.016	.80....	2.40	1.50....	1940	2.50....	8500

Collins has compared the absorption in distilled water to that in salt solutions, and from his results it can be concluded that the dissolved salts in concentrations occurring in sea water exert a negligible effect on the absorption coefficient. The maximum effect appears to be about 1.3 per cent, and the uncertainty of the observed values is greater than

this amount. These results have been confirmed by Clarke and James (1939), who found that the absorption of pure sea water as represented by Berkefeld-filtered oceanic sea water was practically identical with that of distilled water. Their observations indicate that Sawyer's values for distilled water may be too high in the ultraviolet; that is, at wave lengths smaller than 0.38μ .

It has also been concluded that the effect of temperature on absorption, which has been established in the case of distilled water, is applicable to uncontaminated sea water. The effect of temperature changes is to increase the absorption in certain parts of the infrared by about 0.5 per cent for every temperature increase of 1°C , but over a large part of the spectrum the temperature effect is much smaller. When dealing with sea water the effect can be neglected.

EXTINCTION COEFFICIENTS IN THE SEA. In oceanography the greater interest is attached to the rate at which downward-traveling radiation decreases. The rate of decrease can be defined by means of a coefficient similar to the absorption coefficient:

$$\kappa_{\lambda} = 2.30 (\log I_{\lambda,z} - \log I_{\lambda,(z+1)}),$$

where $I_{\lambda,z}$ and $I_{\lambda,(z+1)}$ represent the radiation intensities of wave length λ on horizontal surfaces at the depths z and $(z + 1)$ m. Different names have been proposed for this coefficient, such as *transmissive exponent* (Clarke, 1933) or *extinction coefficient* (Pettersson, 1936a). The latter name has been widely used and will be employed here, although the process by which the intensity of radiation is reduced will be called *absorption*. The absorption of radiation in the sea is complicated by the increased scattering due to suspended particles and by the presence of dissolved colored substances. The extinction coefficient of radiation of a given wave length therefore varies within wide limits from one locality to another, and in a given locality it varies with depth and time.

The first crude measurements of absorption in the visible part of the spectrum were made by lowering a white disc of standard size (30 cm), the Secchi disc, and observing the depth at which the disc disappeared from sight. Comparisons with recent exact measurements by other methods have shown that in the English Channel the extinction coefficient of visible rays can roughly be attained from the formula $\kappa = 1.7/D$, where D is the maximum depth of visibility in meters, as determined by the Secchi disc (Poole and Atkins, 1929).

The next step in the investigation of the absorption of radiation in sea water was made by subsurface exposure of photographic plates enclosed in watertight containers. Such experiments, which were conducted by Helland-Hansen (1912a) on the *Michael Sars* Expedition by exposing panchromatic plates at different depths in the vicinity of the Azores, showed that photographic plates were blackened at very great depths.

A plate exposed for 40 minutes at a depth of 500 m showed strong blackening, another exposed for 80 minutes at 1000 m was also blackened, but a third plate which was exposed for 120 minutes at 1700 m showed no effect whatever. These experiments were made at noon on June 6, 1910, with a clear sky. At 500 m it was found that the radiation had a distinct downward direction, because plates exposed at the top of a cube were much more strongly blackened than those exposed on the sides.

In other experiments colored filters were used, which showed that the red portion of the spectrum was rapidly absorbed, whereas the green and blue rays penetrated to much greater depths. Quantitative results as to the absorption at different wave lengths were obtained by using spectrophotometers (Knudsen, 1922), but the methods were laborious and not sensitive enough to be used at great depths.

The introduction in recent years of photoelectric cells has made possible rapid and accurate determinations of extinction coefficients in different parts of the spectrum. A number of different instruments have been and still are in use, but a standardized technique has been proposed by a committee of the International Council for the Exploration of the Sea (Atkins *et al*, 1938). Because of the wide variation in absorption at different wave lengths, efforts have been directed toward measuring exactly the absorption in narrow spectral bands. The determinations are accomplished by lowering stepwise a photoelectric or photronic cell enclosed in a watertight container and provided with suitably colored filters, and by observing on deck the photoelectric current by means of a sensitive galvanometer or a suitable bridge circuit. The measurements must be made at constant incident light either on clear, sunny days or on days when the sky is uniformly overcast, because the rapid variations in incident light that occur on days with scattered clouds will naturally lead to erroneous results as to the absorption. To determine the percentage amount of radiation that reaches a certain depth, it is necessary to make simultaneous readings of the incident radiation on board ship. For the different precautions that have to be taken, reference is made to papers listed in the bibliography, particularly to Atkins *et al* (1938).

These methods give information as to the absorption in layers of definite thickness. Instruments for measurements of the transparency of sea water at given depths and of the scattering of light have been designed by H. Pettersson (1936b) and have been used for determining relative values. It has been demonstrated, particularly, that at boundary surfaces sharp variations in transparency and scattering occur. The study of the absorption of radiation in the sea is in rapid progress, and several of the following generalizations are therefore presented with reservations.

The main results as to the character of the extinction coefficient in the sea of radiation of different wave lengths can be well illustrated

by means of data which Utterback (1936) and Jorgensen and Utterback (1939) have published. Utterback attempted to determine the extinction coefficients within spectral bands as narrow as possible, and has assigned the observed coefficients to distinct wave lengths, but it should be understood that the wave length actually stands for a spectral band of definite width. He has made numerous observations in the shallow waters near islands in the inner part of Juan de Fuca Strait and at four stations in the open oceanic waters off the coast of Washington, and these can be considered typical of coastal and oceanic water, respectively. Table 22 contains the absorption coefficients of pure water at the wave

TABLE 22
ABSORPTION COEFFICIENTS PER METER IN PURE WATER AND
EXTINCTION COEFFICIENTS IN THE SEA
(From Utterback's data)

Type of water	Wave length in μ								
	.46	.48	.515	.53	.565	.60	.66	.80	1.00
Pure water.....	.015	.015	.018	.021	.033	.125	.280	2.40	39.7
Oceanic water	lowest.....	.038	.026	.035	.038	.074	.199		
	average.....	.086	.076	.078	.084	.108	.272		
	highest.....	160	154	143	140	167	333		
Coastal water	lowest.....	.224	.230	.192	.169		.375	.477	
	average.....	.362	.334	.276	.269		.437	.623	
	highest.....	510	454	398	348		.489	.760	

lengths used by Utterback, the minimum, average, and maximum extinction coefficients observed in oceanic water, and the minimum, average, and maximum coefficients observed in coastal water. The minimum and maximum coefficients have all been computed from the four lowest and the four highest values in each group. The data in table 22 are represented in fig. 20. In the clearest oceanic water the extinction coefficients were only twice those of pure water and the average values were four to five times the latter, whereas the maximum values were up to ten times as great. In the coastal waters the minimum values were up to sixteen times greater than the absorption coefficients of pure water, the average values were up to twenty-four times as great, and the maximum values were up to thirty-four times as great. The increase of the extinction coefficients, however, varied widely in the different parts of the spectrum and was much greater for shorter wave lengths than for longer.

The transparency of the water for radiation of different wave lengths can be expressed by means of the percentage amounts of radiation which penetrate a 1-m layer. These percentage amounts are given in table 23, from which it is seen that the greatest transparency of the clearest oceanic water is at a wave length of 0.48μ —that is, in the blue part of the spectrum—whereas the greatest transparency of coastal water is at wave lengths 0.53μ or higher—that is, in the green or green-yellow part of the spectrum. It is also seen that 97.5 per cent of radiation of wave length 0.48μ passes through 1 m of the clearest oceanic water, but

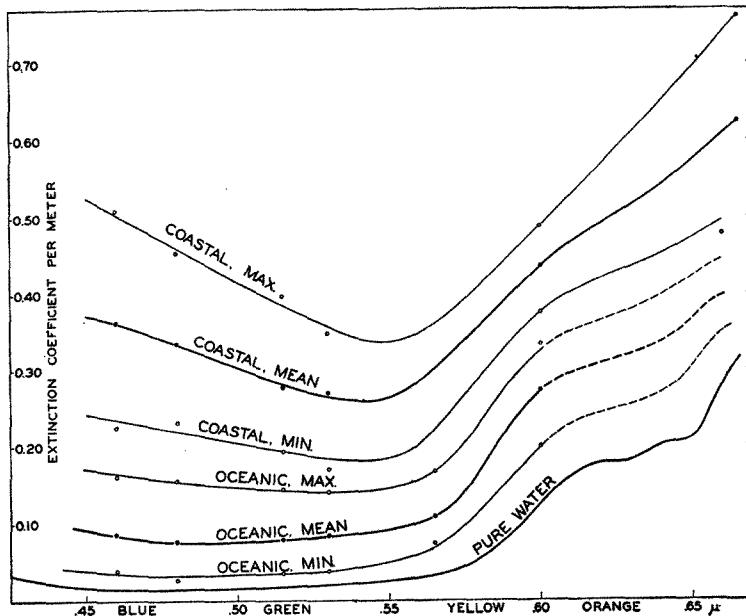


Fig. 20. Extinction coefficients of radiation of different wave lengths in pure water and in different types of sea water.

only 63.5 per cent of radiation of the same wave length passes through 1 m of turbid coastal water.

The great difference between the mean and the maximum values of the extinction coefficients shows that the absorption of sea water varies within very wide limits. In the example presented in table 22 the percentage variations are about the same in coastal and oceanic water, and the maximum values in the oceanic water approach the minimum values in the coastal water. In any given locality, great variations also occur in a vertical direction, layers of low absorption alternating with layers of high absorption, and this feature further complicates the actual conditions.

Similar results have been obtained by other investigators from such widely different areas as in the English Channel (Poole and Atkins, 1929;

TABLE 23
PERCENTAGE OF RADIATION OF GIVEN WAVE LENGTH TRANSMITTED
BY 1 M OF WATER
(Based on data in table 22)

Type of water	Wave length (μ)							
	.46	.48	.515	.53	.565	.60	.66	
Pure water.....	98.5	98.5	98.2	97.9	96.8	88.3	75.9	
Oceanic water	lowest.....	96.4	97.5	96.6	96.3	92.9	81.8	
	average.....	91.8	92.7	92.5	91.8	89.8	75.9	
	highest.....	85.1	85.7	86.7	86.9	84.5	71.6	
Coastal water	lowest.....	80.0	79.4	82.6	84.5		68.7	62.0
	average.....	69.7	71.6	75.9	76.4		64.6	53.6
	highest.....	60.0	63.5	67.1	70.6		61.4	46.7

Poole, 1936), in the waters off the east coast of the United States (Clarke, 1933), and off southern California (Young and Gordon, 1939). In all instances it has been found that the absorption is less in oceanic than in coastal water, but it varies within wide limits both locally and with depth. Where examination of absorption in different parts of the spectrum has been conducted, it has been found that the absorption is much less in the blue than in the red end of the spectrum and that the blue light penetrates to the greatest depths in clear water, whereas the green or yellow light reaches further down in turbid water.

INFLUENCE OF THE ALTITUDE OF THE SUN UPON THE EXTINCTION COEFFICIENT. The extinction coefficient is a measure of the reduction of intensity on a vertical distance and depends, therefore, upon the obliquity of the rays. The obliquity of the incident rays is reduced, however, by refraction when entering the water from the air and by the effect of scattering. When the sun's rays pass the water surface, the angle of refraction increases from zero with the sun in zenith to 48.5 degrees with the sun at the horizon, and therefore the most oblique rays penetrating into the water form an angle of less than 48 degrees with the vertical. Owing to the scattering and the sifting out by absorption of the most oblique rays, the radiation that penetrates to moderate depths will become nearly vertical, and the measured extinction coefficients will be independent, within wide limits, of the altitude of the sun. The reduction of the obliquity of the incident radiation has been directly demonstrated by Johnson and Liljequist (1938). Conditions at very low sun have not been examined, but it is probable that at low sun the extinction coefficients are increased, and this increase may have bearing

upon the diurnal variation of the incoming energy at greater depths (p. 779).

THE SCATTERING OF RADIATION IN THE SEA. The scattering of radiation has been examined both directly by means of Pettersson's scattering meter (p. 83) and by measuring the relative intensities of downward- and upward-traveling radiation or vertical and horizontal radiation. Jorgensen and Utterback (1939) found that in coastal waters the intensity of the upward-traveling radiation ranged for the short wave lengths from 1 to 3 per cent of that of the downward-traveling radiation, and for the long wave lengths from 0.5 to 2 per cent. In oceanic water Utterback (1936) found ratios between 1 and 2 per cent at the shorter wave lengths. Clarke (1936) found considerably higher values in shallow coastal waters, but similar values in the deep basin of the Gulf of Maine.

The relative intensities of horizontal and vertical radiation have been measured by Clarke off the east coast of the United States and by Poole and Atkins in the English Channel. The greatest value found by Clarke was 17 per cent, but Poole and Atkins (1929) have reported an average value of 50 per cent for the horizontal radiation down to a depth of 25 m in the English Channel. The conclusion that can be drawn from these experiments is that the subsurface illumination becomes more and more diffuse with increasing depth, particularly in coastal waters, but that the directional character of the radiation is lost only slowly. This conclusion is particularly true in clear oceanic water, where Helland-Hansen (1912a) found that the vertical radiation was distinctly more intense than the horizontal at a depth of 500 m (p. 82).

CAUSE OF THE GREAT EXTINCTION COEFFICIENTS IN THE SEA. The great extinction coefficients in the sea as compared to those of absolutely pure water are as a rule ascribed to the presence of minute particles which cause scattering and reflection of the radiation and which themselves absorb radiation. If such particles are small compared to the wave length, λ , of the radiation, the scattering will be proportional, according to Lord Rayleigh, to λ^{-4} , and the effect therefore at wave length, say, $.46 \mu$ will be 2.86 times greater than at wave length $.60 \mu$. This selective effect leads to a shift toward longer wave lengths in the region of minimum absorption.

Clarke and James (1939) found that the increased absorption in oceanic water was chiefly caused by suspensoids that could be removed by means of a "fine" Berkefeld filter, and that these suspensoids were largely nonselective in their effect. Utterback's data (1936) indicated, on the other hand, that the increased absorption in oceanic water is at least in part due to selective scattering, because at short wave lengths the extinction coefficients were increased more, above those of pure water, than at longer wave lengths (table 22). Kalle (1938) is of the

opinion that selective scattering is of dominant importance (p. 89), but the question is not yet settled as to the mechanism which leads to increased absorption in oceanic water as compared to pure water. The fact that even in the clearest oceanic water the absorption is greater than in pure water indicates, however, that finely suspended matter is always present. One could state that the ocean waters always contain dust.

The increase of the absorption coefficients in coastal waters appears to be due in part to another process. Clarke and James (1939) conclude from their examination that in coastal water both suspensoid and "filter-passing" materials are effective in increasing the absorption, and that each exerts a highly selective action, with greatest absorption at the shorter wave lengths. These great absorptions at the shorter wave lengths are demonstrated by Utterback's measurements (table 22). Clarke does not discuss the nature of the "filter-passing" material, but Kalle (1938) has shown that in sea water water-soluble pigments of yellow color are present. These pigments appear to be related to the humic acids, but their chemical composition has not been thoroughly examined, for which reason Kalle calls them "yellow substance." This yellow substance seems to occur in greatest abundance in coastal areas, but Kalle has demonstrated its presence in the open ocean as well and believes that it represents a fairly stable metabolic product related to the phytoplankton of the sea. The selective absorption of this yellow substance may then be responsible, in part, for the character of the absorption in coastal water and for shift of the band of minimum absorption toward longer wave lengths.

It has not been possible anywhere to demonstrate any direct influence of phytoplankton populations on the absorption, but very dense populations may cut down the transparency. At present it appears that the major increase of absorption of sea water over that of pure water is due to two factors: the presence of minute suspended particles, and the presence of dissolved "yellow substance." The former factor dominates in oceanic waters, and the latter is particularly important in coastal waters.

THE COLOR OF THE SEA. The color of the sea, as it appears to an observer ashore or on board a vessel, varies from a deep blue to an intense green, and is in certain circumstances brown or brown-red. The blue waters are typical of the open oceans, particularly in middle and lower latitudes, whereas the green water is more common in coastal areas, and the brown or "red" water is observed in coastal regions only.

The color of the sea has been examined by means of a Secchi disk (p. 82) by observing the color that the water appears to have when seen against the white submerged surface of the disc. This color is recorded according to a specially prepared color scale, the "Forel scale" (Krümmel, 1907). The method is a rough one and the scale is not adapted for

recording the extreme colors in coastal waters. In order to obtain more exact results, Granquist, working in the Finnish waters, used a long tube that was blackened on the inside, and this type of instrument has been greatly improved by Kalle (1938).

Kalle has critically reviewed earlier theories as to the causes of the color of the sea and arrives at conclusions that appear to be consistent with all available observations. The blue color is explained, in agreement with earlier theories, as a result of scattering against the water molecules themselves, or against suspended minute particles smaller than the shortest visible wave lengths. The blue color of the water is therefore comparable to the blue color of the sky. The transition from blue to green cannot be explained, however, as a result of scattering, and Kalle concludes that this transition is due to the above-mentioned "yellow substance," pointing out that the combination of the yellow color and the "natural" blue of the water leads to a scale of green colors as observed at sea. Fluorescence may contribute to the coloring but appears to be of minor importance.

The color of suspended larger particles, if present in great abundance, can give color to the sea. In this case the color is not determined by the optical properties of the water or by dissolved matter, but by the colors of the suspended inorganic or organic particles, and the water is appropriately called "discolored." Discoloration can be observed when large quantities of finely suspended mineral particles are carried into the sea after heavy rainfall, or when very large populations, several million cells per liter, of certain species of algae or dinoflagellates are present very near the surface. Thus, the "red water" (often more brown than red) which is quite frequently observed in many areas and after which the Red Sea and the Vermilion Sea (Gulf of California) have been named is due to abundance of certain algae (in the Red Sea, *Trichodesmium erythraeum*) or dinoflagellates. The discoloration, beautiful examples of which have been given by Gunther (1936), is, however, a phenomenon of the typical coastal waters, the green colors being frequent in waters near the coast or at sea in high latitudes, and the blues characteristic of the open ocean in middle and lower latitudes (fig. 214, p. 784).

Eddy Conductivity, Diffusivity, and Viscosity

In the preceding discussion it has been repeatedly stated that the coefficients of heat conduction, diffusion, and viscosity that have been dealt with so far are applicable only if the water is at rest or in laminar flow. By *laminar flow* is understood a state in which sheets (*laminae*) of liquids move in an orderly manner such that random local fluctuations of velocity do not occur. However, the molecules of the liquid, including those of dissolved substances, move at random, and, owing to this

random motion, an exchange of *molecules* takes place between adjacent layers. Consequently, there is a transfer of heat if adjacent layers are at different temperatures, a diffusion of dissolved substances if their concentrations are variable in space, or a transfer of momentum if their velocities differ. The amounts that are transferred are proportional to the gradients of temperature, concentration, or velocity, and the factors of proportionality—that is, the coefficients of thermal conductivity, diffusion, and viscosity—are among the characteristic properties of the liquid. For a given liquid they are functions of temperature, pressure, and concentration, and can be exactly determined in the laboratory.

In nature, laminar flow is rarely or never encountered, but, instead, *turbulent flow*, or *turbulence*, prevails. By *turbulent flow* is understood a state in which random motion of smaller or larger *masses* of the fluid is superimposed upon some simple pattern of flow. The character of the turbulence depends upon a number of factors, such as the average velocity of the flow, the average velocity gradients, and the boundaries of the system. Under these conditions the exchange between adjacent moving layers is not limited to the interchange of molecules, but masses of different dimensions also pass from one layer to another, carrying with them their characteristic properties. As a consequence, a snapshot of the instantaneous distribution of velocity, temperature, salinity, and other variables in the sea would show a most complicated pattern, but so far no means have been developed for establishing this picture. Measurements by sensitive current meters have demonstrated that in a given locality the velocity fluctuates from second to second, but in most cases observations of ocean currents give information as to *mean* velocities for time intervals that may vary from a few minutes to twenty-four hours or more. Similarly, special measurements have demonstrated that the details of the temperature distribution are very complicated, but in general observations are made at such great distances apart that only the major features of the temperature distribution are obtained. Inasmuch as it is impossible to observe the *instantaneous* distribution in space of temperature, salinity, and velocity, it follows that the corresponding gradients cannot be determined and that no basis exists for application to the processes in the sea of the coefficients of thermal conductivity, diffusion, and viscosity that have been determined in the laboratory. Since only certain *average* gradients can be determined, another approach has to be made when dealing with the processes in the sea. In order to illustrate this approach, let us first consider the viscosity.

In the case of *laminar* flow the coefficient of viscosity, μ , is defined by the equation $\tau_s = \mu dv/dn$ where τ_s is the shearing stress exerted on a surface of unit area, and dv/dn is the shear normal to that surface.

In the case of turbulent flow a coefficient of eddy viscosity, A , can be defined in a similar manner:

$$\tau_s = A \frac{d\bar{v}}{dn},$$

where $d\bar{v}/dn$ now represents the shear of the observed velocities. The numerical value of the eddy viscosity depends upon the size and intensity of the eddies—that is, on the magnitude of the exchange of masses between adjacent layers; and the symbol A that is commonly used is an abbreviation for the term "Austausch" introduced by Schmidt (1917). The numerical value of A also depends upon how the "average" velocities have been determined; that is, upon the distribution in space of the observations and upon the length of the time intervals to which the averages refer.

The definition of the eddy viscosity in the above manner appears purely formalistic, but it is based on the concept that masses which leave one layer carry with them the momentum corresponding to the *average* velocity in that layer, and that by impact they attain the momentum corresponding to the average velocity of their new surroundings before again leaving them (p. 472). Thus, A is an expression for the transfer of momentum of mean motion. This transfer is much increased by the turbulence, as is evident from the fact that the eddy viscosity is many times greater than the molecular viscosity.

The eddy viscosity can be determined only by examination of the effect on the mean motion. This effect is discussed on pp. 492 and 577, but a few points will be mentioned here. It has been found practical to distinguish between two types of turbulence in the sea—vertical and lateral. In the case of vertical turbulence the *effective* exchange of masses is related to comparatively slight random motion in a vertical direction or, if the term "eddy motion" is used, to small eddies in a vertical plane. Actually, the eddies are oriented at random, but only their vertical components produce any effect on the mean motion. The corresponding eddy viscosity has been found to vary between 1 and 1000 c.g.s. units, thus being one thousand to one million times greater than the molecular viscosity of water. In the case of lateral turbulence the *effective* exchange of masses is due to the existence of large quasi-horizontal eddies. The corresponding eddy viscosity depends upon dimensions of the system under consideration and has been found to vary between 10^6 and 10^8 c.g.s. units.

The distinction between vertical and lateral turbulence is particularly significant where the density of the water increases with depth, because such an increase influences the two types of turbulence in a different manner. Where the density of the sea water increases with depth (disregarding the effect of pressure), vertical random motion is impeded by

Archimedean forces, because a mass which is brought to a higher level will be surrounded by water of less density and will tend to sink back to the level from which it came, and, similarly, a water mass moving downward will be surrounded by denser water and will tend to rise. In this case the stratification of the water is called *stable*, because it cannot be altered unless work against gravity is performed. Stable stratification reduces the vertical turbulence; where the stability is very great the vertical turbulence may become nearly suppressed and the eddy viscosity small. The effect of stability on the lateral turbulence, on the other hand, is negligible, because the lateral random motion takes place mainly along surfaces of equal density.

With regard to the eddy conductivity, similar reasoning is applicable. When dealing with eddy viscosity it was assumed that the exchange of mass leads to a transfer of momentum from one layer to another, which is expressed by means of A . Correspondingly, when dealing with eddy conductivity, one can assume that the transfer of heat through any surface is proportional to the exchange of mass through the surface, as expressed by A , and to the gradient of the observed temperatures, $-d\bar{\theta}/dn$; that is, $dQ/dt = -rA d\bar{\theta}/dn$, where r is a factor that depends upon the specific heat of the fluid and upon the manner in which the heat contents of the moving masses are given off to the surroundings. When dealing with *homogeneous* water, it is assumed that a large mass which is transferred to a new level breaks down at that level into smaller and smaller elements, and that equalization of temperature ultimately takes place by molecular heat conduction between the small elements and the surroundings. If such is the case, both the difference in momentum and the difference in heat content are leveled off, and the proportionality factor, r , is equal to the specific heat of the liquid. Since the specific heat of water is nearly unity, the numerical values of eddy conductivity and eddy viscosity are practically equal. However, where *stable stratification* prevails, the elements, being lighter or heavier than their surroundings, may return to their original level before completion of temperature equalization, but equalization of momentum may have been accomplished by collision. In this case, the factor of proportionality, r , will be smaller than the specific heat of the liquid; that is, in the sea, r is smaller than unity, and the eddy conductivity is smaller than the eddy viscosity. Thus, stable stratification reduces the vertical eddy viscosity. Taylor (1931) has presented the above reasoning in mathematical language (p. 476).

The discussion has so far been limited to a consideration of the vertical eddy conductivity, but lateral eddy conductivity due to lateral turbulence has also to be introduced. The numerical value of lateral

eddy conductivity must be nearly equal to that of the lateral eddy viscosity, because the lateral turbulence is not affected by stable stratification.

Numerical values of the coefficients of eddy conductivity can be derived only from a study of the effect of mixing processes on the observed distribution of temperature. Methods of such determinations and numerical values are presented on pp. 483 and 484. The results have confirmed the above conclusions and have also demonstrated that the eddy conductivity varies within wide limits. In the upper layers of the sea, where stable stratification prevails, the vertical coefficient of eddy viscosity varies between 1 and 1000, whereas the corresponding eddy conductivity is smaller and varies between 0.01 and 100; in homogeneous water, however, no difference has been established (p. 485). In the cases in which the lateral coefficients of viscosity and conductivity have been examined, nearly equal numerical values have been found in agreement with the conclusion that the stability of the stratification does not influence the lateral turbulence.

The transfer of salinity or other concentration is similar to the heat transfer. The eddy diffusivity is also proportional to the exchange of mass as expressed by A , the factor of proportionality being a pure number. In sea water of uniform density $r = 1$, but in the case of stable stratification, when complete equalization of concentration does not take place, $r < 1$; that is, the vertical eddy diffusivity is smaller than A and equals the eddy conductivity. This conclusion has also been confirmed by observations (p. 484).

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