SOUND TRANSMISSION IN SEA WATER

A PRELIMINARY REPORT

Prepared By The

Woods Hole Oceanographic Institution

For The

National Defense Research Committee

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CHAPTER I

INTRODUCTION

During the last 20 years or more commercial and naval ships have made increasing use of a variety of instruments in which the transmission of sound in sea water plays an important part. Throughout this development the engineers and physicists have devoted most of their attention to the instruments, both sending and receiving units, for from the beginning it seemed likely that through improvements in design and construction very great gains would result in the usefulness of such equipment. This has proved to be the case, and the instruments have now reached a high state of technical perfection, yet from the standpoint of naval tactics the great fluctuations in the effective range of some types of underwater sound equipment has been disappointing.

While improvements in oscillators or projectors (sending elements) and hydrophones (receiving elements) still continue, it now seems clear that one important part of the problem, namely the role played by the sea water, has perhaps received too little attention. Especially when sound must travel horizontally and near the surface, the distribution of temperature and dissolved salts at depths of less than 100 fathoms determines the range. At some seasons and in some areas these and other physical characteristics of the water can be very favorable to lateral sound transmission, and at other times or in other regions the reverse may be the case.

One result is that it has been difficult to make a reliable comparison of the performance of different types of equipment. Another is that full confidence cannot be placed in supersonic methods of detecting submarines, unless the sound transmitting qualities of the water are known to be good. This report is an attempt to summarize existing knowledge in a manner which will throw light on these difficulties, and also to explore the problem in such a way as to provide a basis for future investigations. Special attention will be given to the possibility of predicting the effective range of supersonic signaling methods in various parts of the ocean and at various seasons.

In studying the transmission of sound in sea water two fields of science must be utilized, physics and oceanography. The physics of the problem is relatively well known, but the oceanographic factors seem, until now, to have been generally overlooked. Thus in this preliminary report the latter will be emphasized. In order to round out the discussion a brief summary of the development and uses of the various types of sound equipment has been included.

CHAPTER II

DEVELOPMENT OF THE ART OF SUBMARINE SIGNALING

1. HISTORICAL BACKGROUND

Leonardo da Vinci (1452-1519) knew that water was a particularly good conductor of sound. He recorded the fact that ships could be heard "at a great distance" by use of a trumpet-shaped tube, one end of the tube being placed in the water and the other held to his ear. Presumably the sound heard was the splash of galley oars.

To signal between boats, the native fishermen in Ceylon produce a sharp sound by striking an earthenware pot, known as a "chatty", held underwater. This sound can be heard at considerable distances by placing the ear against the hull of the boat.

In 1826 J. D. Colladon and J. K. F. Sturm made the first reliable measurement of the speed of sound in water in Lake Geneva. They used a 140 lb. church bell, held 10 ft. under water and struck with a hammer, as a sound source. The flash from a charge of gunpowder which exploded when the bell was struck, signaled the instant of origin of the sound. The detector was a large ear trumpet held with one end in the water. They were able to signal to distances of 14000 yards in water which averaged 70 fathoms in depth. In 1841 Colladon repeated his experiment using a very much heavier church bell, and could still detect the sound at a distance of 35000 yards. He stated that the use of suitable equipment would permit ranges of several hundred kilometers under favorable conditions. He was thus the first to recognize and clearly point out the possibility of practical use of underwater sound signaling.

In 1888 M. Banare published "Les Collisions en Mer" in which he discussed fully the status of underwater sound technique. He developed an underwater microphone and gave a method for using the sound shadow of the ship to determine the bearing of the sound source. At about this time several British and German investigators made studies on underwater sound but none of them developed apparatus suitable for practical use. By 1898 it was evident that submarine signaling was feasible, but the apparatus which would make it commercially available had not been perfected.

2. EARLY DEVELOPMENTS

Starting in 1898 A. J. Mundy and E. Gray, later joined by J. B. Millet to form the Submarine Signal Company, developed a system which could be adapted to commercial use. In this system, the sound from a submerged bell could be heard at a considerable distance by means of a submerged microphone, but the microphones could only be used when cast overboard from a motionless vessel in a calm sea. Following this, Gray mounted the microphones in a fish-like housing and obtained satisfactory performance while towing it behind a moving ship.

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The next improvement was introduced by Gray and Mundy in 1902. It consisted of placing the microphones in water filled tanks built inside the hull well below the water line so that the hull formed one side of the tank. When such tanks were placed on both sides of the hull it was possible to determine the approximate bearing of the sound source. Specially designed submarine bells were constructed for use on lightships, the first being installed on Lightship No. 54 in Boston Harbor in 1903. The service given by this bell proved so valuable an aid to navigation under conditions of poor visibility that within a few years the system was adopted on many important lightships of the United States and also in other parts of the world. This system was installed on a number of naval vessels for signaling purposes, but the bell was not a particularly suitable instrument for sending signals in Morse code.

Many other investigators sought to find sound generators more suitable for sending Morse code. Water sirens, both the rotor and the oscillator type, were tried. Water-blown organ pipes and large underwater electromagnetic telephones were tested, all without success.

3. THE FESSENDEN OSCILLATOR

In 1912 R. A. Fessenden of the Submarine Signal Company produced an electrodynamic oscillator which, in contrast with previous electromagnetic types, exerted great forces on a diaphragm on both the "push" and the "pull" phases of the cycle. This device greatly increased the range of underwater sound signaling and permitted rapid transmission of Morse code. At the outbreak of war in 1914 it became standard equipment for our submarines and following the war it replaced the bells on many lightships, giving remarkable improvement in range and quality of the signals emitted. The Fessenden oscillators used in the United States had a frequency of 540 or 1050 cycles per second.

4. ECHO DEPTH SOUND

When suitable transmitters and receivers for underwater sound became available it was a natural step to apply them for measuring the depth of water by timing echoes from the ocean bottom.

In 1914 R. A. Fessenden tested his oscillator during the ice patrol cruise of the Coast Guard cutter MIAMI, in an effort to obtain echoes from icebergs. He detected signals from the ocean bottom, and the travel time of the echoes could be used to calculate the depth of water. In later tests he was able to measure echoes from 3000 fathoms off the Azores.

Following the world war development of echo sounding equipment was achieved by providing one of the Fessenden oscillators with an arrangement for accurately timing the echoes. This development was largely due to H. G. Dorsey. Various other systems have since been invented in America, France, Germany and Great Britain, which are similar to this system in that they use sound in the audible range of frequencies (usually about 500 to 1000 cycles per second). They differ only in details of the methods for generating, receiving and timing the sound waves.

In June 1922 the U.S.S. STEWART conducted an epoch-making test between Newport and Gibraltar. While maintaining a constant speed of 15 knots she averaged 100 echo soundings per day. A Fessenden type oscillator was used for transmitting sound signals and a Navy Sonic Depth Finder, developed by H. C. Hayes, was used for timing the echoes. Depths as great as 3200 fathoms were measured without difficulty. During November and December 1922 the U.S.S. CORRY and the U.S.S. HULL made hydrographic surveys from San Francisco to Point Descanso, Mexico, the results of which were shown on Hydrographic Office Chart No. 5194. These voyages were of greatest scientific interest and opened a new era in knowledge of oceanic depths.

5. HIGH FREQUENCY SOUND

The proposal to use supersonic waves (20,000 to 100,000 cycles per second) for submarine signaling was put forward by an Englishman, E. Richardson. He pointed out that at these frequencies it would be possible to produce directional beams of sound which could be directed like the beam of a searchlight. To produce such a beam, either the source of sound itself or a suitably shaped reflector must have dimensions several times as large as the wave length of sound used, and this is practical only when the waves are short. Richardson was unable to produce a suitable generator for sounds of high frequency.

About 1915 two Frenchmen, P. Langevin and M. Chilowsky, began experiments on production of high frequency sound by means of the mechanical action between the plates of a condenser, but this system could not be made to give sufficient intensity for practical use. The problem was simply one of transforming high frequency electrical oscillations into mechanical oscillations, and it was solved when Langevin thought of using the piezoelectric properties of quartz. If plates cut from these crystals at suitable angles are used as the insulator in a condenser and subjected to electric fields, they undergo changes in thickness. Conversely the same plates will generate electric charges on the condenser if they are subjected to mechanical pressure. A "piezo-electric sandwich" was constructed by placing a mosaic of thin quartz slabs between two plates of steel which were used as the plates of a condenser. By suitable choice of thickness of quartz and steel the resonant frequency of the "sandwich" could be given any desired value in the supersonic range. In 1924 a device of this type, about 4 inches in diameter, mounted in the hull of a ship with one steel plate in contact with the water gave a directional sound beam with sufficient power to transmit signals 4.9 miles, to obtain echoes from floating objects at 2200 yards, and to take soundings up to 245 fathoms. The same unit was used as a transmitter of the sound beam and receiver for the echo. It was not quickly taken into general use because the apparatus appeared too complicated for use at sea.

Rochelle salt crystals exhibit the phenomena of piezo-electricity more strongly than quartz and can be used to construct efficient sound projectors suitable for high frequency operation. These projectors have the advantage that they can be made to work efficiently without taking advantage of mechanical resonance, which is necessary in the case of quartz. Many of the projectors in use today are Rochelle-salt units.

Some of the projectors in use at present operate on the principle of magnetostriction. When placed in a magnetic field, a magneto-strictive metal changes its dimensions. High frequency changes in magnetic field, produced by high frequency electric currents, are used to set nickel tubes or plates into vibration at supersonic frequencies.

6. SUBMARINE DETECTION

For several reasons supersonic waves are far more useful than sound of lower frequency in detecting submarines. High frequency sound transmitters and detectors are strongly directional in their action, and they are less disturbed by noises originating in the ship or in the movement of the ship through the water. Furthermore a smaller target will serve as an efficient reflector for supersonic waves than for those of audible frequencies. Finally, when a supersonic detector is used in listening for sound generated by a submarine, it exhibits the same directional properties which it possesses in its application in echo ranging.

When water conditions are suitable for effective transmission of sound, supersonic methods are highly effective for submarine detection.

Supersonic echo ranging has been used by submarines travelling below periscope depth for determining both the bearing and the range of a target. For listening to sounds generated by the target, it is often preferable to use a detector sensitive to the supersonic components of these sounds instead of one designed to receive the low frequency components because the former can be made with far more directional sensitivity than the latter.

CHAPTER III

PHYSICS OF SOUND IN SEA WATER

1. THE SPEED OF SOUND IN SEA WATER

General Statement

The speed of sound in sea water depends on the temperature and composition of the water. In general these quantities vary from time to time, from place to place, and with the depth beneath the surface. Variations in the speed of sound play a most important part in echo ranging and in other methods of signaling in which sound is transmitted horizontally through sea water. Obviously changes in the speed of sound will cause changes in the time required for a signal to travel between two given points, but in echo ranging this effect is small and never serious. The important point is that changes of velocity with depth, even slight ones due to warming of the surface water on a bright calm day, deflect the sound beam from the horizontal plane and may cause it to undershoot or overshoot the target. These deflections of the sound beam are far more important than any other factor in the horizontal range of underwater sound.

The increase in the speed of sound per foot increase in depth is called the vertical velocity gradient. It usually varies with depth. It is the factor which determines the deflections of a sound beam, and it is therefore the most important property of the water from the viewpoint of echo ranging.

Factors Which Determine The Speed Of Sound.

The speed of sound at any point in the ocean may be calculated if the temperature, salinity and pressure of the water at that point are known. The temperature is determined by lowering a suitable thermometer to the desired point. The <u>salinity</u>, which is defined as the number of grams of salt in 1000 grams of sea water, can be determined chemically, electrically, or by use of a hydrometer, provided a sample of water from the point in question is available. The pressure depends primarily on the depth beneath the surface, and may readily be calculated.

In echo ranging work, where depths greater than a few hundred feet are not involved, the temperature is far more important than either of the other factors in determining the speed of sound. A temperature variation of only 0.1°F per hundred feet of depth can produce a marked influence on the range of a sonic beam, and changes many times larger than this are commonly present.

The salinity of the water in the open ocean is reasonably constant and does not introduce serious changes in the speed of sound. In coastal waters there are frequently larger variations in salinity due to the presence of river water which has not become thoroughly mixed with the sea water, but in nearly all situations thus far examined the temperature exerts more influence than salinity in determining the speed of sound.

The velocity changes caused by pressure are not large because the depths involved in echo ranging are slight. Although the pressure effect is small it is quite important in cases where the temperature and salinity are both constant, as in a thoroughly mixed surface layer. It will be shown later that in such cases the rays of the sound beam are bent upward into circles whose centers lie approximately 280,000 ft. or about 47 nautical miles above the surface. This situation is favorable for sound ranging, and maximum ranges will be obtained provided the mixed layer extends deep enough so that the lower rays of the beam do not run out of it before recurving.

All three factors change the velocity in the same direction. An increase in temperature, salinity, or pressure causes an increase in the velocity of sound. For purposes of comparison the following list shows the changes in temperature or salinity which produce velocity changes equal in magnitude to that caused by the effect of pressure alone:

$1^{\circ}F$ in 76 fm at $32^{\circ}F$

$1^{\circ}F$ in 61 fm at $50^{\circ}F$

$1^{\circ}F$ in 48 fm at $68^{\circ}F$

1 part per thousand salinity in 39 fm.

Calculation Of The Speed Of Sound

The two graphs in Figure 1 may be used for computing the speed of sound in sea water. The main part of this graph gives the speed in feet per second at zero depth and at any desired temperature and salinity. The auxiliary graph gives the correction which must be added to allow for the effect of pressure at a point beneath the surface.

In echo ranging work the vertical velocity gradient is more important than the velocity itself. It is most conveniently computed from the gradients of salinity and temperature. The <u>salinity gradient</u> is the rate of increase of salinity with depth, in parts per thousand ($^{o}/oo$) per foot. The <u>temperature gradient</u> is the rate of increase of temperature with depth, in degrees Fahrenheit per foot. The vertical velocity gradient in feet per second per foot, may be computed from the formula:

<u>Vertical velocity gradient = 0.0182+(K x temperature gradient)+(4.3 x salinity gradient)</u>, where 0.0182 is the gradient due to pressure and K, the coefficient of the temperature gradient, may be read from the graph of Figure 2 for the <u>mean</u> temperature of the range involved. These gradients are <u>positive</u> if the quantity in question <u>increases</u> with depth, <u>negative</u> if it <u>decreases</u>.

The graphs for computing the speed of sound are based on tables published in the "Hydrographic Review" Vol. XVI, pp. 123-40, 1939, by S. Kuwahara.

Sample Computations

(a) Depth = zero	From Figure 1, Temperature and salinity, . .	5022.0
Temperature = $75.2^{\circ}F$.	" " Pressure, ...	0.0
Salinity = 36.5 $^{o}/oo$	Sound speed	5022.0
Find the sound speed		ft/sec

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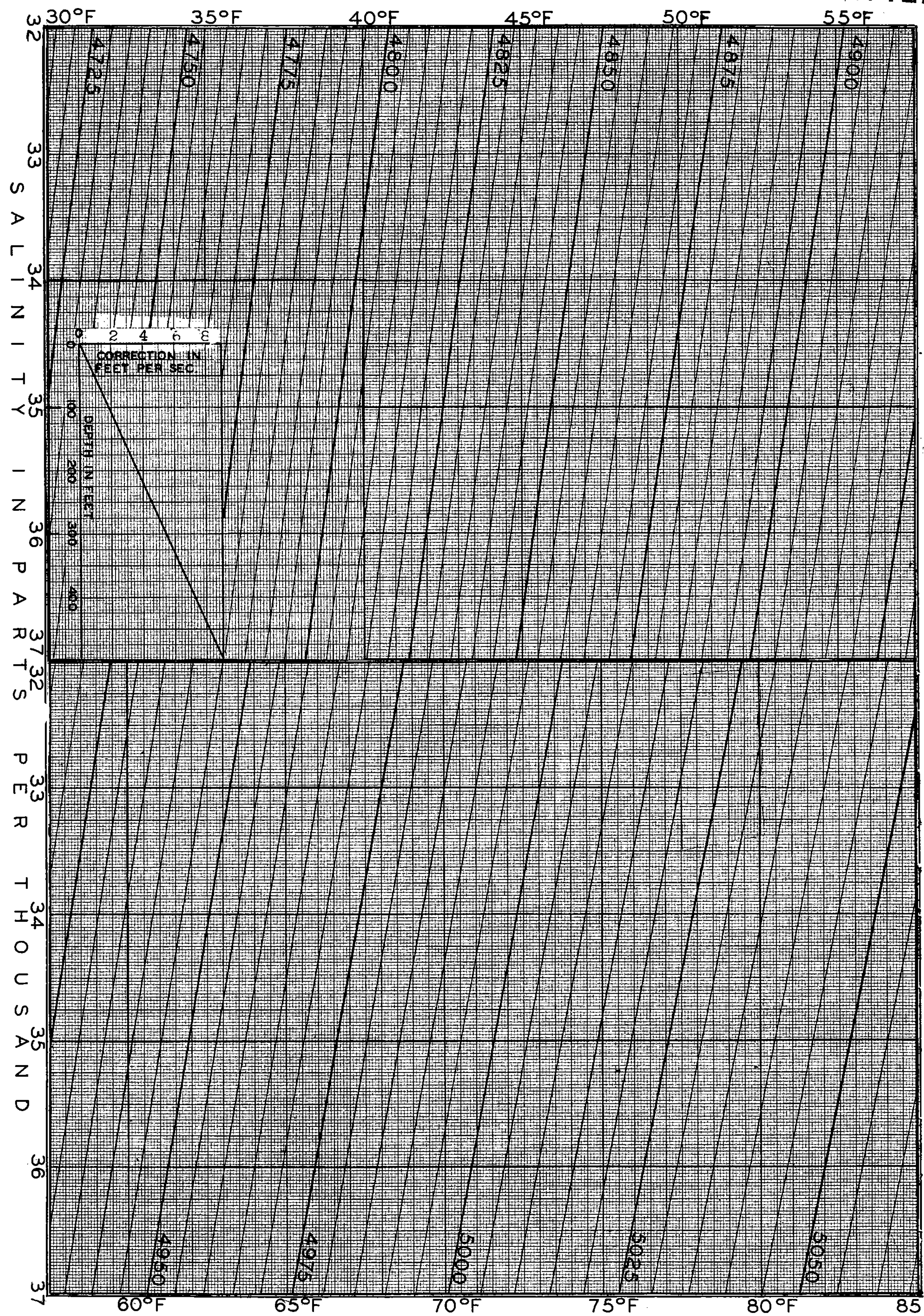

Fig. 1. SPEED OF SOUND IN SEA WATER

(b) Depth - 230 ft.
Temperature = 70.7°F
Salinity = 36.5 °/oo
Find the sound speed-

From Figure 1,	Temperature and salinity,	5001.1
" "	Pressure	4.1
	Sound speed	5005.2 ft/sec

(c) Temperature constant
Salinity constant
Find the vertical gradient of sound speed-

From Figure 2,	Temperature effect	0.0
" "	Salinity effect	0.0
	Pressure effect	0.0182
		0.0182 ft/sec/ft

(d) Temperature, 42.5°F at surface, 40.0°F at 65 ft.
Salinity, 32.3 °/oo at surface, 32.4 °/oo at 65 ft.
Find the vertical gradient of sound speed by assuming constant temperature and salinity gradients from surface to 65 ft.

Temperature gradient = 2.5/65 =-0.0384°F per ft.
From Figure 2, K at 41.3°F = 7.5
Gradient of sound speed due to temperature = 7.5x-.0384 =-0.288

Salinity gradient = 0.1/65 = 0.00154 °/oo per foot
Gradient of sound speed due to salinity = 0.0015 x 4.3 = 0.006

Gradient of sound speed due to pressure = 0.018
GRADIENT of sound speed (in ft. per sec per ft.) -0.264

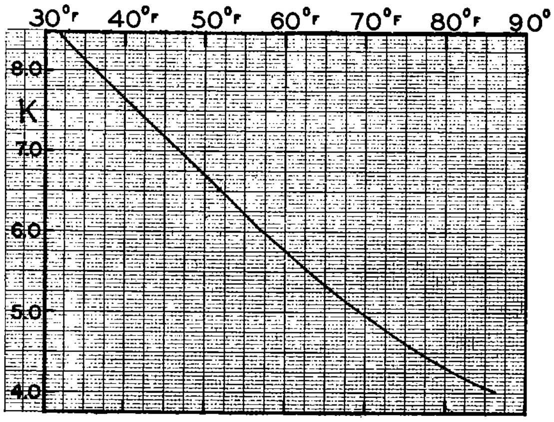

FIG. 2. GRADIENT OF THE SPEED OF SOUND. It is computed by the equation: 0.0182+(K x temp. grad.)+(4.3 x sal. grad.). The temperature gradient is in degrees Fahrenheit per foot and the salinity gradient is in parts per thousand per foot.

Measurement of Temperature and Salinity

Surface water. For the water at the surface a technique for the measurement of temperature with simple apparatus has been evolved which is well known and widely available. The salinity also could be determined for surface water to useful accuracy by means of a relatively simple apparatus. From these two measurements the speed of sound in the surface water could be derived. This value alone would have practically no value for predicting the performance of echo ranging apparatus because the variation of speed with depth is the controlling factor.

Nansen bottles and reversing thermometers. This is the standard apparatus for oceanographic research at the present time. The Nansen bottles with reversing thermometers attached are lowered to the desired depth by means of a wire. A "messenger" is then released at the surface and allowed to slide down the wire. The messenger causes the frame which holds the bottle and thermometer to invert itself, closing the water bottle and breaking the mercury thread in the thermometer in such a way that the reading of the thermometer will not change while the apparatus is being hauled to the surface. Thus a reading of the temperature and a sample of water for salinity determination are obtained from any desired depth. Usually two thermometers are attached to each bottle, one having its bulb protected and the other having its bulb exposed to the water pressure. The difference in the two thermometer readings indicates the depth, giving a more accurate value of depth than can be obtained from the length of wire paid out and its inclination at the surface. In actual practice ten to twelve water bottles are attached to the wire at suitable intervals, giving data for a number of depths for a single lowering. While this technique yields accurate observations, its drawback is that the ship must be stopped for half an hour or more each time readings are obtained.

The electrical resistance thermometer. This thermometer consists of a resistor whose change in resistance with temperature is accurately known. The resistor is lowered to the desired depth by means of an insulated cable and the resistance is measured on board ship by means of a suitable bridge circuit. The principal advantage of this type of thermometer is that the temperature can be read at many different depths during a single lowering. This type of thermometer has not been widely used for sea water temperatures, its principal disadvantage being that it is difficult to read the galvanometer at sea.

The bathythermograph is an instrument for obtaining a continuous record of temperature against depth. It consists of a pressure responsive element which moves a small glass slide parallel to the length of the instrument, and a temperature responsive element which moves a pen across the slide. The instrument is convenient to use and has given satisfactory performance to depths of 500 feet while operating from a vessel moving as fast as 14 knots. It seems to be the ideal instrument for detailed measurements in the first few hundred feet of water, its greatest advantages being that it measures temperature continuously and that it can be used while the ship is under way.

Regional And Seasonal Charts Of Water Conditions

The easiest way to find out how sound speed varies with depth at a given time and place would be to read it off a suitable chart. The need for such charts is new and the data necessary for their preparation are not plentiful because oceanographers have not concentrated their attention on the first few hundred feet of water. The sections and diagrams in Chapters IV and VI include most of the information which is available at the present time, and they can serve as a valuable guide. Additional data are being collected for use in the preparation of a more extensive set of charts. In Chapter V methods are given for determining the performance of underwater sound apparatus from these charts.

2. DENSITY OF SEA WATER. STABILITY

Factors Which Determine The Density

The density of sea water may be considered as the ratio of the weight of a given volume of sea water to the weight of an equal volume of distilled water at $4^{o}C$. The values of density observed at the surface fall between about 1.020 and 1.030.

The factors from which the density of sea water is determined are temperature, salinity, and pressure. The density increases with increasing salinity and pressure, but decreases with increasing temperature because water expands when heated. In the range of depths involved in echo ranging, the effect of pressure upon density is slight, but the effects of temperature and salinity are important and must be understood before one can grasp the principles governing variations of sound speed in the upper layers.

By the use of the graph in Figure 3 the density of sea water under surface pressure may be determined for given values of temperature and salinity. An auxiliary graph for allowing for the effect of changes in pressure is shown at the bottom of Figure 3.

Calculation Of Density Of Sea Water

The following calculation of the density of the water is given for the temperatures and salinities observed at a station in coastal water in 41 fathoms off Block Island, April 13, 1938. The temperature decreases down to 20 meters but increases from there to bottom, becoming higher at the bottom than at the surface. The salinity increases all the way to the bottom, its effect upon density more than offsetting the effect of the rise in temperature of the lower part of the water column. This example confirms the general rule that density increases all the way to the bottom.

Depth		Temp.		Salinity	Density	
meters	fathoms	^{o}C	^{o}F	o/oo	at surface pressure	at actual pressure
1	0.55	5.84	42.5	32.39	1.02554	1.02554
10	5.47	5.09	41.2	32.31	1.02556	1.02560
20	10.94	4.46	40.0	32.47	1.02575	1.02585
30	16.40	4.55	40.2	32.56	1.02581	1.02596
40	21.87	4.85	40.7	32.74	1.02592	1.02611
50	27.34	6.35	43.4	33.39	1.02626	1.02650
60	32.81	6.59	43.9	33.49	1.02631	1.02660

The densities are calculated in two ways: first the potential density, which is the density the sample would have if brought to the surface, and second the actual density, which is the density of the sample in place. In much oceanographic work the potential density is the quantity considered, because it indicates directly whether the water from one level would be heavier or lighter than its surroundings if it were transported to another level.

This discussion of density has not been introduced for use in calculating the speed of sound because the equation $\text{velocity} = \sqrt{\text{elasticity} \div \text{density}}$ is not convenient for use in this calculation. It has been included to enable the reader to understand stability of the water column, the vertical circulation which occurs when the stability is destroyed, and the vertical distribution of temperature to which this circulation can lead.

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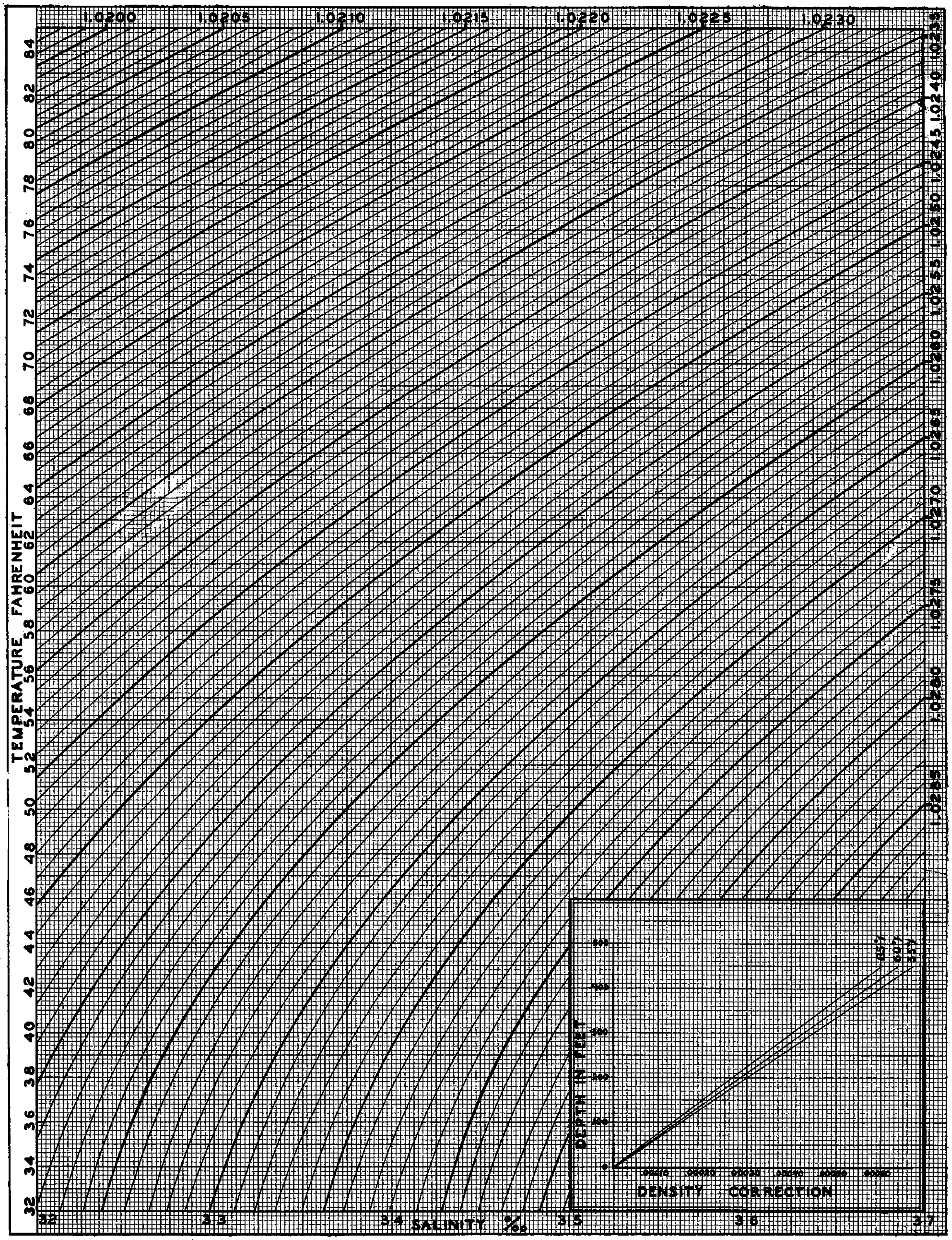

Fig. 3. DENSITY OF THE SEA WATER

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Stability

The densities calculated in the above example illustrate the stability of the water column. In other words the lighter water is at the top and the heavier water is at the bottom. Sea water is such a free-flowing fluid that it always adjusts itself in this way. The degree of stability is measured by the rapidity with which the potential density of the water increases with depth, and the more stable the column becomes the greater is its ability to resist vertical mixing by the action of the wind.

If a water-column is warmed at the top by the sun or by warm air, the surface density is diminished and the stability is thereby increased. If the top of the column is cooled, the density of that part increases and the stability decreases. If the process of cooling is continued until the stability becomes slightly negative, convection begins, the water from the surface sinking until it reaches a level where it finds denser water beneath it. This action produces a thoroughly mixed surface layer which is very favorable for echo ranging provided it extends sufficiently deep. As is shown in the oceanographic diagrams of Chapter IV, Figure 13, for example, the homogeneous layer becomes deeper and more suitable for echo ranging in winter than in summer.

3. REFRACTION OF SOUND WAVES

Refraction Described

Sound waves travel in straight lines only when in a medium in which the speed is everywhere constant. If there existed a large body of water in which the speed of sound was the same at every point, the beam from a sound projector would be a cone with its axis perpendicular to the face of the projector. In an actual body of water the speed of sound will vary with depth. Suppose for example the speed increases with depth. In that case every ray of the sound beam will be curved toward the surface. The more rapid the change of speed with depth, the more strongly the rays will be curved. This bending of the sound rays is called refraction. It will aid in understanding refraction to imagine that the lower part of the beam, in the example just cited, keeps getting slightly ahead of the upper part, due to the slight difference in speed in the water at the levels at which the two parts of the beam travel. It is a fundamental law of wave motion that the rays, which indicate the direction of travel, are always perpendicular to the wave fronts. A wave front is the surface occupied by the front of a sound signal at any instant. Because the lower part of the wave front keeps gaining on the upper, the beam will curve upward.

If the speed had been decreasing with depth, the beam would have bent downward. Refraction always causes a sound ray to shift its course toward the water in which the speed is lower.

Snell's Law Of Refraction

Snell's Law is used in the study of optics to calculate the bending of light rays when they pass from one transparent substance to another. The bending depends on differences in the speed of light in the two substances. Snell's Law may also be applied in calculating the bending of sound rays when they pass through successive layers of water in which the speeds are different. Suppose the speed in the water depends only on depth and not on horizontal position, as is usually the case. Snell's Law, when applied to a sound ray such as that shown in Figure 4, states the following:

$$V_o/\cos i_o = V_1/\cos i_1 = V_2/\cos i_2 = V_3/\cos i_3 = V_m \quad \ldots (1)$$

where i is the inclination of the ray to the horizontal at any point, and V is the speed of sound in water at the same point. If V_2 is greater than V_1, equation (1) shows that $\cos i_2$ must be greater than $\cos i_1$, or i_2 must be less than i_1. If V_2 is also at a greater depth than V_1, the ray is bent upward. V_m represents the speed at the point where the ray becomes horizontal, and it can be seen from equation (1) that this is the greatest speed at any point on the entire path. The point at which a ray becomes horizontal is called the vertex of the ray. For the rays shown in Figure 6, the vertices are at the points G, F, P, E, C and B.

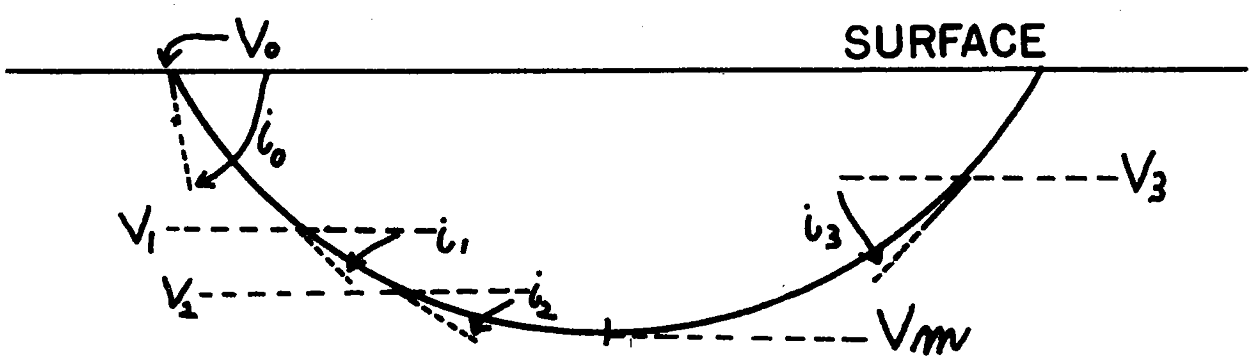

Fig. 4. SNELL'S LAW OF REFRACTION. $V_o/\cos i_o = V_1/\cos i_1 = V_2/\cos i_2 = V_3/\cos i_3 = V_m$

Theory of Circular Sound Paths

General statement. In any layer of water in which the vertical velocity gradient is constant (see definition, page 9) it can be proved by Snell's Law that the paths of all sound rays are circles. The centers of all the circles for a given layer lie on a horizontal line whose distance from the surface of the layer is V_o/g, where V_o is the speed of sound at the surface of the layer and g is the vertical gradient. It may be of aid to the memory to think of the line of centers as the elevation at which the speed in the layer would become zero if the layer is imagined to extend that far and to have the same gradient throughout.

Positive gradient. A schematic diagram showing the paths of the sound rays in a layer where the vertical velocity gradient is positive appears in Figure 5. It is seen that the radius of any one of the circular paths is given by

$$R = V_o/(g \cos \theta_o) \quad \ldots (2)$$

where R is the radius of the path and θ is the inclination of the ray to the horizontal at the surface of the given layer.

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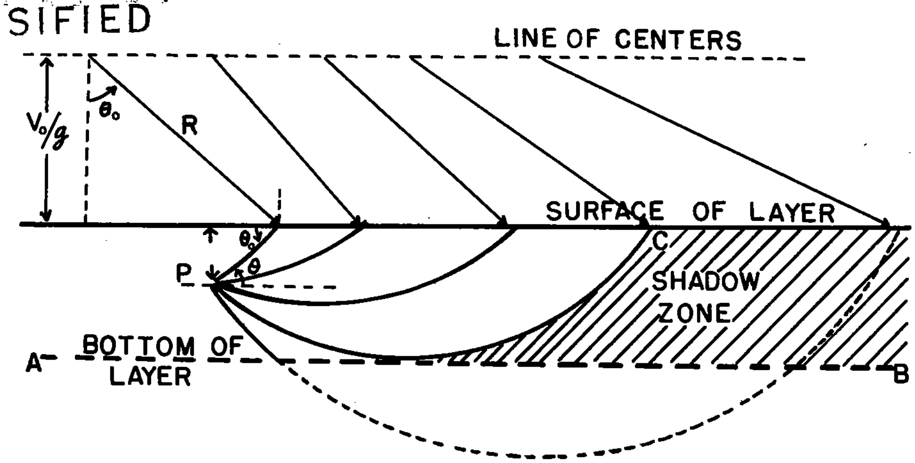

Fig. 5. REFRACTION WITH POSITIVE GRADIENT. The rays are all circles with centers Vo/g above the surface of the layer. A shadow zone for direct rays is formed by the bottom of the layer.

It can also be seen from Figure 5 that if the layer is of limited thickness, terminating for example at the line AB, the direct rays for all ranges beyond the point marked C will suffer interference. The type of interference suffered will depend on the nature of the underlying layer. If the layer beneath AB has a negative vertical gradient of sound speed so that all sound rays which enter it are curved downward, no direct rays will reach the surface at ranges beyond the point C, producing a shadow zone as indicated in the figure.

In all schematic sound ray diagrams in this report (Figs. 5, 6, 7 and 11) the gradient has been exaggerated for the sake of graphical convenience. For example in Figure 5 the projector depth is actually about 15 feet while the radius of one of the circles is about 50 miles. This simply means that the gradient used in the drawing was enormously larger than those which actually occur.

When constructing the ray diagrams for actual water conditions it is necessary to make the vertical scale of the section much greater than the horizontal one. This distorts the circular path into ellipses, and they cannot be drawn with compasses but must be constructed from the graphs in Figures 29 and 30.

Negative gradient. In case there is a negative gradient, which remains constant throughout a given layer, the paths of the sound rays will be circles with centers below the layer as shown in Figure 6. The direct rays for all ranges beyond the ray marked C are cut off by the surface of the water, giving rise to a shadow zone as indicated.

Inspection of the diagrams indicates that direct rays reflected from the surface will be far more effective in carrying sound into the shadow zone in case of a positive gradient in the upper layer (Fig. 5) than in the case of a negative gradient in the upper layer (Fig. 6).

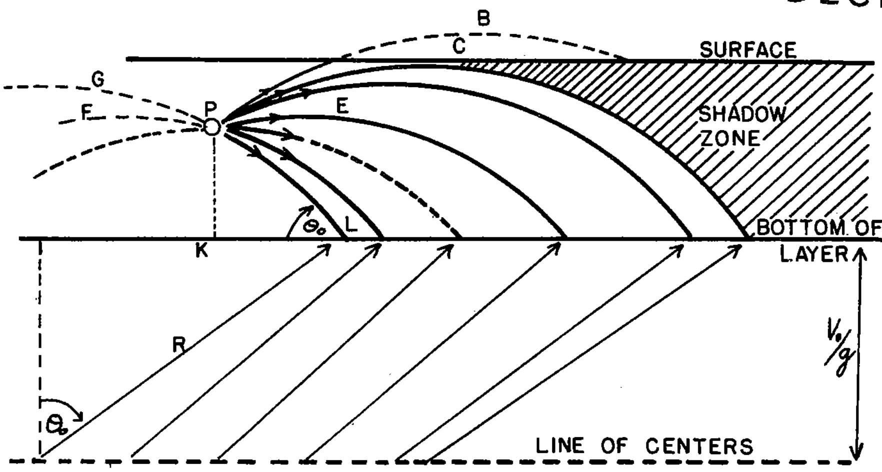

Fig. 6. REFRACTION WITH NEGATIVE GRADIENT. The circular rays are concave downward, and a shadow zone for direct rays is formed by the upper surface.

Negative gradient above a positive gradient. In Figure 7 schematic diagrams for the sound rays are drawn for the case where the top layer of water has a negative velocity gradient while the second layer has a positive one. The variation of speed with depth is shown in the small graph at the left side of the figure. All rays from the projector are curved downward at the start, and the top portion of the beam is cut off by the surface, causing a shadow zone starting at the range A. The beam which enters the second layer is limited at point B by the angle of the cone of the projector, and at the point C by the "shadow" of the surface at A. The rays which enter the second layer are all bent upward. Those which enter at sufficiently small inclination are curved back and reenter the first layer, but those which enter with too much inclination reach the bottom of the second layer before recurving completely. The axis of the beam leaves the projector horizontally, enters the second layer with minimum inclination and hence recurves to reenter the first layer at the minimum distance (point E). The part of the direct beam which reenters the first layer is limited on its two sides by the shadows of the bottom of the second layer at F and G.

From Figure 7 it can be seen that, for ranges beginning near E and extending for a considerable interval, a sound detector placed at the same depth as the projector P would be in the direct beam, although the detector would have been out of the beam over a considerable interval of lesser ranges. This situation explains satisfactorily many cases where the signal strength first falls and subsequently rises as the distance is increased. Several actual cases of this type have been observed and are described in a later section of this report. These cases are fully explained by consideration of the distribution of temperature and salinity in the water.

Natural "channels" for sound conduction. The graph at the left side of Figure 7 indicates the speed of sound versus depth. The dotted line HK sets off a portion of the water which acts as a natural channel for conducting sound. The velocity at H is equal to that at K, and is greater at these points than at any point between them. Any sound ray (for instance the axis of the beam) which becomes horizontal inside the region HK will be refracted back and forth across the level of minimum velocity, never leaving the channel. From Snell's Law of Refraction (Equation 1, page 16) it is known that the point at which any given ray becomes horizontal is the point of highest speed on that particular path, so the only rays which can become horizontal in the channel HK are rays which originate from a projector inside the channel. Any ray which originates outside the channel must obviously enter the channel with an inclination greater than zero if it enters the channel at all. A ray always increases its inclination when it goes into water having lesser sound speed, and the speed inside the channel is lower than at the boundaries H and K, so the inclination of the ray is greater inside the channel than at the point of entry.

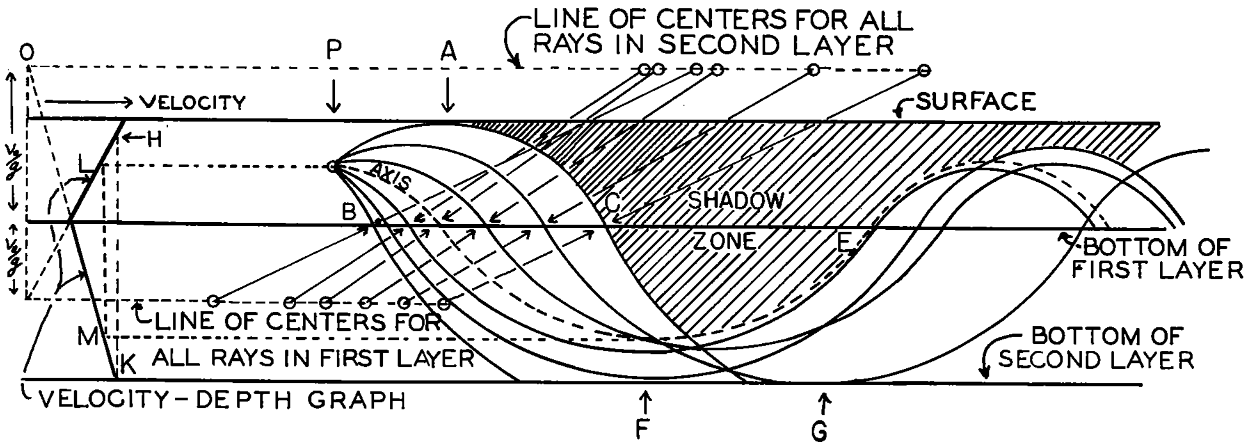

Fig. 7. NEGATIVE ABOVE POSITIVE GRADIENT. This case frequently occurs when the upper layer gains stability through diurnal heating. It is the basis of the "afternoon effect".

Thus a channel, as defined above, serves as a good conductor for sounds which originate inside it, but not for sounds which originate outside. From Snell's Law it may be seen that if a ray enters a sound channel through the top side it will pass entirely through the channel, having the same inclination at the bottom that it had at the top, because the velocities at top and bottom are equal. These channels are responsible for "skip distances" as, for example, in Figures 34-38. The sound channel HK in Figure 7 may be subdivided by considering separately the part LM, which constitutes a sound channel with the projector outside. As may be seen from the drawing and from the preceding discussion, no ray has its vertex inside the channel LM, and the only rays which recurve toward the surface are those which originally have vertices in HL. The zones HL and MK together form a sound channel with the projector inside it. They are responsible for returning the sound to the upper layer beyond the range E.

Reversibility Of Sound Paths

The refraction of sound waves, in all cases discussed so far, depends only on the velocity distribution in the water. Such refraction is governed completely by Snell's Law, and it is important to know that all paths computed for sound rays by this law are completely reversible. If sound proceeds from a projector to a target, undergoing refraction on the way, it can return to the projector along the same path. Furthermore the principle of reversibility of paths leads at once to the conclusion that if a target is in a shadow zone relative to a given projector then the projector is in a shadow zone relative to the target. In other words, if refraction conditions prevent a destroyer from obtaining echoes from a given submarine, these refraction conditions will also prevent that submarine from obtaining echoes from the destroyer. (This statement makes no allowance for the fact that the destroyer may be a more efficient reflector than the submarine).

Focussing Effect

By reference to Figure 7 it can be seen that refraction sometimes causes the sound beam to be concentrated into a relatively small cross section at certain distances. For instance the beam in the region between points F and G is more compressed than at slightly greater distances. This corresponds to the action of an imperfect lens and undoubtedly affects the intensity of echoes. In comparison with other effects this one is relatively minor, and at the present stage it will be ignored except in special cases.

Convective Refraction

The refraction of sound waves in water can take place in two ways, either by the variation of the speed of sound in the water or by the effect of currents. The effect of currents, which is known as convective refraction, may be compared to the effect of wind on the propagation of sound in air. In general the speed of the wind is least near the surface and gradually increases with height up to such heights that the retarding effect of surface friction becomes negligible. If a sonic beam is transmitted downwind speed of the wind is added to the speed which the sound would have if the air was at rest, so the effective velocity of sound increases with height. This curves the sound rays toward the surface and permits reception of the sound by a detector at the surface at a great distance. For transmission to windward the speed of the wind is subtracted from the true speed of sound in air, resulting in a decrease in the effective speed of sound with increasing height. This curves the sound rays upward and causes them to pass above a receiver located at the surface, reducing seriously the range at which signaling is possible. The important way in which convective refraction differs from velocity refraction is that it is not reversible, for the wind aids transmission downwind but hinders it in the reverse direction.

There are two situations in which this type of refraction may occur in water, but it will never be as serious in water as in air because the speed of flow is always a much smaller fraction of the speed of sound in water than it is in air. In wind driven currents the velocity is maximum at the surface and decreases downward. A sound beam travelling to windward would be curved toward the surface by convective refraction, while one travelling to leeward would be bent away from the surface. At the present time there are not enough data available on the velocities of wind driven currents, their variation with depth, and the time lag between the action of the wind and the establishment of the current to permit evaluation of the importance of convective refraction in underwater sound work. About the only positive statement which can be made is that the wind drift will be confined to a shallower surface layer in stable water columns than in unstable ones.

Near the boundaries of the major oceanic currents there are obviously lateral and vertical variations in currents capable of causing convective refraction. At present there are so many other complications in these regions that effects due to convective refraction have not been isolated.

Difference between upwind and downwind performance would not be noted in echo ranging because the distance between target and projector is always traversed by the sound beam in both forward and backward directions. If the convective refraction aids transmission of the projected beam it will impair the transmission of the reflected beam, and vice versa. Difference between upwind and downwind performance may be expected however in all underwater sound signaling where the beam traverses the path in only one direction. Whether this effect is large enough to be given serious consideration is not known at present.

4. SUMMARY

1. Variations in the speed of sound, caused by variations in the temperature and salinity of the sea water, are the controlling factors in determining the horizontal range of sonic apparatus. The effect of these variations is to deflect the sonic beam in a vertical plane so that in some cases it misses the target. In the open ocean the vertical salinity gradient is small and the speed of sound is therefore determined primarily by the temperature. In case the temperature is constant in the upper layer, the slight increase in speed due to the increase in water pressure with depth becomes important. By means of graphs, Figure 1 and 2, the speed of sound and the rate of change of speed with depth may be calculated if the temperature and salinity are known.

The most practical instrument for measuring the temperature of the water to depths of about 75 fathoms is the bathythermograph, for it can be used from a moving ship.

2. By means of Figure 3 the density of the water may be computed from temperature, salinity and depth. The distinction is made between potential density and actual density. The densities are calculated at a typical station in coastal water to show that the bottom water is more dense than that above it, even though its temperature is higher. Heating the water at the surface produces stability of the water column, while cooling it destroys stability and leads to the formation of a surface layer of thoroughly mixed water.

3. The paths of sound waves are bent into curves when there are vertical gradients in the water. This process is known as refraction, and the amount of bending may be calculated by use of Snell's Law. In any layer of water in which the vertical velocity

gradient has a constant value, all sound rays are circles having centers a fixed distance from the surface of the layer. Schematic ray diagrams showing the effects of refraction in typical cases are discussed.

CHAPTER IV

OCEANOGRAPHY OF THE SURFACE LAYER

1. INTRODUCTION

Physical oceanography is the subdivision of the earth sciences which deals with the structure and circulation of the ocean. In the development of this rather obscure field of knowledge the United States has played a leading part, but since about 1900 considerable research has also been published from the northern European countries. At the present time the subsurface distribution of temperature and salinity (or saltiness) has been well examined only in the North and South Atlantic, where observations have been secured at various depths between surface and bottom at as many as 6000 stations. In the Pacific and Indian Oceans much less detail is available except along the west coast of the United States, in the Dutch East Indies and near Japan. It is for this reason that the diagrams which follow have largely been constructed from observations made in the North Atlantic.

In spite of the lack of data from the central areas of the other oceans, it is possible to learn a good deal about the sound transmitting qualities of their waters through a study of the North Atlantic. The basic pattern of the currents and the seasonal variations of temperature in the superficial layer varies but little from ocean to ocean. The North Pacific and the North Atlantic are particularly similar in these respects, while the South Atlantic, South Pacific and Indian Oceans are also closely comparable. Thus, provided we compare geographically analogous parts of each of these oceans, a knowledge of the sound transmitting qualities of the waters in the Atlantic will enable us to predict the general conditions elsewhere. For example, the waters near Japan are very similar in structure to those off the eastern coast of the United States, while conditions in the eastern part of the North Pacific resemble those off the European coast. These important considerations will be more fully discussed in Chapter VI.

The main ocean basins average some 2500 fathoms in depth, but near the continents is usually found a shallow (less than 100 fathoms deep) shelf of varying width. This is known as the continental shelf and over it the vertical structure of the water and the causes of the currents are somewhat different than in deep water. North of Cape Hatteras the continental shelf bordering the eastern coast of North America is broad (40 to 200 miles wide) and well developed. It is usually narrow off mountainous coasts such as the coast of California.

2. THE BASIC VERTICAL STRUCTURE OF THE OCEANS

With very few exceptions the deep oceanic waters can be subdivided into three principal layers: 1) a relatively warm surface layer which is subject to daily and seasonal changes in temperature and salinity, and to wind stirring; 2) a layer of transition at mid-depths, the main thermocline layer, in which temperature decreases rapidly with depth; and 3) the much colder deep water layer in which temperature decreases only gradually with depth. This basic structure is illustrated in Figure 8; however, the relative thickness of the three layers varies from one part of the sea to another. In general, the surface layer and the layer of transition at mid-depths are relatively deep in mid-latitudes (A) and shallow near the equator (B), but in high latitudes there is a less pronounced vertical thermal structure (C). Indeed in midwinter in very high latitudes the water may be completely mixed, surface to bottom.

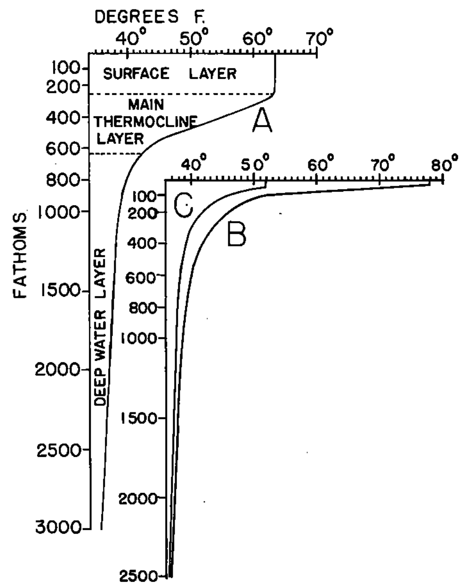

Fig. 8. Typical temperature-depth curves from the deep ocean. Curve A is representative of winter conditions in mid-latitudes, curve B of the equatorial belt and curve C of high latitudes in summer. The depth scale is in fathoms and the temperature scale in degrees Fahrenheit.

In this report we will be dealing mainly with the structure of the surface layer, for only where the transition layer at mid-depths is especially shallow and sharp does it enter the sound ranging problem. The thick layer of the relatively cold, deep water, which in the main ocean basins constitutes from 70 to 85% of the water column, can be completely neglected as a factor in horizontal sound ranging. On the other hand, inside the 100 fathom curve, as will be shown below, the relatively cold but saline bottom water often does play a part.

In the deep ocean, as has already been explained, the temperature decreases with depth, although not at the same rate in each of the three primary layers. In deep water the concentration of dissolved salts, which is customarily expressed in parts per thousand (°/oo), also in general decreases with depth and in a manner quite parallel with temperature (Fig. 9A). This is in contrast to typical coastal waters on a broad continental shelf (Fig. 9B). Especially in mid-latitudes and off the eastern coast of the continents a layer of minimum temperature is usually encountered at mid-depths during the spring and summer. But more important, over the continental shelf the salinity (for definition see page 6) increases with depth. This contrast results from the fact that in the open ocean the basic structure of the water layers is the result of the currents, while near the land it is the distribution of temperature and especially salinity which is often their cause. In the deep ocean the energy for the currents comes mainly from the general wind system, while along the coast the currents result also from horizontal density gradients maintained by the inflow of fresh water from the land. This important distinction will perhaps become clearer as the discussion proceeds.

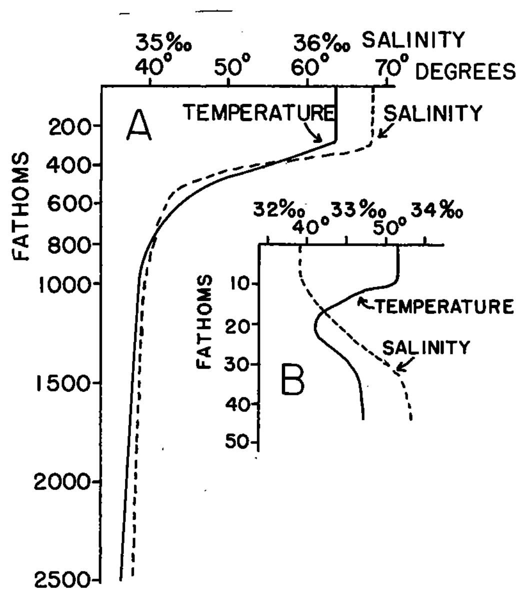

Fig. 9. Comparison between typical temperature-depth curves and salinity-depth curves. Curves A are based on observations made in winter near Bermuda, while curves B are from coastal water off southern New England in early spring. The salinity scale is in parts per thousand by weight.

Thus far we have stressed the vertical changes of temperature and salinity in the ocean, for these are quite tangible characteristics of water. However, it is the <u>density</u> (for definition see page 11) distribution resulting from the thermal structure and from the distribution of dissolved salts which is of fundamental significance in oceanic circulation. Warm waters are of course less dense than cold waters and fresher waters are less dense than more saline waters. The relationship between temperature, salinity and density over the range normally found in the sea is illustrated in Figure 3, Chapter III.

Obviously it is to be expected that density will increase with depth nearly everywhere in the sea, in which case the water is said to be <u>stable</u>. Unstable conditions, where heavier water overlies lighter water, are very rarely encountered, except near the surface during the winter months when slight instability, resulting in excellent sound ranging conditions, may at times prevail.

In the deep ocean the vertical distribution of density is very closely controlled by temperature, in fact from the point of view of salinity alone the water-column is unstable. On the other hand, over the continental shelf both temperature and salinity contribute to the stability, which consequently is often very great. This is illustrated in Figure 10 which shows the density distribution derived from the two pairs of temperature and salinity curves given in Figure 9. As will be explained below, the curves for the coastal water (B) can only be considered typical of the spring months, for outside the tropics the whole water-column over the continental shelf is usually subject to seasonal variations in temperature and salinity.

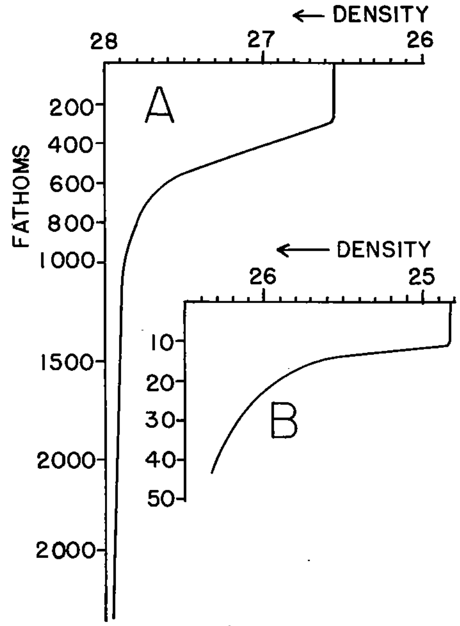

Fig. 10. Typical "density-depth" curves derived from the temperature and salinity observations plotted in Figure 2. As is customary in oceanography the pressure effect has not been included in computing the values expressing density.

The slope of the density-depth curve is a measure of stability. Where the curve is nearly vertical the water has little stability and at depths where it approaches the horizontal the water is very stable. It will be noticed in the examples given that the coastal water (B) is very much more stable at mid-depths than the oceanic water (A). For example, at a depth of 12 fathoms in the coastal water the density increases with depth about 80 times as rapidly as in the most stable part of the offshore water-column, and this contrast would be even more marked if midsummer observations from coastal water had been used.

Stability (the rate of change of density with depth) is an important factor in the sound transmission problem for it is this characteristic of the water which resists the downward penetration of wind stirring. Were the waters less stable, even a moderate wind would be able to maintain a relatively deep homogeneous layer at the surface, a condition favorable to the horizontal transmission of sound. Especially over the continental shelf and in summer the stability is so high that only a very shallow layer of wind stirred water is found at the surface.

As already explained, in the open ocean the density-depth curve is closely parallel to the temperature-depth curve and largely dependent on it. In addition, except near land, the vertical range of salinity in the surface layer is relatively small. Consequently in oceanic areas it is usually permissible as a first approximation to estimate stability and density from the temperature distribution. On the other hand, in coastal waters, where density throughout most of the year is largely controlled by the salinity-depth distribution, temperature observations alone can give an incorrect picture of stability (compare Figure 9B with Figure 10B).

3. EFFECT OF TEMPERATURE, SALINITY AND PRESSURE ON SOUND

The speed of sound in sea water, as has been explained in detail in Chapter III, varies with the temperature, salinity and pressure. The higher the temperature, salinity, or pressure, the greater will be the speed of sound. Since the magnitude of the temperature effect is considerably greater than that of the other two factors and since temperature ordinarily decreases with depth, in most areas the sound velocity will do the same. The pressure is of course a function of depth, and therefore in thoroughly mixed water the speed of sound will increase with depth. Because allowance for the effect of pressure can be made by calculation and because variations in salinity are relatively small, in most parts of the sea it can be considered that the speed of sound is approximately known if the temperature distribution has been observed.

In interpreting the oceanographic diagrams which follow in terms of sound transmission, it is also necessary to realize that as sound travels in water in which there are vertical velocity gradients the beam is bent slightly in the direction of the lower velocities (for particulars see Chapter V). Thus in water in which the speed of sound decreases with depth, a sound beam which starts out horizontally will be deflected downward. This is the case in all layers where stability exists due to a decrease of temperature with depth. On the other hand, where the water is homogeneous, as in the wind stirred layer at the surface, the pressure effect will cause an increase in speed with depth and therefore a sound beam which starts out horizontally in this layer will be bent slightly towards the surface. From this it follows that the horizontal range of sound in the superficial waters is much greater in mixed water than in layers where the temperature or salinity decreases rapidly with depth. Good sound water is water which is thoroughly mixed and thus has no stability. Poor sound water is water which is thermally stable or in which salinity decreases rapidly with depth.

As the discussion proceeds the importance of these considerations will become more apparent, but before going into more detail it is advisable further to examine the basic structure of the ocean.

4. THE BASIC HORIZONTAL STRUCTURE OF THE OCEANS

Horizontally the water in general changes much more gradually in temperature and salinity (and therefore in density) than it does vertically, yet the horizontal structure can by no means be neglected when one examines the ocean from the point of view of sound ranging.

The density of the surface water is least near the equator in the doldrum belt, because of high rainfall and high temperature, and increases quite rapidly both north and south to the horse latitude belts (roughly Lat. 30°) where the temperature is still quite high, but where the salinity is at a maximum because of rapid evaporation (caused by the relatively dry air) and small rainfall. With increasing latitude the density continues to rise, because of lower temperatures and in spite of increased rainfall. The broad features of the changes of temperature, salinity and density with latitude in mid-

Atlantic are shown for the surface in Figure 11A. As will be discussed below, near the borders of strong currents and near the coast the horizontal gradient of temperature and salinity in the surface layer is sometimes so abrupt as to interfere with sound ranging.

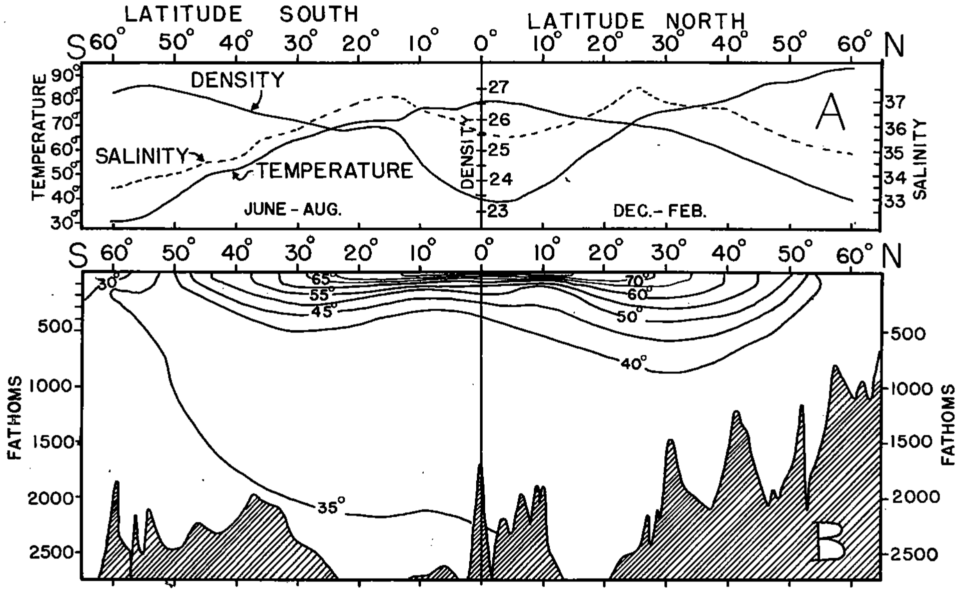

Fig. 11. The curves marked A show the changes with latitude of temperature, salinity and density at the surface in mid-Atlantic. In the lower part of the diagram (B) the temperature-depth distribution along the same profile is shown, with the vertical scale of course much exaggerated.

In Figure 11B the basic structure of the deeper layers of the Atlantic is given in north-south profile. This diagram is introduced to show that near the equator and at about Latitude 50° the permanent thermocline layer is near enough to the surface so that it can affect horizontal sonic signaling results. The same is in general true in the western part of the oceans. For example, along the northern edge of the Gulf Stream and the Japan Current the main thermocline is particularly sharp and shallow. This basic horizontal structure of the oceans, with a relatively deep surface layer in mid-latitudes and in mid-ocean, is the result of the major current systems driven by the anticyclonic winds. North of the equator the main currents circulate clockwise with an axis at about Latitude 30°, while south of the equator corresponding great eddies rotate in an anticlockwise direction. At the center of these great swirls, for example near Bermuda, the main thermocline is relatively deep and gradual, while all around the edge of the currents, especially in the west where they are swift, the main thermocline is shallow and sharp. From this it follows that in general sound ranging conditions are better in mid-ocean and in mid-latitudes than elsewhere.

5. THE SEASONAL THERMAL CYCLE OF THE SURFACE LAYER

Neither the deep mass of colder water nor the permanent thermocline layer is affected by the local climate. However, above the main thermocline there is everywhere a more or less marked seasonal cycle of temperature which is of major importance to sound ranging. Where the surface layer is deep, as in mid-ocean, these seasonal thermal changes extend down to 150 fathoms or more. But near the equator, where the surface layer is shallow, the seasonal temperature change is slight and unimportant to sound ranging compared with the permanent thermocline.

In Figure 8, curve A was constructed from observations made during the winter months when the surface temperature is at a minimum and when the depth of wind stirring is at a maximum. In midsummer, due to the absorption of heat near the surface a secondary thermocline develops at depths between 10 and 50 fathoms so that temperature-depth curves have the form shown in Figure 12A. This great increase in thermal stability so near the surface (comparable to the stability of coastal water in summer) results in general in poor sound ranging conditions during the spring and summer months, for, as was explained in Chapter III, whenever the temperature decreases with depth at a rate in excess of 1°F in 40-60 fathoms sound is deflected downwards.

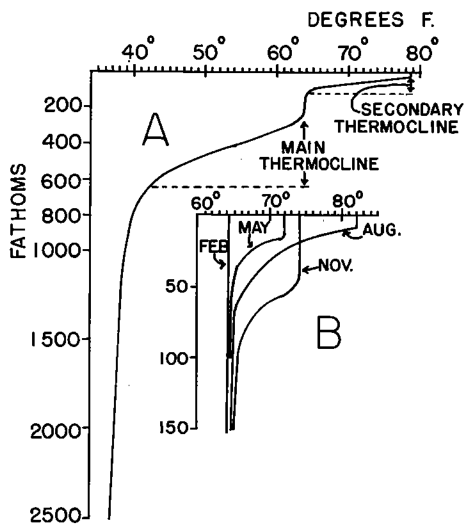

Fig.12. The curve marked A shows the temperature-depth distribution in midsummer near Bermuda. Curves B, with an increased depth scale, show the thermal situation in the surface layer at four different times of year.

The extreme lower limit of the seasonal temperature cycle varies from one part of the sea to another, and as might be expected the total range of the cycle varies with latitude, being of course small near the equator and large in high latitudes. However, there are quite large areas in which the seasonal thermal changes in the surface layer are rather uniform. For example, the whole Sargasso Sea with Bermuda near its center exhibits much the same pattern of seasonal temperature changes in the upper layers. In Figure 12B a number of temperature-depth curves from various months are shown for the upper 150 fathoms in this area. As already explained, as a first approximation these temperature-depth curves can be interpreted in terms of density and therefore of stability. Starting in late February, when surface temperatures are at a minimum and when wind stirring extends down to 150 fathoms or more, there is no stability above the permanent thermocline. Then as spring commences the upper 20 fathoms or so begin to warm up and a secondary thermocline is established. As soon as this is formed stability develops at depths vital for sound ranging. As more heat is absorbed the secondary thermocline increases in intensity and depth, for the heat is also mixed downward gradually by turbulence. By summer only 5 to 7 fathoms of mixed water may overlie the very stable layer. During the autumn, when the surface begins to lose heat, the slight tendency for instability at the surface greatly increases the effectiveness of wind stirring and the thickness of the homogeneous water at the surface increases rapidly. By mid-October as much as 20 fathoms of mixed water will be found at the surface and sound ranging conditions rapidly improve thereafter.

This sort of seasonal gain and loss of heat can be illustrated more clearly by another form of diagram (Fig.13) in which isotherms (lines of equal temperature) are plotted with depth and time as coordinates. In such a diagram (assuming that the salinity gradients are slight and therefore that temperature can be interpreted in terms of sound transmission) where the isotherms are vertical there is no stability and the pressure effect will curve the sound beam slightly toward the surface; where they have a horizontal component the water is stable and sound rays which enter this layer will be bent sharply downward. Thus, provided the vertical salinity gradients are slight, the broken curve on this diagram indicates the lower limit of good sound water in the Sargasso Sea area month by month. It is hoped that before long observations will be available to construct similar diagrams for many other areas.

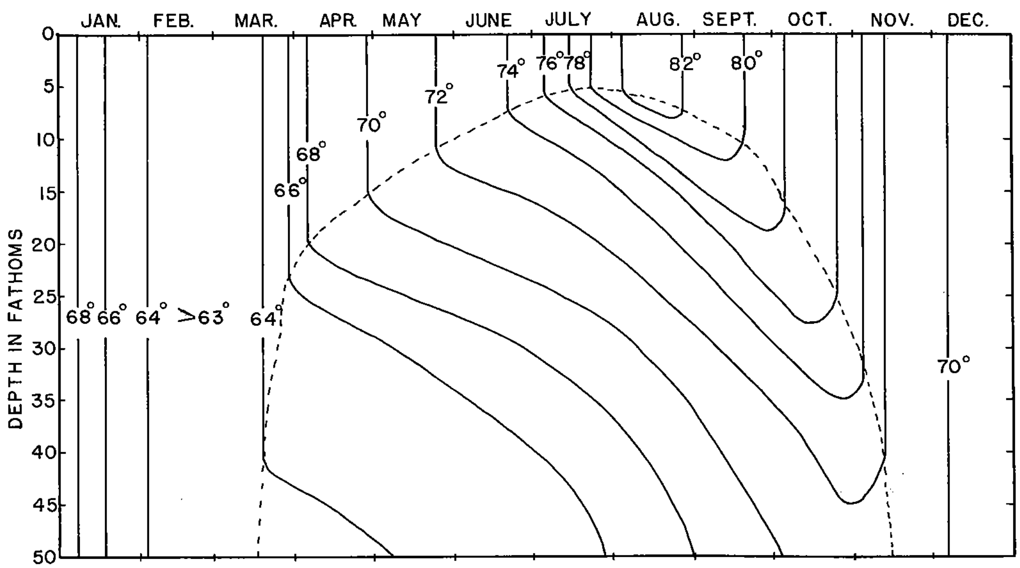

Fig.13. Seasonal cycle of temperature at depths of less than 50 fathoms in the western Sargasso Sea, based on frequently repeated observations obtained at approximately Lat. 35°N, Long. 67°W. The broken curve shows the lower limit of wind stirring during the part of the year when a secondary thermocline is present.

In offshore waters the vertical salinity gradients are indeed relatively small. For example, if a diagram similar to Figure 13 were constructed to show the seasonal cycle of salinity in the upper 50 fathoms of the Sargasso Sea, nowhere on this diagram would the salinity change with depth at a rate in excess of .025 °/oo per fathom. Moreover, this extreme gradient would be located in the secondary thermocline layer where its effect on sound speed is completely masked by the vertical temperature gradient.

To illustrate how closely a knowledge of the seasonal temperature cycle enables one to estimate vertical gradients of sound speed in the open ocean, the temperature observations used in constructing Figure 13 and the corresponding salinity determinations have been used to derive a seasonal sound speed diagram (Fig. 14). This shows the seasonal variations in sound velocity at depths of less than 50 fathoms in the Sargasso Sea. The observations came from a point about 100 miles northwest of Bermuda (see Fig. 20), but, as will be explained below, the conditions there are typical of a very wide area.

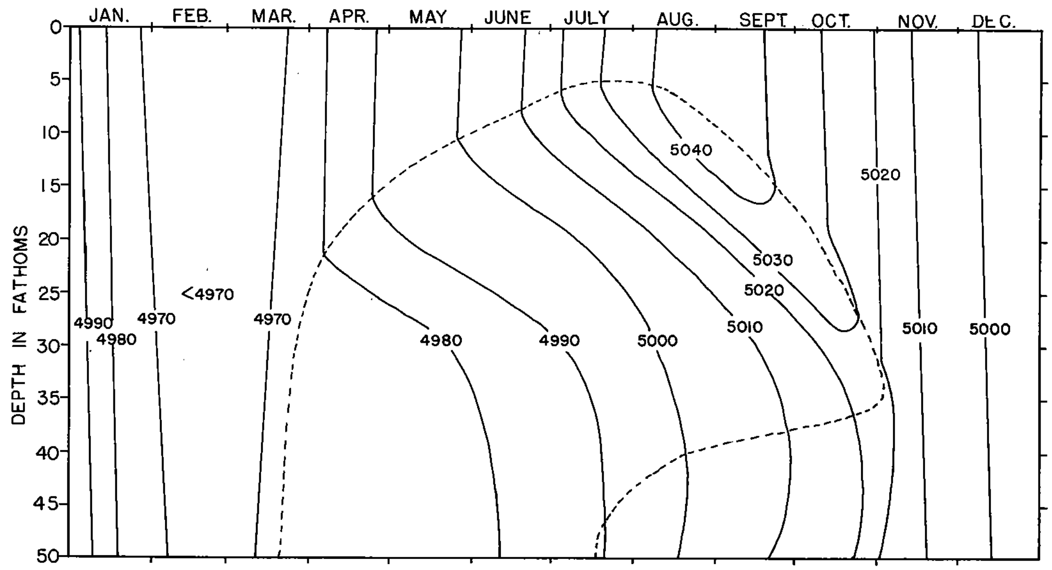

Fig. 14. Seasonal charges of sound speed (in feet per second) at depths of less than 50 fathoms in the western Sargasso Sea, computed from the temperature observations from which Figure 10 was constructed and from the salinity determinations made at the same time. Inside the broken curve the velocity gradient is negative, elsewhere it is positive.

The interpretation of Figure 14 in terms of horizontal ranging will be fully analyzed in Chapter V. It will suffice for the present purposes to explain again that in the parts of the diagram where the sound velocity increases with depth (positive velocity gradient) a sound beam will be bent slightly towards the surface and good sound ranging conditions prevail. On the other hand, in the parts of the diagram where the velocity values decrease with depth (negative velocity gradient) sound will be deflected downward and the range will be restricted.

During the winter months, when the upper 50 fathoms of the surface layer are well stirred by the winds, the sound velocity lines are slightly slanted because of the effect of pressure, but it will be noted that this slope is such that the velocity increases with depth. Elsewhere in the diagram the lines of equal sound velocity have an almost identical trend with the isotherms in Figure 13, showing the very close control which temperature exerts on the speed of sound in sea water. It is clear that for the whole Sargasso Sea area temperature observations alone are sufficient to predict the sound transmitting qualities of the water.

At the present time adequate data for such diagrams are unfortunately not available from many representative parts of the ocean. Except for the Sargasso Sea, it is only possible to show the complete seasonal cycle in coastal waters. Figure 15 has been constructed from observations made at a point about 60 miles south of Montauk Point, Long Island (see Fig. 20). This diagram is representative of the seasonal temperature cycle of the waters over the continental shelf between Delaware Bay and Nantucket.

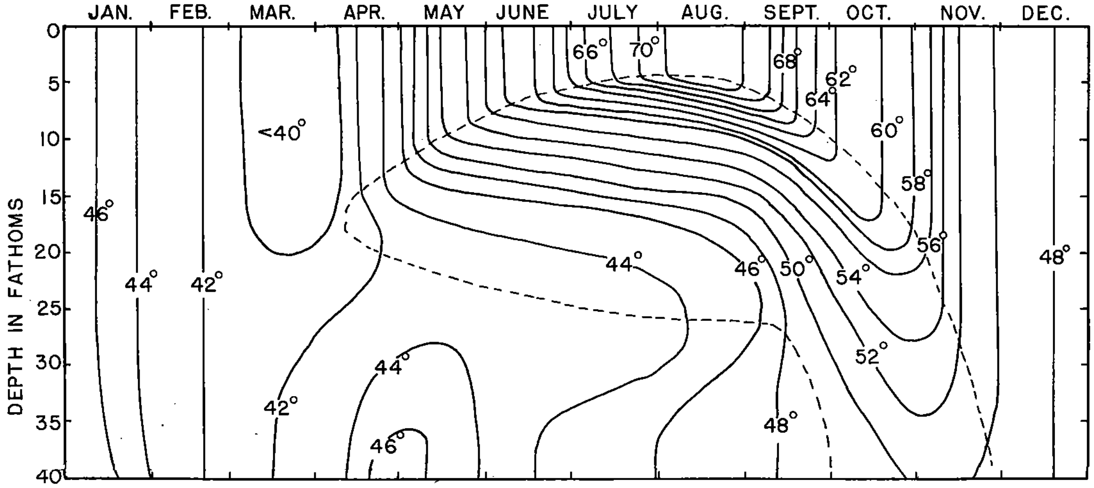

Fig. 15. Seasonal cycle of temperature, in coastal water off Montauk Point (approximately Lat. 40° 38', Long. 71° 38'). Within the broken curve the water is stable from the point of view of temperature. Elsewhere it is completely mixed by the wind or the water column is unstable from the point of view of temperature alone. The bottom is at 45 fathoms.

While the general form of this diagram is similar to Figure 13, representative of the Sargasso Sea, it will be seen that the temperature changes are more extreme near the coast and that sharper vertical stratification is produced in summer. The temperature minimum at mid-depths during the spring months should also be noted. Since we are here dealing with coastal waters, a temperature diagram may not give the full picture of the variations in stability or of sound ranging conditions. As mentioned above, in coastal waters the vertical gradient in salinity often contributes more to the stability of the water-column than does the temperature.

Unfortunately it is difficult from the available salinity observations off Montauk Point to construct a satisfactory diagram showing the seasonal cycle of salinity which accompanies the thermal changes given in Figure 15. The main trouble is that in summer the salinity of the wind stirred surface layer is quite variable, depending on the wind direction. At the point where the observations were made, offshore winds cause the salinity of the surface water to fall, while onshore winds have the reverse effect. During the period of the secondary thermocline the surface salinity at times may be as low as 30.5 °/oo and at other times it will rise to 32.4 °/oo, depending on the wind direction but such changes are accompanied by only slight temperature variations. Within the secondary thermocline (the area enclosed by the broken curve in Figure 15) the salinity increases from about 32 °/oo at a depth of 10 fathoms to as much as 34.6 °/oo at the bottom (45 fathoms). This large increase in salinity with depth during the season when the thermal stability of the water-column is high, has a beneficial effect on sound transmission, for such a distribution of salinity will cause less reduction of sound velocity with depth than would be expected from temperature alone. However, off Montauk Point the vertical salinity gradient is never great enough to counteract the control of sound velocity by the temperature. Nearer the bottom and during the summer months the increasing pressure and the increase in salinity with depth combine with the temperature gradient to produce a positive velocity gradient.

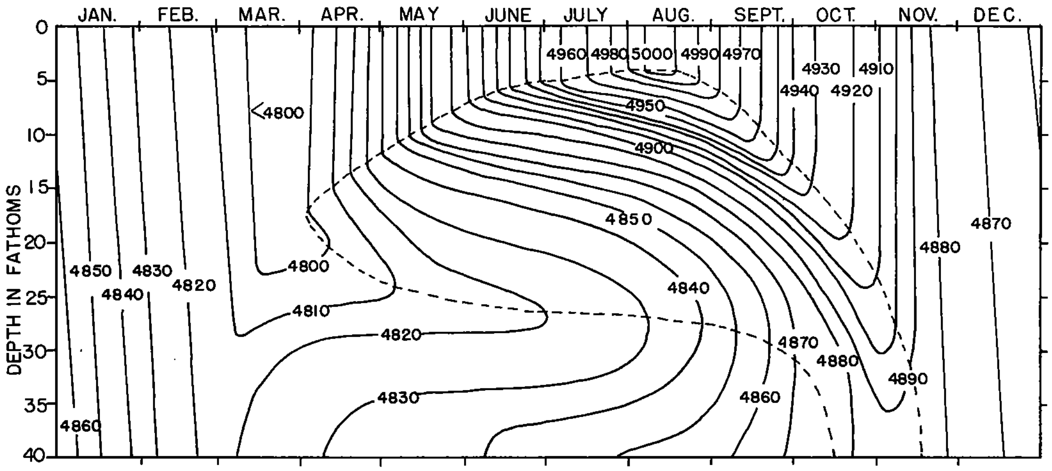

Fig. 16. Seasonal changes of sound speed (in feet per second) in coastal water off Montauk Point, computed from the temperature observations used in Figure 15 and from the salinity determinations made at the same time. The broken curve encloses the water in which the velocity gradient is negative, elsewhere it is positive.

This is brought out by Figure 16 which shows the seasonal cycle of sound speed derived from the temperature and salinity data available for the waters 60 miles south of Montauk Point. As mentioned above, Figures 15 and 16 are typical of the conditions over the continental shelf between Delaware Bay and Nantucket. It must be added, however, that in shallower waters, within 10 or 15 miles of the shore, the tidal stirring is often sufficient to considerably lengthen the part of the year when good sound ranging conditions prevail.

Between Cape Cod and Nova Scotia the continental shelf is wide and partly blocked off from the offshore water by the relatively shallow Georges Banks. This area is called the Gulf of Maine, and except for the bank water and the waters of the Bay of Fundy where strong tides keep the water column homogeneous even in summer, the seasonal cycle of temperature and salinity in this region is everywhere comparable to the conditions shown in Figure 17. These diagrams were constructed from monthly observations obtained at a point about 30 miles east of Cape Cod where the depth is about 130 fathoms. While the thermal cycle (A) is much the same as off Montauk Point (except that no layer of minimum temperature appears in spring), the seasonal variations in salinity (B) are more regular and the total vertical salinity gradient in midsummer is somewhat less.

The basic interchange between coastal water and offshore water is well illustrated in the Gulf of Maine salinity diagram (Fig. 17B). Over the continental shelf relatively saline water penetrates towards the land along the bottom, while fresher water spreads offshore at the surface. Turbulence is continually at work to reduce the contrast between surface and bottom salinities. Thus in the spring, as soon as thermal stability develops, lower salinities appear at the surface. At this season the rivers flowing into the Gulf of Maine are of course in freshet, and as the much diluted waters from bays and estuaries spread offshore, they intensify the thermal stability due to surface warming.

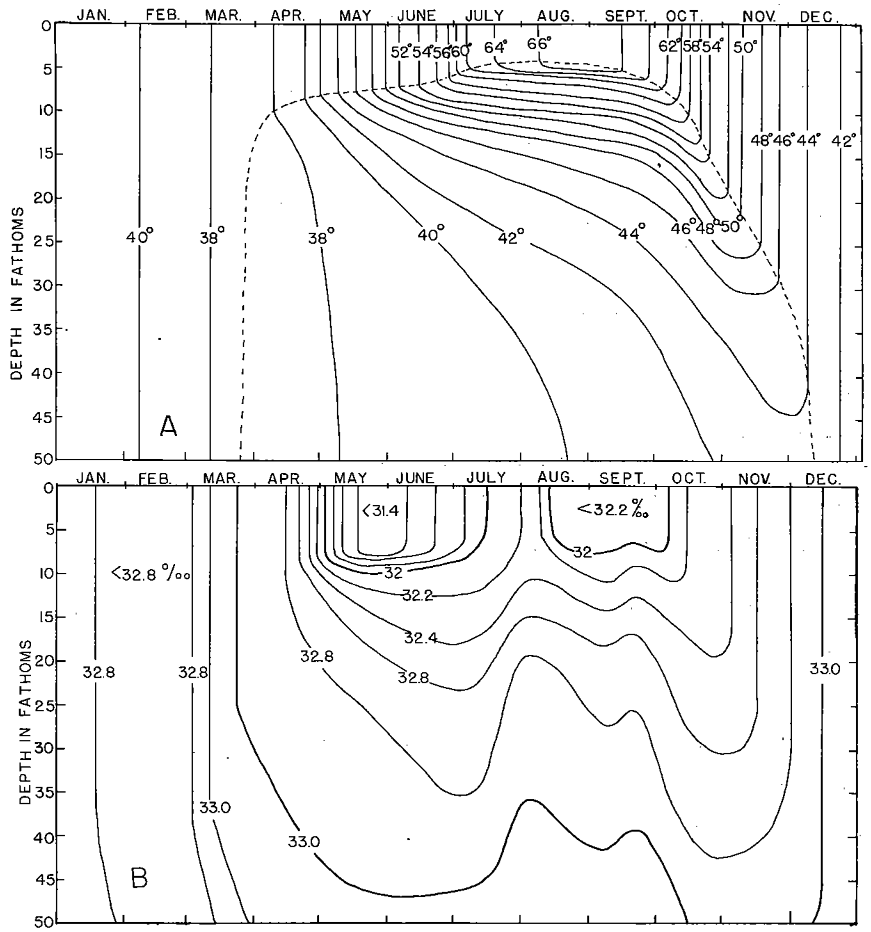

Fig. 17. Seasonal cycle of temperature (A) and salinity (B) in the Gulf of Maine. The observations were obtained at a point about 30 miles east of Cape Cod where the depth is approximately 130 fathoms.

In Figure 18 the temperature and salinity observations from which the diagrams in Figure 17 were constructed have been converted to sound velocity. It will be seen that in the Gulf of Maine temperature data alone is sufficient for a prediction of the seasonal variations in sound velocity. At no time in the upper 50 fathoms is the increase of salinity with depth sufficiently marked to overcome the control of the secondary thermocline on the sound velocity lines. It will be remembered that this was also the case off Montauk Point (Figs. 15 and 16). Apparently in coastal waters, as well as in offshore waters, the vertical gradient of salinity is unimportant from the point of view of sound transmission. But until more data is available we cannot be absolutely sure that this will be the case on all continental shelves and at all seasons.

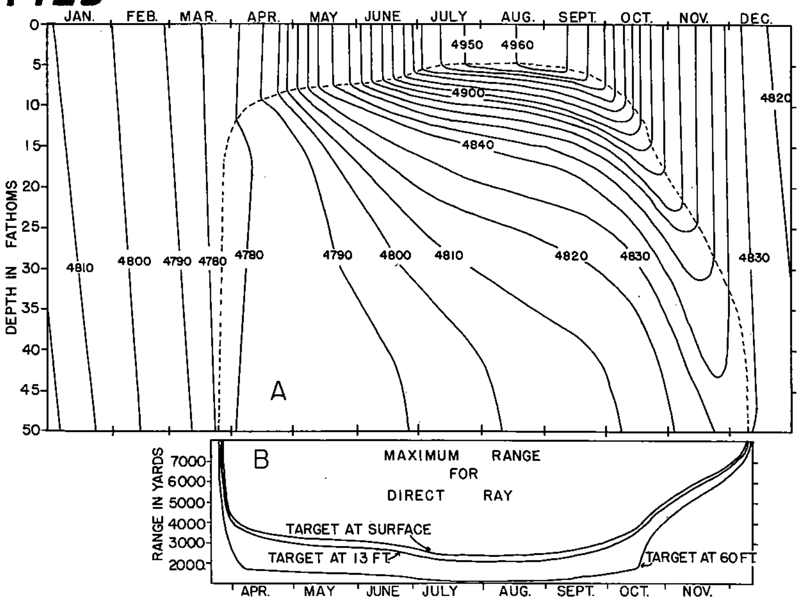

Fig. 18. Seasonal changes in the speed of sound (A) in the Gulf of Maine, computed from the data used in the construction of Figure 17. The broken curve encloses the water in which the velocity gradient is negative, elsewhere it is positive. In diagram B the maximum range for a direct sound ray has been calculated for the months in which negative gradients appear.

In order to give the seasonal sound velocity diagrams (Figs. 14, 16 and 18A) a more practical meaning in terms of range, Figure 18B has been constructed. The method by which these curves were calculated will be explained on page 54. Assuming that the depth of the sound projector is at 13 feet, the maximum range for a direct sound ray has been computed for three different levels. During December, January and February a direct sound ray will reach a target in the upper 60 feet at ranges of over 8000 yards. As soon as the secondary thermocline is formed in early spring the range is sharply decreased and then diminishes more gradually until midsummer. In the autumn the range increases again but more gradually, following the increase in thickness of the wind stirred layer at the surface. This correlation between the depth of the homogeneous layer at the surface and the maximum range of the direct sound ray also applies to Figure 16 for the waters off Montauk, and to Figure 14 for the Sargasso Sea, It is to be regretted that at the present time we have little reliable evidence to indicate how closely under ordinary conditions the limit of submarine detection corresponds to the maximum range of the direct ray (see page 72).

There are unfortunately very few parts of the sea where sufficient oceanographic observations have been secured at all seasons to warrant constructing such diagrams as have been given here for the waters of the western Sargasso Sea, and for the coastal waters off Montauk Point and in the Gulf of Maine. In order to demonstrate the seasonal change of temperature and salinity due purely to the local weather, the region must be relatively free from currents, while most oceanographic observations have been made in order to study currents. Moreover, data must be available from every month and at closely spaced depth intervals. In coastal areas temperature and salinity observations have usually been obtained in the course of fisheries investigations. Thus most of these data are from the spring and summer months when the important biological changes are in progress. On the other hand, for the offshore waters of mid-latitudes few observations have been made in winter, while in low latitudes few data are available from the summer months.

However, it is hoped that during the next year or two our knowledge of the seasonal cycle of the surface layer will be greatly extended in sufficient detail so that it can be reliably interpreted in terms of sound transmission. Once such information is available it will be possible to predict the general sound transmitting qualities month by month in the areas free from marked lateral gradients of temperature and salinity.

6. THE DIURNAL THERMAL CYCLE OF THE SURFACE LAYER

We have seen how the simple three layered ocean illustrated in Figure 8 really only holds true for the midwinter months, for superimposed on this sort of temperature-depth curve in the upper 50 fathoms or more is the seasonal cycle illustrated in Figure 13. Thus at most seasons of the year the wind stirred layer at the surface of the open ocean is separated from the permanent thermocline layer by a secondary thermocline of varying intensity and of varying depth. This sort of structure is, unfortunately, strictly true only during the night, for there is a daily (or diurnal) thermal cycle which is superimposed on the seasonal cycle.

The importance of this daily gain and loss of heat to supersonic ranging methods is not yet clearly understood, nor has it been accurately observed in many parts of the sea. It is known that, with a light wind and under certain circumstances to be discussed more fully below, during the morning a third thermocline can develop at a depth of a few fathoms. As the day progresses this gradually deepens and loses its intensity up to about four o'clock in the afternoon. From then on until early morning the surface is losing heat and the diurnal thermocline rapidly dissipates in the same manner as the secondary thermocline deepens and decreases in intensity during the autumn. In short, under certain conditions of wind and weather the diurnal thermal cycle is comparable to the seasonal thermal cycle, although on a much smaller scale.

The diurnal cycle can be illustrated by a series of observations obtained on a day when the winds were light (less than force 3) off Guantanamo, Cuba. Several temperature-depth curves of the upper 15 fathoms are given in Figure 19A, which is comparable to Figure 12B. Illustrated in another way, the diurnal cycle can be shown as in Figure 19B in which case the diagram is comparable to Figure 13, but with the time scale reduced to 24 hours and with isotherms drawn for each fifth of a degree.

The depth to which this diurnal cycle extends depends not only on the strength of the wind and on the time of year, but also on the cloudiness and on the humidity of the air within a few feet of the sea surface. It is believed that the diurnal cycle illustrated in Figure 19 is rather extreme both in the amount of heat absorbed and in the degree of stability which developed at depths critical to sound ranging. With more wind and clouds, and less humid air the diurnal cycle can only be observed with extremely sensitive thermometers. In short, during much of the time and over most of the ocean the daily gain and loss of heat in the surface layer is probably not sufficiently marked to interfere seriously with sound ranging, nor is this diurnal cycle ordinarily accompanied by a marked variation in salinity. Before discussing the factors involved more fully it is advisable to continue the discussion of larger scale phenomena.

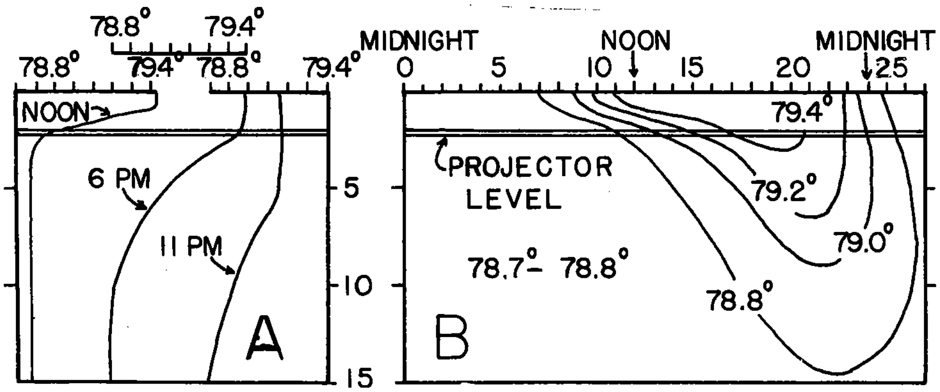

Fig. 19. Temperature-depth curves (A) of the upper 15 fathoms showing the diurnal heat exchange during light winds off Guantanamo. The complete cycle is shown diagramatically in B.

From the oceanographic data thus far presented the following general conclusions concerning the effective range of horizontal sound transmission in sea water can be drawn:

1) Sound ranging conditions should improve gradually during the early autumn and by November or December a sufficiently deep wind stirred surface layer is usually present in mid-latitudes so that the secondary thermocline is below the depth of submarine operation. The conditions will remain good until mid-March or later when, because of the warming at the surface, sufficient stability develops to greatly lessen the lower limit of wind stirring. This sequence of changes applies of course to the northern hemisphere, for south of the equator the seasons are reversed.

2) To a somewhat lesser degree the diurnal interchange of heat between the sea and the atmosphere, especially when the wind is light, may during the daytime produce some thermal stability at depths critical to sound ranging. Thus, in general, conditions should be better at night and in cloudy weather than during the part of the day when the sun is high.

3) While over most of the ocean the depth of the wind stirred surface layer is largely dependent on the seasonal and diurnal temperature cycles, in other areas the effect of the currents is equally important. Thus in any attempt to estimate the horizontal range of sound near the surface it is also necessary to take certain geographic factors into account.

7. HORIZONTAL STRUCTURE OF THE SURFACE LAYER

To illustrate the geographic factors which enter the sound ranging problem the surface layer of the western half of the North Atlantic will next be examined in the light of modern physical oceanography.

The permanent currents in this area are shown diagramatically in Figure 20. They can at once be subdivided into two groups: a) the currents in deep water, such as the Gulf Stream and the Northern Equatorial Current, which derive their energy from the general wind circulation over the whole ocean and b) the currents flowing over the continental shelf and along the continental slope, such as the Labrador Current, which are caused by the inflow of river water. As already explained, in the first group (with certain minor exceptions) the salinity decreases with depth, while in the second group (except in mid-winter) the salinity increases with depth and therefore adds to whatever thermal stability may be present.

Within a current, turbulence is of course greater than in relatively motionless water. Thus within a current, turbulence caused by the flow is added to the wind turbulence and a deeper layer of mixed water is usually present at the surface than on either side. The surface layer over a current may also differ somewhat from the surface waters on either side due to the actual transport of the water, especially where the current has a pronounced north-south trend. As might be expected the surface layer of a current flowing towards the north is warmer than the water on either side, and one flowing towards the south is colder. For this reason the edges of the currents having north-south components consitute boundaries in the horizontal structure of the surface layer as shown by the broken lines in Figure 20. Within each of the geographic subdivisions thus indicated the horizontal sound ranging conditions will be more or less uniform and will have comparable seasonal and diurnal cycles. However, at the edges of such currents, especially along the left hand edge in the northern hemisphere (right hand edge in the southern hemisphere), the two contrasting water masses do not blend readily, at times producing a complex thermal structure which can greatly interfere with sound transmission. Thus in general the boundaries of each subdivision in Figure 20 (shown by the broken lines) are rather uncertain for sound ranging. On the other hand, in the east-west flowing currents of low latitudes, such as the Northern Equatorial Current, the horizontal gradients of temperature and salinity are much less marked.

It will be noticed that the boundaries in Figure 20 are in general parallel to the coast so that as a vessel proceeds offshore from a port north of Cape Hatteras she will pass through three rather narrow bands of water, in each of which the sound ranging conditions may be more or less different. On the other hand, in an along shore direction a vessel will keep the same sort of water for a considerable distance. It is therefore convenient to examine the vertical and horizontal structure along profiles which are roughly at right angles to the current system, for example, along the line (shown in

Fig. 20) extending from Montauk Point, Long Island to Bermuda.

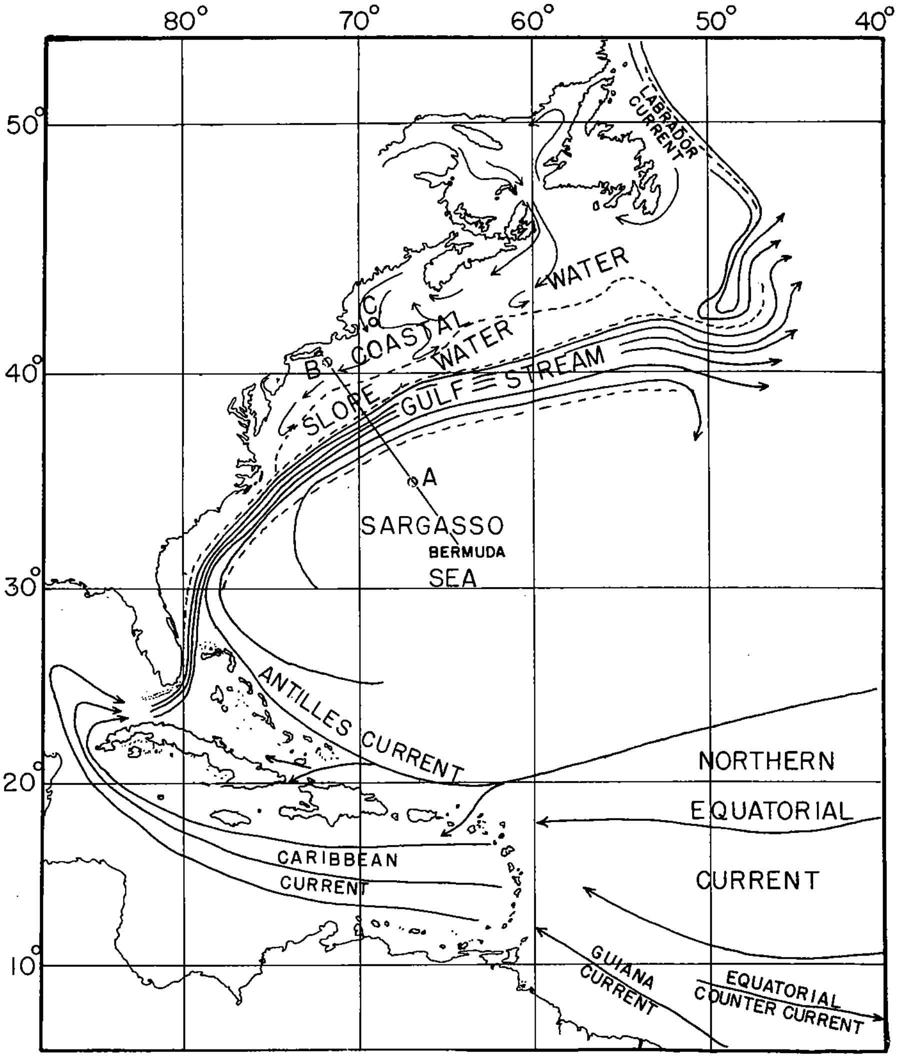

Fig. 20. Diagramatic representation of the main currents of the western North Atlantic. The stream lines represent the direction of flow at depths below the limit of local wind action. In the case of the Gulf Stream the lateral boundaries (broken lines) are somewhat further apart than is the case at any one time, for the current varies slightly in position.

Included will also be found the location of the three areas for which complete seasonal data are available: western Sargasso Sea (A), coastal water off Montauk Point (B) and coastal water in the Gulf of Maine (C).

The winter temperature conditions in the upper 50 fathoms along this section are shown in Figure 21. At this season of course there is little stability in the surface layer and sound ranging conditions are at their best. Near the coast the waters are mixed, surface to bottom, although temperature increases slightly in an offshore direction. As a result the isotherms in this area are nearly vertical and no sharp lateral contrasts occur at the surface until a point somewhat beyond the edge of the continental shelf has been reached. Here the closely packed isotherms on the diagram mark the offshore limit of the relatively cold and fresh coastal water. It will be noticed that below 20 fathoms along this slanting boundary the water is apparently unstable. However, this increase of temperature with depth is more than counteracted by the relatively low salinity of the coastal waters. Between this point and the Gulf Stream is a band about

80 miles wide, known in oceanography as the slope water area. In this zone at other seasons, as will be shown below, a complicated structure is found, but in winter the surface layer is well mixed to a sufficient depth for good sound ranging. It will be seen from the diagram that in midwinter temperatures are about 10°F. higher in the slope water than in the coastal water.

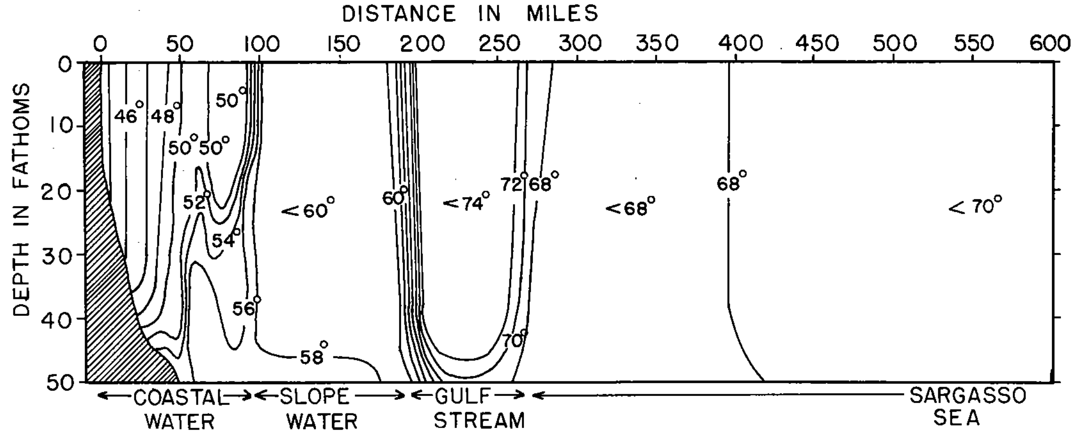

Fig. 21. Temperature distribution in January along a profile extending from Montauk Point to Bermuda. It should be noticed that the depth scale is so exaggerated in this and in the next two diagrams that closely packed, vertical isolines do not indicate lack of stability. On the contrary, at the contact between the slope water and the Gulf Stream for example, the warm water usually overlies the colder water for a distance of 10 miles or so.

The slope water in turn is separated from the still warmer Gulf Stream water by another sharp zone of thermal transition, but on the southern edge of the current there is a less marked change as one enters the slightly colder Sargasso Sea. Thus the four subdivisions shown in Figure 20 can be clearly made out from the midwinter temperature profile (Fig. 21). But in all four types of water the isotherms of the upper 40 fathoms are so nearly vertical (or indicate an increase of temperature with depth) that little stability could have prevailed. Thus with the possible exception of the layer between 40 and 50 fathoms in the slope water and in the Gulf Stream and at the three boundary zones one would judge that in midwinter excellent sound ranging conditions must prevail along this 600 mile section.

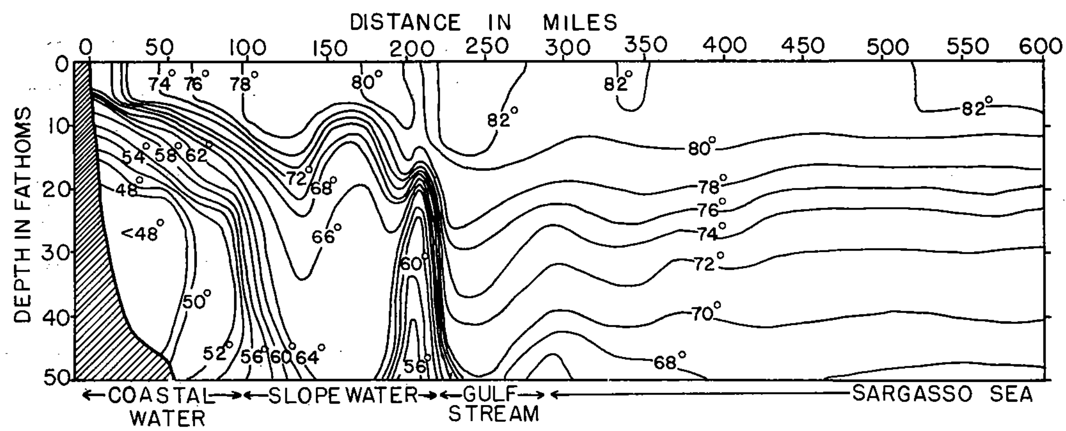

Fig. 22. Temperature distribution in August along a profile extending from Montauk Point to Bermuda.

The midsummer conditions along the same profile (Fig. 22) are in sharp contrast, for in this case many of the isotherms have a fairly horizontal course, indicating stable conditions. This results from the development of the secondary thermocline, caused by the warming of the superficial layer. It will be seen that over the continental shelf the surface water is about 26°F. warmer than the bottom water and that the greater part of this gradient is at depths between 5 and 21 fathoms. This layer of maximum stability increased in depth slightly in an offshore direction, but north of the Gulf Stream the layer of mixed water at the surface was nowhere more than 12 fathoms thick.

The offshore limit of the coastal water is marked by the outward bulge of the relatively cold water at depths between 25 and 45 fathoms. It will be noticed also that below this layer of minimum temperature, just as in winter, the water appears to be unstable from the point of view of temperature alone.

In the slope water band the isotherms, especially those between 58° and 64°, have a rather wavy course. This results from a large clockwise eddy which was present north of the Gulf Stream at the time this section was occupied. Such eddies are frequently met with in this band.

The northern edge of the Gulf Stream is clearly marked by the closely packed isotherms at a distance of about 210 miles from the coast. Over the current the thickness of the wind stirred layer at the surface, with temperatures higher than 82°, increased to 15 fathoms. It will be noticed that the Gulf Stream water was only about 2° warmer than the water at corresponding depths in the Sargasso Sea and that in the latter area the secondary thermocline was much less sharp than north of the current. Nevertheless, south of the Gulf Stream sufficient thermal stability existed everywhere below a depth of 12 or 14 fathoms so that it is believed that very indifferent sound ranging conditions prevailed.

The midwinter salinity distribution has a similar pattern to the thermal structure shown in Figure 21, because in each of the four zones wind stirring had penetrated down to at least 40 fathoms. Therefore, from the point of view of sound ranging little would be gained by adding a diagram to show the salinity observations obtained at the time the temperature data were secured. Such a diagram would only show that the coastal water had a salinity of about 33 °/oo, the slope water 35 °/oo, the Gulf Stream 36.4 °/oo and the Sargasso Sea 36.6 °/oo.

On the other hand, in midsummer the vertical salinity gradients are by no means insignificant as shown in Figure 23. From this diagram it can be seen that in the coastal water the surface salinity was about 1 part per thousand less than near the bottom, while in the slope water band there was a difference of as much as 1.2 parts per thousand between the surface and 50 fathoms. In the Gulf Stream and in the Sargasso Sea this vertical salinity gradient was much less, amounting to only about 0.4 parts per thousand.

In addition, in these two water masses the increase in salinity with depth was more gradual.

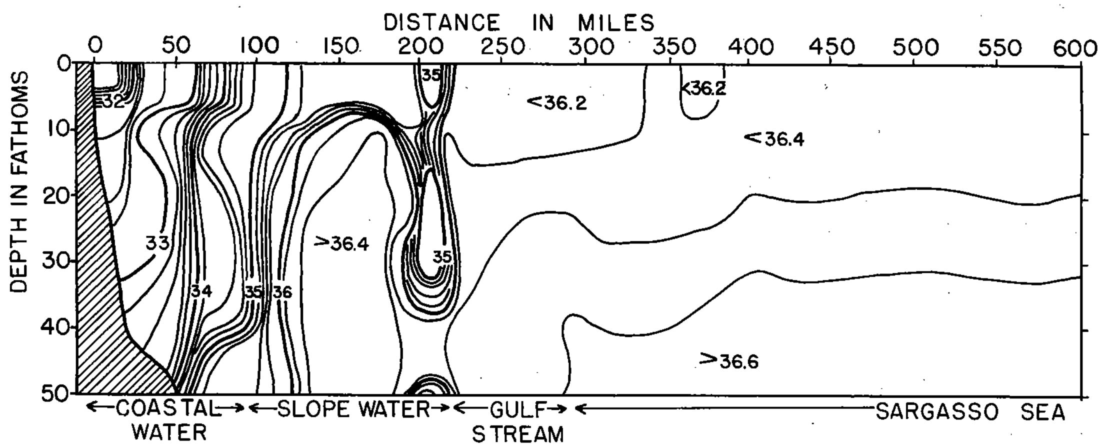

Fig. 23. Salinity distribution in August along a profile extending from Montauk Point to Bermuda.

It was earlier stated that in general in the open ocean salinity decreased with depth. Why then does the Gulf Stream water and Sargasso water in Figure 23 show the opposite gradient? Considering the whole offshore water-column the salinity does decrease with depth, but in the surface layer some exceptions are found. The explanation in this case is that the observations come from summer when the surface salinity is slightly lower than in winter. In offshore waters the salinity near the surface is determined by the balance between evaporation and rainfall. In the western North Atlantic the rainfall is relatively constant throughout the year. However, in winter evaporation is much more marked than in summer. Not only are the winter winds much stronger, but they are a great deal drier than in summer. This marked seasonal cycle in evaporation causes a seasonal variation in surface salinity, the change in the Sargasso Sea area being from 36.6 °/oo in winter to 36.2 °/oo in summer. It is for this reason that summer sections, such as Figure 23, show a slight increase in salinity with depth, but from the point of view of sound transmission this favorable factor is completely masked by the secondary thermocline.

The salinity diagram also shows that along the northern edge of the Gulf Stream quite fresh (35 °/oo) masses of water were in close contact with the saline waters (36.2 °/oo) of the current. Such sharp horizontal contrasts in salinity are very common along the left hand edge of the Gulf Stream and introduce an added difficulty to sound ranging. Thus both from the point of view of vertical and of lateral gradients the northern boundary of the current is a very difficult zone in summer for successful sound transmission.

As mentioned above, any section normal to the trend of the coast line from Cape Hatteras eastward would have the same basic structure exhibited by Figures 21-23. Of course the width of the coastal water and the width of the slope water varies somewhat from one part of the coast to another, but in each band the stability will be comparable in an along shore direction, whatever the season.

The complete seasonal cycle of thermal stability has already been given for the coastal water off Montauk Point (Fig. 15) and for the western Sargasso Sea (Fig. 13). The seasonal changes within the surface layer of the Gulf Stream are almost identical with the latter, except that the increased turbulence maintains a slightly deeper homogeneous layer at the surface throughout the year. The conditions in the slope water band cannot be so easily systematized. In the first place, the offshore limit of the relatively fresh coastal water varies quite widely, depending on the prevailing winds, and in the second place warm Gulf Stream water is frequently discharged at the surface north of the current. Thus in the slope water band at one time the vertical salinity gradients may be comparable to coastal water conditions, while nearby or at another time the water may be almost indistinguishable from Gulf Stream water. In short, the slope water is a mixing zone for coastal and Gulf Stream water and only occasionally is the mixture sufficiently complete so that it can be said that intermediate conditions prevail. For these reasons only during the midwinter months is it possible to predict with any certainty what the sound transmitting qualities of this band of water will be. Fortunately such zones, where contrasting water masses are being continually blended by large scale turbulence, do not occupy a large part of the sea. Moreover, their limits can on the whole be mapped from existing oceanographic data. Elsewhere the vertical and horizontal gradients are much more gradual, and predictable with some certainty when seasonal and diurnal factors are taken into consideration.

To further emphasize the close control which temperature exerts on the sound velocity gradients in the various geographical subdivisions of the western North Atlantic the midwinter and midsummer oceanographic data from the Montauk Point-Bermuda section have been converted to sound velocity and are presented in Figure 24. A comparison of these velocity profiles with the corresponding temperature sections (Figs. 21 and 22) shows the striking similarity between the distribution of temperature and sound velocity. In midwinter (A), except at depths of between 40 and 50 fathoms in the slope water and in the Gulf Stream, the sound velocity gradient is everywhere positive; while in midsummer (B) it is negative with the exception of the deeper waters over the continental shelf. In the first case the exception is caused by slight thermal stability (see Fig. 21) and in the second case by the relatively warm bottom water near the edge of the continental shelf (see Fig. 22). But as already noted in connection with Figure 16, in summer the good sound water near the bottom over the continental shelf is not usable by a surface craft.

From this examination of the seasonal and diurnal cycles and of the vertical and horizontal structure of the waters off the eastern coast of the United States it seems safe to conclude that the variations in salinity, although at times important in maintaining stability, play a very small part in determining the sound transmitting qualities of the upper 50 fathoms. For all practical purposes a knowledge of the vertical temperature gradients is all that is required.

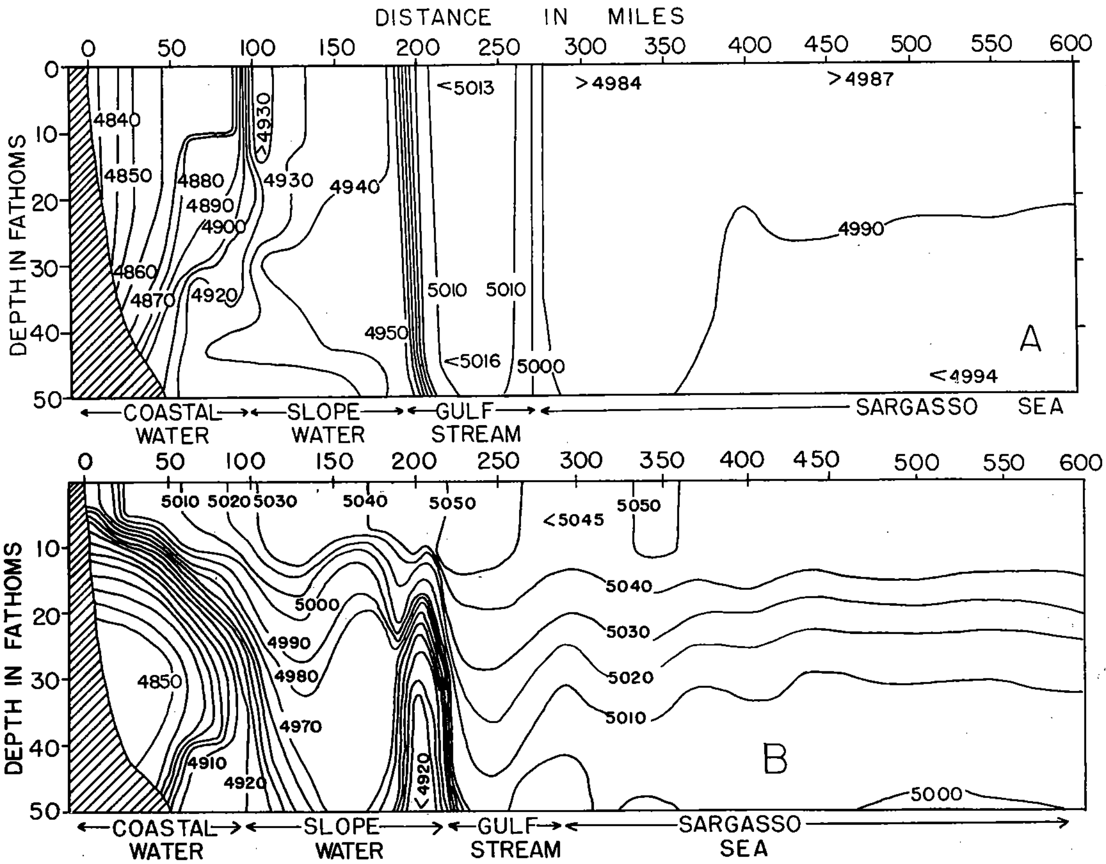

Fig. 24. Speed of sound (in feet per second) in January (A) and August (B) on the Montauk Point-Bermuda section.

8. INTERACTION BETWEEN THE SEA SURFACE AND THE ATMOSPHERE

It has been shown repeatedly in the previous pages how extremely important the heat exchange across the sea surface is to the horizontal transmission of sound. When the sea surface is gaining heat, and therefore a tendency for increasing stability prevails, the sound transmitting qualities of the water will become relatively poorer. When the sea surface is losing heat, so that the waters near the surface tend to become slightly unstable, the winds are able to maintain a relatively deep mixed layer within which the pressure effect will prevent sound from being refracted downwards. The importance of the heat exchange at the surface is equally great whether diurnal or seasonal variations are being considered. Furthermore, the question of turbulence in the superficial layer is also closely connected with the thermal interchange between the sea and the atmosphere.

At the sea surface two fields of science, oceanography and meteorology, meet. From the oceanographic standpoint the force which the winds exert on the surface drives the currents from above. From the meteorological point of view the sea is retarding the winds from below. In the ocean the currents decrease in velocity with depth, while in the atmosphere the winds increase in velocity with altitude. These vertical velocity gradients and the accompaning turbulence are intimately connected with the stability or lack of stability in both mediums.

For convenience turbulence can be thought of as having two components, vertical and horizontal. Where stability is great the vertical component is small and the stirring must express itself mainly in lateral movements. In either stable air or stable water the turbulent motion can be thought of as being flattened from above. Under such circumstances the horizontal component of turbulence is increased. On the other hand, where there is no stability the horizontal and vertical components of turbulent motion will be roughly equal.

Turbulence in the surface layer of the sea is comparable to a system of eddies, especially under stable conditions. These eddy motions seem more often than not to occur on several definite scales. There is first of all molecular motion, but this is such an extremely small scale phenomenon that it accomplishes only a relatively small part of the mixing. Next there is the ordinary small scale turbulence whereby most of the wind stirring is carried out. Here the scale of the eddies is of the order of magnitude of a few inches or at most a few feet. Turbulence on this scale no doubt also is caused by the shear in waters where the velocity of flow is changing gradually with depth. Then there are eddies which may be roughly a mile or more in diameter, for example, the smaller eddies one would expect to find along the edges of a powerful current. The next jump seems to be to eddies which are roughly 60-100 miles in diameter. Finally, mixing occurs on an ocean wide scale, and this produces the permanent current systems. It is the second of these typical eddy sizes, with dimensions of the order of magnitude of a few inches or a few feet, which is of major importance in the transfer of energy between the sea surface and the atmosphere. The waves are one of the manifestations of this sort of turbulent motion.

Under stable conditions the heat absorbed from the sun in the upper few feet is carried downward largely by wind stirring. As has been shown above, when the sea surface is gaining heat the winds are only able to maintain a relatively shallow layer of completely mixed water. During a calm period marked thermal stability may be found in the upper 10 feet or so. But even strong winds in summer cannot maintain a homogeneous layer at the surface more than 60 or 100 feet deep.

When the surface is losing heat, <u>convection</u> is added to wind turbulence and the homogeneous layer at the surface deepens rapidly. Convection is independent of the wind and will take place either in a calm or in a gale. Convection is the motion which occurs when, through cooling and through an increase in salinity (resulting from evaporation), water near the surface becomes heavier than the water just beneath.

What are the processes whereby the surface gains or loses heat? There is first of all <u>conduction</u>, the direct interchange of heat due to the contact between the air and the water. Only about 5% of the heat transfer across the sea surface can be accounted for by conduction.

A much more effective mechanism is the absorption of solar radiation, for, since water is relatively transparent to light rays, the thermal change in this case is not confined to the immediate surface. Transparency of sea water is rather variable, but in round numbers it may be stated that half the energy of solar radiation penetrates the first half fathom of water. Some wave lengths are almost completely absorbed in this layer and the remaining wave lengths are more penetrating, so one quarter of the incident energy usually goes deeper than two fathoms.

Radiation is not a reversible process. Water is far more opaque to the long wave radiation originating in the ocean (or atmosphere) than it is to solar radiation, of which a large part of the energy is in the short wave or light band. Consequently the back radiation from the water originates in a thin surface film; most of the emitted energy comes from a surface layer less than a thousandth of an inch in thickness. Thus the loss of heat by back radiation comes from a much shallower layer than that which absorbs solar radiation in daytime. From this point of view the ocean surface is comparable to the glass roof of a greenhouse.

The rate at which the absorption of solar radiation raises the temperature of the water depends of course on the altitude of the sun and on the presence or absence of clouds, but on the average this process accounts for about 95% of the heat gained by the ocean. Back radiation on the other hand is responsible for only about 40% of the heat lost from the superficial layer. The remainder is largely accounted for by evaporation.

Evaporation is a surprisingly effective method of cooling the water. It of course acts on the immediate surface and further accentuates the temperature gradient of the surface film produced by back radiation. However, by mixing and internal radiation within the surface film an unstable surface layer an inch or more in thickness is probably often formed. Thus in daytime, especially in clear dry weather with light winds, a relatively dense surface layer is maintained while at the same time solar radiation steadily decreases the density of a much thicker layer just below. Under such circumstances the heavy surface water must break through the warmer layer beneath. The pattern of flow which is brought about by this sort of convection has been studied only very recently. Needless to say it is not easy to observe such phenomena in the open ocean. However, the behavior of the gulf weed in the North Atlantic provides convincing evidence of the importance of this process.

Preliminary observations indicate that when the gulf weed is scattered at random the superficial waters are stable, while when the gulf weed is lined up surface cooling is taking place. The lines of gulf weed are evidence of convergence of the relatively dense surface water as it sinks along a vertical plane through the warmer waters just below. These lines of convergence extend downwind and are spaced 50 to 100 feet apart. Under such circumstances the superficial layer of water has a cellular structure which, as will be discussed below (page 62), may not be without significance to sound ranging.

This type of motion is illustrated diagrammatically in Figure 25.

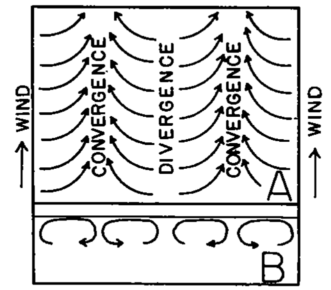

Fig. 25. Schematic representation of the convectional pattern near the surface during a moderate wind. The horizontal view is shown about (A) while a cross wind section is given below (B).

Such convectional cells in the surface layer may not be independent from similar cells in the air just above the water. Much further research is required. However, sufficient observations have been made to indicate that this is probably a general phenomenon. Alternating bands of convergent and divergent motion have been seen, for example, in a lake where it was thought that the wind was the cause, but the importance of evaporation in this process has only recently become apparent. The fact that the cells are always lined up with the wind indicates that the motion is not independent of the wind, but it is also significant that lines of gulf weed have not been observed, except under conditions when the surface film was being cooled by evaporation and conduction.

The rate at which evaporation proceeds depends on the dryness of the air (indicated by the temperature difference between wet and dry bulb thermometers), and ceases completely when the air is saturated. Rate of evaporation depends also on wind speed, becoming very slight in a calm. The <u>relative humidity</u> of the air, an inverse indicator of its dryness, varies locally with the wind direction. In the northern hemisphere southerly winds are usually relatively moist, while northerly winds tend to be drier. But there is also a much larger scale pattern of humidity, not without significance to sound ranging. In the western part of the oceans, especially in winter and in the latitudes of prevailing westerly winds, the winds are relatively dry as they come off the continents. Thus outside the trade wind belts evaporation is greater in the west than in the east. This means that in the latitudes where westerly winds prevail thermal stability will be more common in the surface layer in the eastern half of the oceans than in the west.

Evaporation is also more rapid in the horse latitude belts and along the northern edge of the trade wind belts than nearer the equator. The very high surface salinity of these areas (see Fig. 11A) is good evidence of the relative dryness of the air. The general wind system is such that in the horse latitude belts the air is descending. Thus it is being compressed and therefore warmed as it descends. This increases its capacity to take up water vapor. As the trade winds converge towards the equator they become saturated with water vapor so that only along the northern (and, in the southern hemisphere, the southern) edge of the trades can evaporation be important. At the equator the converging air must rise and this results in much rain and a stable surface film.

Enough has been written to indicate that in the superficial layer of the ocean rather complex processes are at work. The upper few feet of water are by no means structureless, especially when the air is dry and the wind is light. Whether or not any of this structure is of much importance to horizontal sound transmission is not yet known, but, as will be suggested in Chapter V, large scale convection cells near the surface may be the cause of reverberations in sound reception. Moreover, turbulence and convectional flow within this layer may break up the directional qualities of high frequency sound as it approaches the surface. In all probability until more is known about the physical oceanography of the upper few feet of the sea, a full understanding of underwater sound transmission will not be possible.

The effect of the winds in producing turbulence in the surface layer has been briefly discussed, but the winds also cause the actual transport of surface water. Where the winds are steady, as in the trade wind belts, after sufficient time deep powerful currents develop, because turbulence carries the energy from the winds downwards. But such movements are not particularly detrimental to sound transmission, for in permanent wind currents turbulence due to vertical shear also minimizes the thermal structure of the water. On the other hand, light winds which are variable in direction are capable of causing very bad sound conditions when stability prevails. This is especially true near the coast and along the border of a current with a marked north-south component. Where the surface varies in temperature and salinity horizontally, a temporary wind can cause a relatively thin sheet of water to over-ride another body of water with different physical characteristics. For example, relatively warm and saline water from the slope water area off New England may be blown inshore so that it overlies the colder and less saline surface layer of the coastal water. Under such conditions there may be little or no difference in density between the two types of water, but the resulting temporary thermal gradient may be very bad for sound. This is one of the reasons why the boundary zones between contrasting water masses are most uncertain for sound ranging.

As such thin sheets of water with contrasting physical characteristics are moved one over the other by a moderate wind and when there is too little turbulence to break through the stability, a sharply stratified surface layer is produced. Under such circumstances the vertical thermal structure will frequently indicate instability. In Figure 26 two such examples are given. However, in such cases the apparent instability is always counteracted by salinity. In other words, regardless of its temperature the less dense sheet of water is carried by the wind over a somewhat heavier layer.

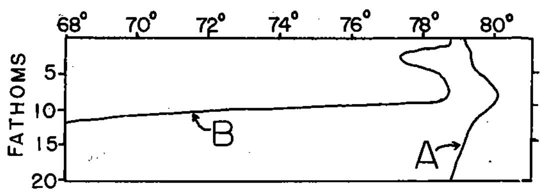

Fig. 26. Temperature-depth curves illustrating the sliding of one layer over another by a light wind. The observations for curve A were made off the Atlantic end of the Panama Canal in winter. Curve B shows a more complex case from the slope water area south of Montauk Point in summer. In both cases the salinity distribution more than counteracted the apparent thermal instability.

This sort of stratification cannot continue if the wind holds steady in direction, nor can it develop during periods when the stability near the surface is low. It is much more common near the coast than in the open ocean and outside of the tropics it can only occur during the summer season. However, it is this sort of layering which in the past has often made the sound ranging problem seem so hopelessly complex. It is fortunate that only over relatively small parts of the sea are the horizontal contrasts in the surface layer sufficiently marked to make this wind effect possible.

9. INTERNAL WAVES, TIDAL STIRRING AND UPWELLING

When the winds are light the diurnal thermal cycle and the sliding of one stable layer over another by a moderate wind may combine to produce a complicated distribution of temperature with depth within 20 or 30 feet of the surface. As will be shown in Chapter V, the position of the sound projector relative to such shallow temperature gradients is extremely important to horizontal signaling methods. For example, the range will be much greater when the thickness of the wind stirred water at the surface is greater than the depth of the sound projector than when the beam starts out in a layer of thermal stability. It becomes necessary, therefore, to consider whether or not the <u>internal waves</u>, which exist in all large bodies of stable water, can so vary the thickness of the various superficial layers that the projector will be first in one sort of water and a short time later in another. The term internal wave is applied to such waves as are propagated along the interface between two layers of water of different density.

This phenomenon has been studied in the main thermocline layer and in the secondary thermocline layer, but never in the diurnal thermocline or in such temperature discontinuities as may be caused by wind transport within 20 feet or so of the surface. In the main thermocline layer 12 hour and 6 hour waves exist which may alter the depth of a given isotherm by as much as 150 feet in the course of a few hours. Such internal waves are apparently not due entirely to the tides. A storm can send out subsurface waves in the same manner as swell can travel great distances from its region of origin.

Within the secondary thermocline internal waves have been frequently observed in coastal waters, where, as might be expected, their amplitude is much smaller than at mid-depths in the open ocean, for obviously the amplitude of such a wave must approach 0 at the surface. In Figure 27 a series of observations, obtained over a 24 hour period in the Gulf of Maine, are shown to illustrate the magnitude of short period waves in a summer thermocline. It is believed that this is an extreme example because of the strong tides. Other areas show waves with about half as much amplitude.

It will be seen from the diagram that the 54° isotherm changed in depth by about 60 feet in roughly a six hour period, while the 44° isotherm varied about 100 feet in a similar length of time. Thus within the secondary thermocline we have evidence that the amplitude of the internal waves decreases towards the surface. What would be their amplitude at a depth of 15 feet which is roughly the depth of the oscillator on most surface ships equipped for supersonic ranging? No observations are available to help answer this question. A reasonable guess would be that at a depth of 15 feet a given water particle might change in depth by 10 feet due to the passage of an internal wave. However, the mean amplitude of such movements is probably more nearly half this figure.

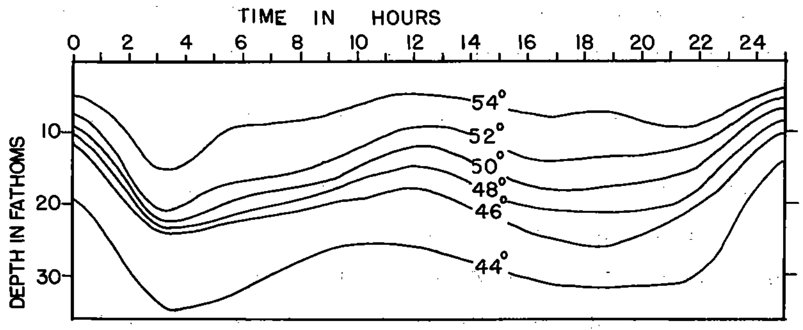

Fig. 27. Internal waves in the eastern part of the Gulf of Maine in summer. Temperature observations were secured at intervals of about one and a half hours during a 25 hour period.

If these estimates are at all reliable, there is a distinct possibility that whenever fine scale thermal structure exists at depths within a few feet of the sound projector the range will vary with the passage of each internal wave. The $6\frac{1}{2}$ hour tidal cycle should largely control the period of these variations in range. The chief significance of these considerations is that when such conditions prevail the analysis of sound transmission tests will be unreliable unless temperature data are secured at least once an hour.

Under most circumstances the tides are not able to break down the tendency for stability in the surface layer of the ocean. In the open ocean, as has been discussed above, turbulence caused by the flow of the tides plays a minor part compared with wind stirring in overcoming stability near the surface. From the point of view of sound transmission in most regions the tides are only important in so far as they partly control the frequency of internal waves. However, on the continental shelf under certain circumstances tidal turbulence is able to keep the water column mixed from surface to bottom.

The Georges Banks area off the Gulf of Maine provides a good example of such a case. The depth over the top of the bank varies mostly between 10 and 50 fathoms and strong tidal currents prevail. Only for about 6 weeks in the middle of summer is the absorption of solar radiation able to establish a secondary thermocline. Throughout the rest of the year wind stirring and tidal turbulence keep the water completely mixed from surface to bottom. The limits of this homogeneous water coincide roughly with the 40 fathom depth contour.

The Grand Banks of Newfoundland is another such area, but here the tidal currents are not as strong as on Georges Banks, so the summer stable period is considerably longer. Few observations are available from the central part of the Grand Banks, but its waters are probably homogeneous, surface to bottom, from early October to late May.

Besides increasing turbulence over a shallow bank, the tides often considerably shorten the period of thermal stability in waters near the coast. Within 5 or 10 miles of the beach in many areas the water-column is well mixed, except for a short period in midsummer. In addition to shortening the period of the secondary thermocline, tidal stirring is no doubt sufficiently vigorous in some areas to prevent the formation of the diurnal thermocline. Thus strong tides and strong winds work together to maintain favorable conditions for horizontal sound transmission.

There is one circumstance in which a moderate or strong wind will have just the opposite effect, namely when it persistently blows in an offshore direction. Steady offshore winds will keep the surface layer moving away from the coast and therefore near the shore will cause deeper water to rise towards the surface. This is know as up-welling. The vertical component of such motion is usually relatively slight, but since the wind stirred surface layer is continually being blown offshore the thermocline may extend almost to the surface, resulting in bad sound ranging conditions.

The best examples of fairly wide spread, unfavorable sound conditions due to upwelling are found in the latitudes of the trade winds on the eastern side of the oceans. The west coast of Africa is one example. Of more importance to naval operations are the waters off the Pacific entrance to the Panama Canal where similar conditions prevail (see page 68).

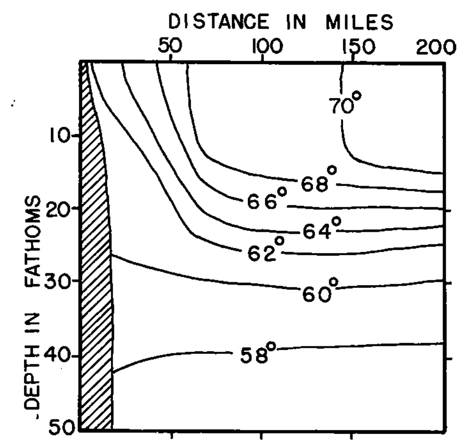

Fig. 28. Temperature section illustrating upwelling due to a steady offshore wind. The observations extended westward from Dakar on the west coast of Africa at approximately latitude 15°N.

The temperature distribution on a profile (Fig. 28) extending offshore from the port of Dakar on the African coast illustrates the phenomenon of upwelling. Here the continental shelf is narrow and at the time the section was occupied (February) the trade winds had an offshore component. It will be seen that for at least 50 miles off the coast thermally stable water extended right to the surface, while further out 12 to 15 fathoms of wind-stirred water overlay the thermocline.

11. SUMMARY AND CONCLUSIONS

Various oceanographic factors which influence the transmission of sound in the surface layer of the ocean have been discussed. It is believed that the available temperature and salinity observations from the western North Atlantic permit the following general conclusions:

1) In the deep ocean the water above the permanent thermocline layer is only entirely without vertical structure for a relatively short period in midwinter. At other seasons varying degrees of thermal stability will be found in the upper 50 fathoms.

2) The development of a secondary thermocline in early spring causes a sharp decrease in the thickness of the wind stirred layer at the surface. This is followed by a more gradual decrease during the early summer so that by midsummer only 5 to 15 fathoms of mixed water may overlie a layer of great thermal stability. During the autumn, as the surface layer loses heat, wind-stirring penetrates deeper and deeper.

3) This seasonal thermal cycle in the surface layer is particularly marked near the coast and in the latitudes of prevailing westerly winds. Nearer the equator the secondary thermocline cannot be so clearly differentiated from the main thermocline, which in low latitudes approaches the surface.

4) In addition, during periods of calm or light winds a diurnal thermocline may develop at a depth of only a few fathoms. Under such circumstances during the daylight hours, especially between 9 A. M. and 4 P. M. the layer of mixed water at the surface may be very thin indeed.

While these three layers of thermal stability (permanent thermocline, secondary thermocline and diurnal thermocline) are unfavorable to horizontal sound transmission, various processes whereby the surface loses heat to the atmosphere (chiefly evaporation and back radiation), especially when accompanied by turbulence, have the opposite effect on sonic ranging results.

5) The greater part of the mixing in the surface layer is caused by the winds, but vertical shear, either in permanent currents or in tidal currents, also plays a part. Thus near the coast (except when prolonged offshore winds prevail), over a shallow bank or along the path of a steady current the wind stirred layer at the surface will be deeper than elsewhere during the period of the secondary thermocline.

6) In regions where the permanent currents have north-south components the seas will also have a horizontal thermal structure. Through wind transport or through large scale mixing processes the neighboring contrasting water masses may temporarily overlie one another. Thus the borders of such currents, especially their left hand edge in the northern hemisphere (right hand edge in the southern hemisphere), in the absence of strong winds are often unfavorable to sound ranging.

7) The surface layer of the western North Atlantic can be subdivided into four parts

on the basis of its vertical and horizontal structure:

a) The coastal water, which covers the continental shelf and often extends offshore at the surface considerably beyond the 100 fathom curve, is characterized by an increase in salinity with depth and during the spring and early summer by a temperature minimum at mid-depths.

b) From Cape Hatteras eastward to the Grand Banks the coastal water is separated from the Gulf Stream by the so-called slope water. This band, which is mostly between 40 and 100 miles in width, has only weak permanent currents, but frequently there are large eddies. The surface layer of the slope water is a mixing zone for bodies of relatively fresh coastal water, which have become detached, and masses of Gulf Stream water carried north of the current by eddies. The vertical and horizontal gradients are therefore variable and unpredictable in this region.

c) The Gulf Stream waters are of course relatively warm and the mixed layer at the surface over the current is always deeper than on either side. Marked lateral gradients of temperature and salinity are always found at the left hand edge of the currents. At such boundaries it is usual for the less dense water to slightly overlie the heavier, and consequently the margins of a current often have bad sound transmitting characteristics.

d) The Sargasso Sea extends from the right hand edge of the Gulf Stream eastward to the Azores. In winter its waters are homogeneous to a depth of 50 fathoms or more. No marked horizontal gradients are met with and the secondary thermocline is relatively weak when compared with the slope water or with the coastal waters.

8) In no part of the western North Atlantic thus far examined has it been possible to show that the vertical salinity gradient was sufficiently marked to overcome the effect of temperature on the transmission of sound. While the distribution of salinity often plays an important part in determining the stability of the water column, its influence on sound is so slight that it can be safely neglected.

9) The rather complex factors determining the distribution of temperature within a few fathoms of the surface have been discussed, but it is impossible yet to demonstrate how important such small scale thermal structure is to the problem of sonic ranging.

CHAPTER V

RANGE AND INTENSITY OF THE DIRECT BEAM

1. RANGE IN AN ISOTHERMAL LAYER

The most frequently occurring condition of the water is that in which the upper layer is isothermal, or in other words is completely mixed and therefore has constant temperature and salinity. From Figure 1 it is seen that in this case the velocity increases with depth at the rate of 0.0182 ft/sec/ft. As explained on page 16, this will introduce refraction, bending all sound rays into the form of circles with centers approximately 4900/.0182 or 269,000 ft. above the surface. By reference to Figure 5 it is seen that the limiting range of the direct beam is determined by the depth of the layer. The graph in Figure 29 is designed to permit rapid calculation of this range in an isothermal layer of known depth.

Let h_1 and h_2 represent the heights of the projector and target respectively above the bottom of the isothermal layer. From Figure 29 it is possible to read values of X_1 and X_2, the horizontal distance from the vertex of the limiting ray to the projector and to the target. The range is the sum of X_1 and X_2.

For example, the maximum range for the direct beam in a layer of thoroughly mixed water 300 ft. deep, with the projector depth 13 ft. and the target depth 100 ft., is found as follows:

$$h_1 = 300 - 13 = 287 \text{ ft}; \quad X_1 = 4220 \text{ yd.}$$
$$h_2 = 300 - 100 = 200 \text{ ft}; \quad X_2 = \underline{3510} \text{ yd.}$$
$$\text{Range} = 7730 \text{ yd.}$$

2. RANGE IN A LAYER WITH GIVEN VELOCITY GRADIENT

For cases in which the velocity gradient is different from the isothermal value, the more general graph in Figure 30 may be used. The significant quantities in the calculation are the speed of sound at the vertex of a path V, the speed $V - \Delta V$ and the inclination Θ of the path at a point whose vertical distance is h and whose horizontal distance is X from the vertex. In figure 30, Θ and also the ratio X/h are plotted against values of the ratio $\Delta V/V$. The graph is drawn in four parts to permit coverage of the entire useful range of values, as follows:

Part	$\Delta V/V$	X/h	Θ
A	.01-.16	3.25-15.5	5.5-30.
B	.001-.016	10.5 -35.	3.0-11.5
C	.0001-.0016	32.5 -155.	.55-3.0
D	.00001-.00016	105. -350.	0-1.25

Since all the quantities plotted in Figure 30 are ratios, it follows that this graph will serve for calculations in any system of units. The units used may be chosen at will, provided X and h are expressed in the same unit, and likewise ΔV and V. (ΔV is read as "delta V", and means "the change in V"). In every calculation made with the aid of Figure 30, the path must be broken up into units such that one end of each unit is

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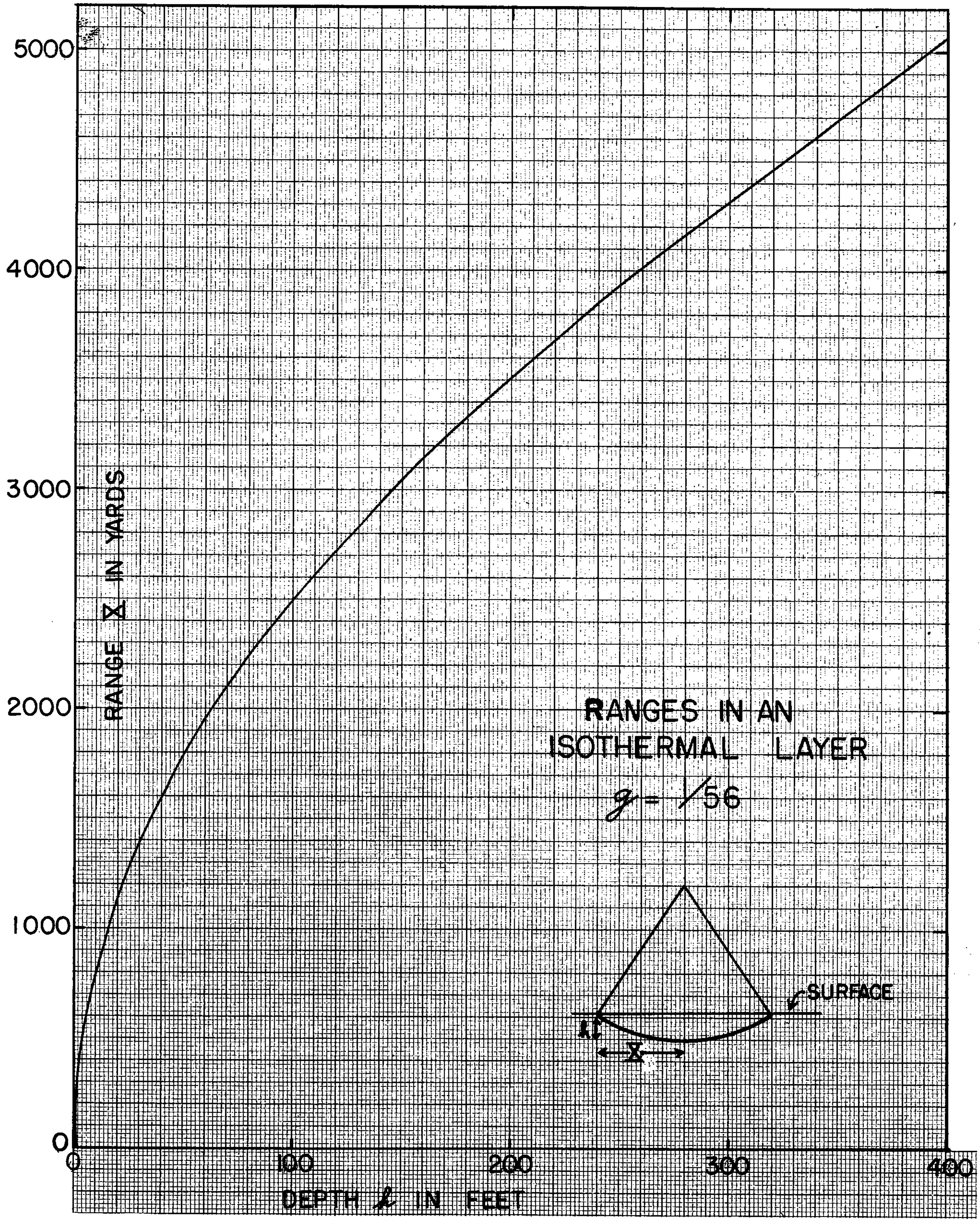

Fig. 29. RANGES IN AN ISOTHERMAL LAYER. Wnen the upper layer of water is thoroughly mixed so the temperature and salinity are constant throughout, this curve may be used to compute ranges of direct rays.

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a vertex, or point at which the path is horizontal.

The following examples illustrate the use of Figure 30 in the calculation of ray diagrams.

Examples

Negative gradient, single layer. A projector is 15 ft. beneath the surface. The sound speed decreases uniformly from 4800 ft/sec at the surface to 4790 ft/sec at a depth of 50 feet. Find (1) the distance at which the shadow zone begins at the surface, (2) the angle at the projector of the limiting ray, (3) the beginning of the shadow zone for a receiver also at the depth of 15 feet, (4) the beginning of the shadow zone for a receiver at 50 feet.

Solution:

(1)
$\Delta V/V = (15 \times 10)/(50 \times 4800) = .000625$
From Figure 30, $X/h = 56.6$
$X = 56.6 \times 15 = 849$ ft. $= 283$ yd.

(2) From Figure 30, $\theta = 2^{\circ}02'$

(3) $2 \times 283 = 566$ yd.

(4) $\Delta V/V = 10/4800 = .00208$
From Figure 30, $X/h = 31.0$
$X = 31.0 \times 50 = 1550$ ft. $= 517$ yd.
Distance $= 283 + 517 = 800$ yd.

Negative gradient, vertex unknown. In Figure 6, take the velocity to be 5000 ft/sec at the surface, and 4995 ft/sec at the bottom of the first layer, which is 50 ft. deep. Calculate the horizontal distance travelled in the layer by the ray which leaves the projector at an angle of 5° below the horizontal.

Solution: Here the position of the vertex is not known and the first step is to determine it. In Figure 30, at $\theta = 5^{\circ}$, $\Delta V/V = .00380$. Hence ΔV = approximately 5000 x .00380 = 19.0 ft/sec. The velocity gradient is 5/50 = .100, so a change in V of 19.0 ft/sec means a change in h of 19.0/.1 = 190 feet. Thus the vertex of the path is 190 ft. above the projector, which happens to be above the actual surface of the water as shown in Figure 31. This point causes no trouble because the vertex lies on a continuation of the ray back of the projector, which is used only as an aid to the calculation. From Figure 30, at $\Delta V/V = .00380$, we find $X/h = 22.9$, so $X = 190 \times 22.9 = 4350$ feet. The vertex of the required path is 190 ft. above the projector and 4350 ft. back of it. The velocity at that point (if the layer actually extended to it) would be

$$5000 + (175 \times 5 \div 50) = 5017.5 \text{ ft/sec.}$$

Having located the vertex, the next step is to calculate the horizontal distance for a ray from the vertex to travel to the required depth of 35 ft. below the projector or 225 ft. below the vertex. For this part of the calculation

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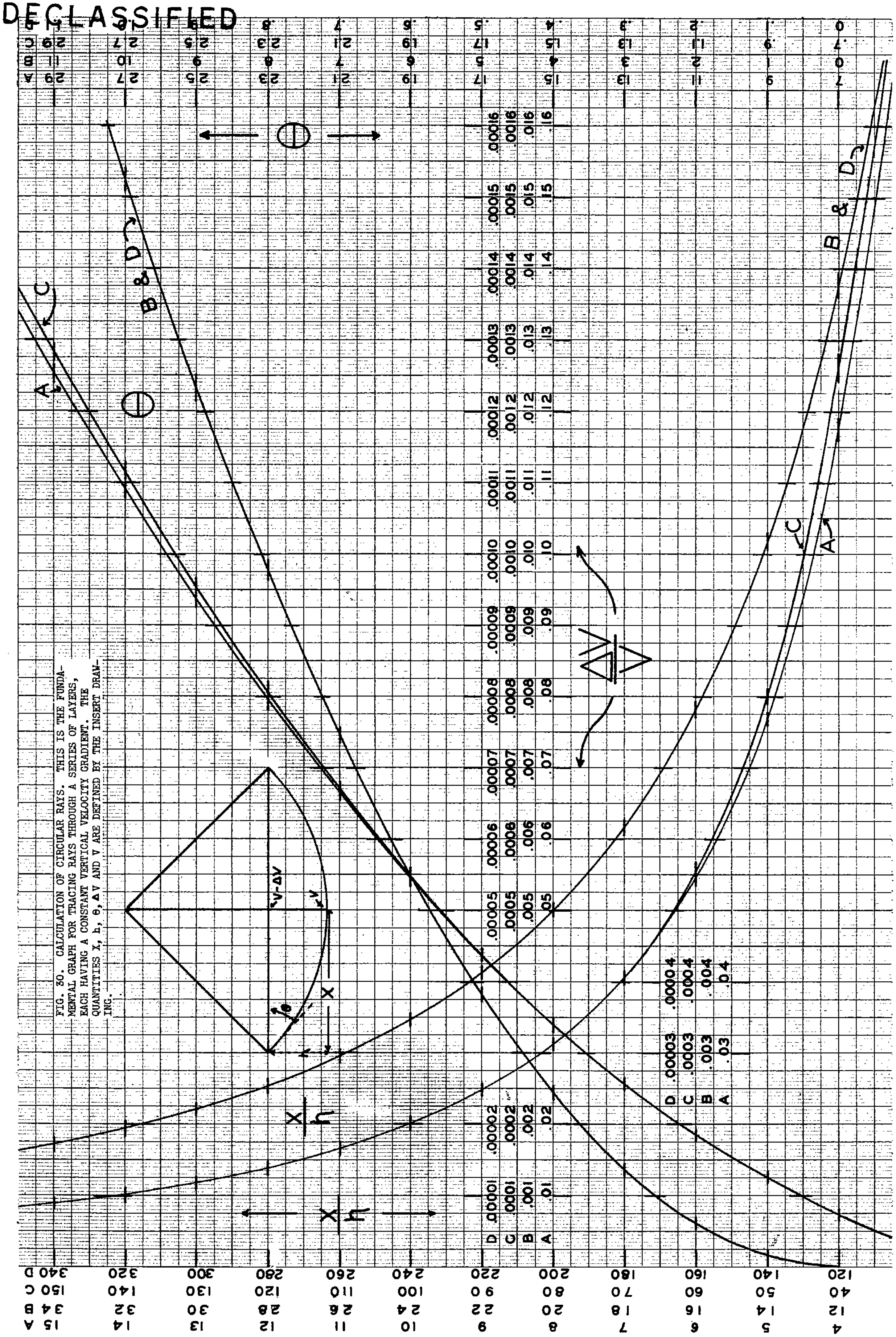

FIG. 30. CALCULATION OF CIRCULAR RAYS. THIS IS THE FUNDAMENTAL GRAPH FOR TRACING RAYS THROUGH A SERIES OF LAYERS, EACH HAVING A CONSTANT VERTICAL VELOCITY GRADIENT. THE QUANTITIES X, h, θ, ΔV AND V ARE DEFINED BY THE INSERT DRAWING.

$\Delta V/V = 22.5/5017.5 = .00449$
From Figure 29, $X/h = 23.8$
$X = 23.8 \times 225 = 5360$ ft.
The required distance is 5360 - 4350 = 1010 feet.

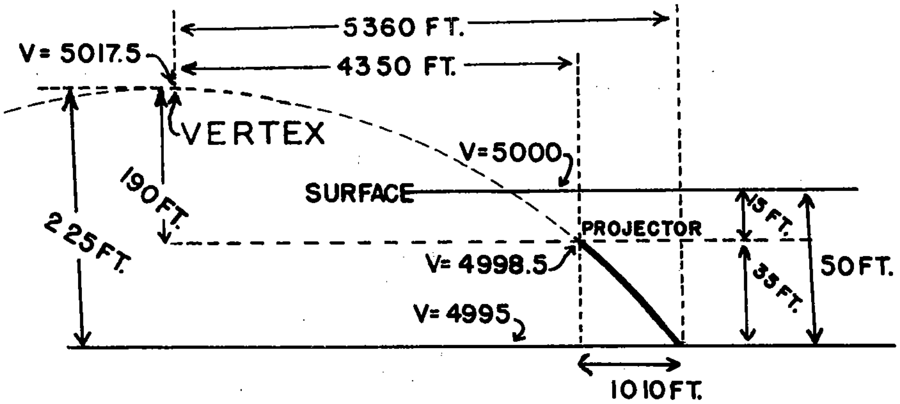

Fig. 31. SOLUTION FOR UNKNOWN VERTEX. Illustrating the solution of an example solved in the text.

3. GEOMETRICAL DIVERGENCE OF THE SOUND BEAM

In the complete absence of speed gradients, surface reflections, and absorption, the rays of a sound beam would all be straight lines and the decrease in intensity of the sound with increasing distance from the projector would result solely from <u>geometrical divergence</u>. The total sonic energy of the beam would be the same at any distance from the projector, but it would be spread over a larger area at a greater distance. The area covered by the beam increases as the square of the distance, so the intensity (amount of sound energy per unit area per second) decreases as the square of the distance.

It is customary to measure the intensity of sound in decibels above or below an intensity which has been selected as a standard of reference. If I_1 and I_2 represent the intensities at two points in a sound beam, the number of decibels denoting their ratio is $10 \log_{10} I_1/I_2$. The following table shows the relation between decibels and intensity ratios:

Intensity ratio	Decibels
1	0
2	3.0
3	4.8
4	6.0
5	7.0
6	7.8
7	8.5
8	9.0
9	9.5
10	10.0
100	20.
1000	30.
10,000	40.
100,000	50.
1,000,000	60.

If intensity is measured in decibels, and if the variations in intensity of the beam with changing distance are due to geometrical divergence alone, the variation of intensity with range is given by the expression

$$db = (db_o + 20 \log R_o) - 20 \log R \quad \ldots\ldots\ldots\ldots\ldots \quad (3)$$

where dbo represents the value chosen for representing the sound intensity at the range Ro. To apply this equation for the purpose of finding whether effects other than geometrical divergence are influencing the intensity in the sound beam it is necessary to select the reference range Ro small enough that reflection and refraction effects are negligible.

In Figures 33 to 39 are shown typical intensity versus range measurements for several experiments conducted by the Naval Research Laboratory and others. Signals sent from a projector at a depth of 13 feet were detected and their intensity measured by a unit at an equal depth and at varying distance. Simultaneous measurements of the thermal condition of the water were made and from them the refraction pattern was calculated. In Figure 33, where the refraction did not cause shadow zones, the variation of observed intensity agrees with the values computed from geometrical divergence. But in Figures 34, 35, 36 and 39, cases where refraction does produce shadow zones, the observed intensity falls far below that computed on the basis of geometrical divergence alone.

The question may arise whether the curvature of the sound beam due to refraction may not cause the actual distance travelled to exceed the straight line distance between source and target. This increase is so small that it can be ignored in computing intensity changes due to geometrical divergence, as will readily become apparent if the ray diagrams are constructed on a true scale without vertical exaggeration.

4. REFLECTION OF SOUND

Reflection From The Surface

Underwater sound is reflected very efficiently from the surface of the water, less efficiently from the bottom, and slightly by abrupt changes in water conditions inside the body of water. The effectiveness of sound reflected from the surface is greatly diminished by the presence of waves, and it is a fair comparison to consider that the difference between the direct sound beam and a beam once reflected from the surface is similar to the difference between the direct sunlight and the sunlight reflected from the surface of the water under the existing condition of the sea. When a ray is reflected from the horizontal surface of the sea, the ray is inclined to the horizontal at the same angle before and after reflections; if the incident ray is 1° above the horizontal the reflected ray will be 1° below the horizontal. But if the surface is inclined at an angle θ° by the presence of a wave, the reflected ray will be deflected through an angle $2\theta^\circ$ from its course as determined by the direction of the incident ray. Thus each ray of the direct beam may be considered to give rise to a sort of cone of reflected rays, the angle of the cone being twice the maximum angle of disturbance of the surface by the wave motion.

Figure 32 illustrates the distribution of surface-reflected sound in the case of (A) positive and (B) negative vertical gradients of sound speed. Five conclusions are readily drawn from these diagrams:

(1) Within the limit of range of the direct rays the received sound always consists of reflected sound in addition to the direct sound.

(2) When the gradient is positive the range of the reflected rays exceeds that of the direct rays, but for negative gradients this is not the case. We would therefore expect a more marked decrease in intensity at the limit of range of the direct beam in the case of the negative gradient than in the case of the positive one.

(3) The direct rays just inside the limiting range will be weaker in negative gradients than in positive ones because these rays pass near the surface and are subject to disturbance by wave action and related surface effects.

(4) Because they arrive over a number of different paths, reflected sound signals will be "mushy" in regard to both direction and time of arrival. The problem of taking a bearing on a signal whose path includes a reflection from the surface may be compared to taking a bearing on the sun by observing the light reflected from the surface of the sea.

(5) Surface-reflected sound will obviously be strongly influenced by the amount of sea and swell.

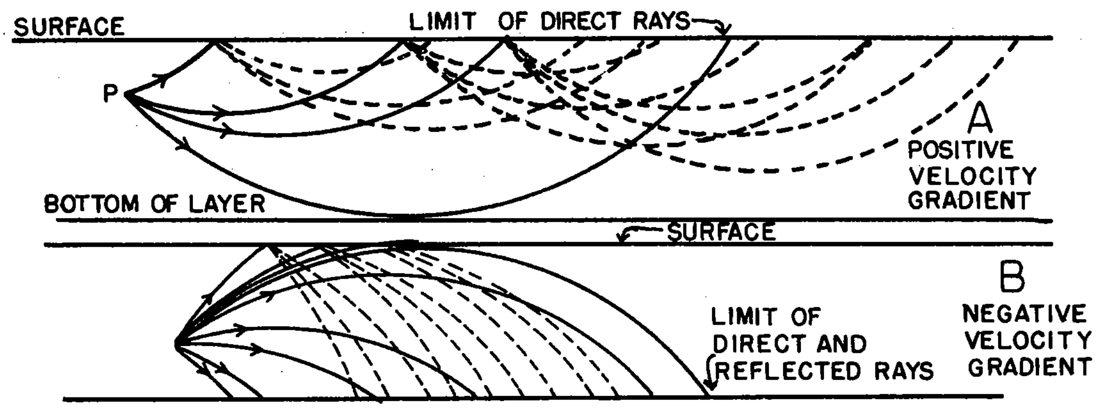

Fig. 32. SURFACE-REFLECTED RAYS. When the surface layer has a positive gradient surface reflections are more effective in sending sound into a shadow zone than when the gradient is negative. The multiple reflected rays arising from each direct ray represent the scattering action of water waves.

Reflection from Bottom

Reflection from the bottom is well known through sonic depth measurements, which show a considerable variation in the reflecting power of the bottom materials. In sound ranging work, incident rays meet the bottom at flat angles (less than 15^{o}), the reflection may be expected to be more efficient than for vertical rays. The ray diagrams for effects of refraction on bottom reflection may be obtained from Figure 32 by simply imagining the figure inverted. There are not available for this report data on the results of sound ranging in shallow water in which other conditions are known with sufficient accuracy to permit isolation and evaluation of the effect of bottom reflection, but it is probable that sound reflected one or more times from bottom plays an important part in operations inside the 100 fathom curve. This point demands further study.

[redacted]

Reflection At Interfaces Within The Water

Several cases in which reflection occurs from an interface between two kinds of water are well known. Reflection from the wake of a ship is due to the persistence of turbulence in the wake, or to the vertical mixing in stratified water, or to both. Reflections which appear to come from the interface between two types of water have frequently been reported near the edge of the Gulf Stream. Supersonic sounding machines when operated near the mouths of the Mississippi have given echoes from the interface separating the overlying river water from the saltier water beneath. Echoes have been reported from schools of fish, arising apparently from the turbulence produced by motion of the school. In general there will be partial reflection of a sound beam whenever it encounters an interface separating two bodies of water which differ in sound velocity, density, temperature, current velocity, or degree of turbulence. The greater the contrast and the sharper the transition, the greater will be the reflecting power.

Until the bathythermograph, or other instrument for measuring water conditions continuously instead of at a set of points, becomes more available it will not be known how commonly interfaces sharp enough to cause reflections occur, or how sharp an interface is required. There are some instances of intensity-range runs which seem to require reflection at an interface to explain the observed intensities. For example, in the ray diagram of Figure 34 there is no chance of getting energy of the direct beam to a detector located between the ranges of about 900 and 4500 yards. The energy which does appear between these ranges is slight but definitely present. Either it came by partial reflection at the interface at 26 ft.depth or by some process not yet discussed in this report, such as diffraction. To obtain definite information on this point will require experiments planned especially for that purpose.

5. DIFFRACTION, SCATTERING, AND ABSORPTION OF SOUND

Diffraction

Any change in the direction of travel of sound waves caused by bending around an obstacle, instead of by variations of the water, is called <u>diffraction</u>. Diffraction is the phenomenon which permits ordinary sound in air to travel around a corner. In a layer of water in which the sound rays follow curved paths, the boundary of the layer on the high speed side may be considered as an obstacle which casts a "sound shadow". Because of diffraction effects the boundaries of shadow zones, such as these shown in 5 and 6, are not absolutely sharp lines. Sound will, to a certain extent, spread into the shadow zone. The extent of such spreading, or in other words the "fuzziness" of the shadow, will be greater the lower the frequency or the longer the wave length of the sound wave used. It is the fact that light waves are very much shorter than sound waves which accounts for the sharper shadow cast by a given obstacle for light than for sound. Thus the 17.6 kilocycle signal penetrates better than the 23.6 kilocycle signal into shadow zones largely because of diffraction effects. For sound in the lower part of the range of audible frequencies it is probably that diffraction would eliminate the shadow zones.

Scattering

The disturbance of sound transmission by irregularities in the properties of the water is called scattering if all linear dimensions of the irregular areas are small compared to the dimensions of the sound beam. For instance, a layer of water 20 ft. deep but extending laterally for several miles is considered to refract the sound but a body of water whose greatest dimension is of the order of 20 ft. would be considered as scattering the sound. Scattering tends to diffuse the beam in a random way while refraction produces a systematic bending of the beam.

Scattering accompanies reflection at the surface or at the bottom provided the surface is disturbed by waves or the bottom is irregular. When there is a zone of turbulence near the surface due either to wave action or to circulation induced by cooling of the surface with resultant sinking of the cooled water, this scattering will affect the range of direct rays which recurve near the surface as those in Figure 34, page 64.

It is likely that the occurrence of excessive reverberations results from scattering of this type, and observations of the water conditions under which reverberations occur are needed.

Absorption

The term absorption has been used in the past in two different connections with underwater sound. True absorption means the loss in energy in the sound beam due to viscosity and imperfect elasticity of the water. This effect is negligible for all frequencies in practical use, and if it alone were responsible for the loss in signal strength the ranges obtained with present equipment would be far greater than are now found even under the best conditions. ?

The absorption is sometimes used to mean the loss of intensity with increasing distance due to the combined effects of true absorption, scattering, reflection, refraction, geometrical divergence, diffraction, etc. When used in this sense it should be called the loss coefficient. It is measured in db/kiloyard, and does not bear any simple relation to water conditions which would permit its calculation from water temperatures without use of the refraction calculations.

In the past there have been attempts to correlate high loss in signal strength with absorption induced by small organisms in the water. The evidence is against this view for in every case where temperatures and salinities have been measured they have been found to give the explanation of the sound conditions observed. The fact that organisms are exceptionally plentiful in some regions where sound conditions are notably bad is due to the coincidence that the particular set of water conditions present in these regions are favorable for the growth of the organisms and unfavorable for the transmission of sound.

6. DATA FROM SOUND TRANSMISSION EXPERIMENTS

Sources of Data

Four series of measurements of intensity vs. range have been made by the SEMMES in collaboration with various other vessels. During the intensity measurements complete data on water temperatures to depths of several hundred feet were observed. These data are of the greatest value because they offer an opportunity for applying the calculations of refraction and for comparing results predicted by the refraction calculations with the intensities which were actually observed.

In these experiments the SEMMES usually approached the vessel which acted as target for its sound beam, keeping its projector trained on the target. The target vessel kept its receiver trained on the approaching SEMMES and measured the intensity of received sound. The receiver was usually located at a depth of 14 feet.

Presentation of Data

General Statement. The data are presented in Figures 33 to 39. Four intensity versus range runs, made under similar water conditions, are shown in each figure. For each run a theoretical intensity curve is drawn representing intensities calculated on the assumption that geometrical divergence is the sole controlling factor. In many cases the observed points fall below the smooth curves, indicating that factors other than geometrical divergence must be considered.

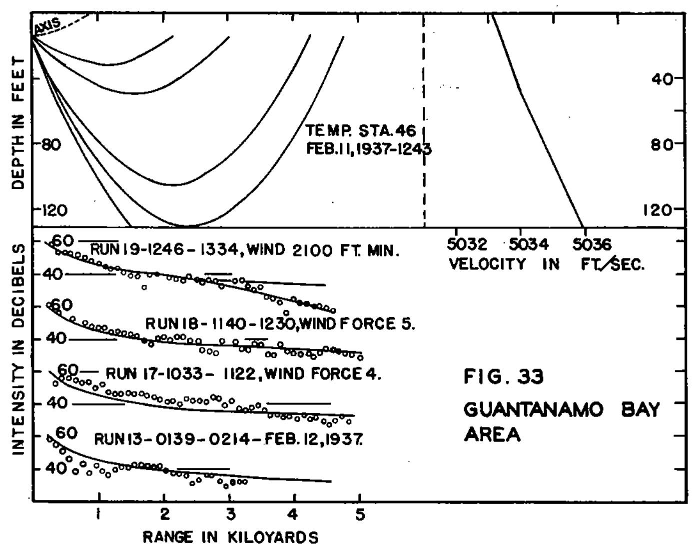

FIG. 33
GUANTANAMO BAY AREA

Each figure also contains a graph of sound velocity vs. depth and a drawing of the paths taken by various rays of the sound beam under the influence of the given velocity distribution. To permit ready intercomparison the speed chart has a common vertical scale with the ray diagram, and the ray diagram has a common horizontal scale with the intensity curves. Each speed chart, and the resulting ray diagrams, is selected as characteristic for the group of intensity curves which are included in the same figure with it.

Each speed diagram was constructed from given temperature data by the use of Figure 1. Each ray diagram was constructed from the speed diagram by the use of Figure 30. Each inverse square intensity curve was drawn from equation 3, page 58. After practise, one can construct one of these complete diagrams in 30 to 50 minutes. When the entire process is well understood it is possible to make reasonably accurate predictions of sound ranging performance from the velocity chart alone, but it is desirable to construct the ray diagram if possible.

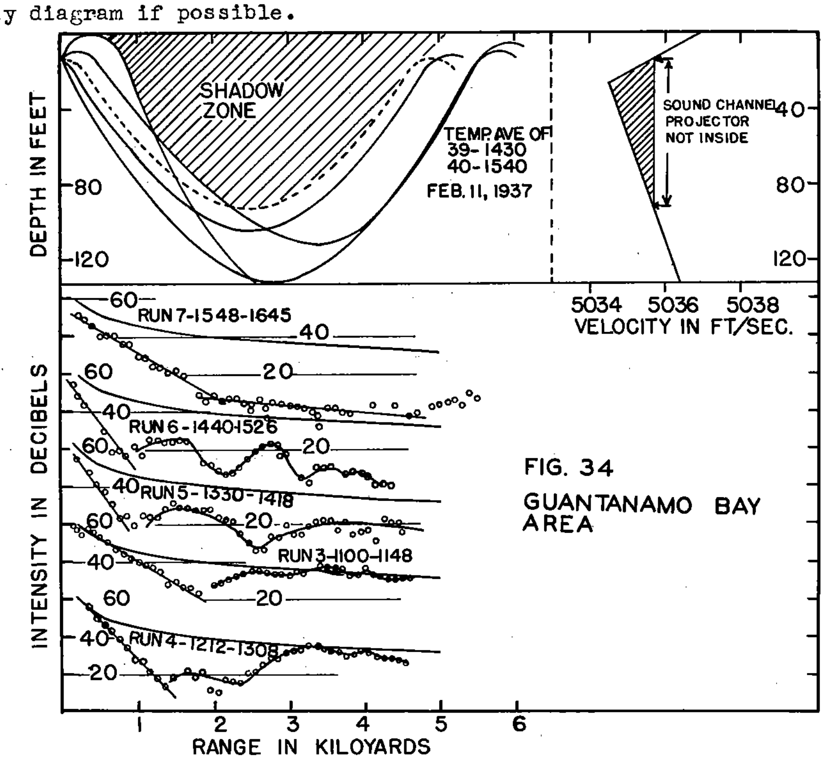

FIG. 34
GUANTANAMO BAY AREA

The absolute intensities on the decibel scales in Figures 33 to 39 are somewhat arbitrary. The sound intensity data were collected on three cruises, and it is assumed that the zero intensity remained constant throughout each cruise, but has a different value from one cruise to the next. The smooth curves for intensity are put in at a constant level for each cruise, the level being selected to average all of the short range points for the cruise.

<u>Figure 33, Feb. 11-12, 1937. Guantanamo Bay Area</u>. Here the speed increases with depth to the greatest depth measured, giving ideal conditions for sound ranging as shown by the ray diagram and the intensity curves. An appreciable amount of surface-reflected energy probably reaches the receiver, but the wind was force 4 to force 5 and there was probably enough sea to scatter the surface reflections badly.

The good sound conditions are attributed to the fact that the wind had thoroughly mixed the surface layer.

Figure 34, Feb. 11, 1937. Guantanamo Bay Area. This situation, by contrast with that of Figure 33, shows the adverse effects produced by heating of the top 30 feet of water, which caused a negative velocity gradient at the surface. The intensities fall off rapidly and at a range of only 1000 yards the receiver is shadowed from the direct beam. Some of the intensity runs show the intensity returning to the theoretical curve at longer ranges, in general agreement with the ray diagram. Because of the negative gradient in the top layer surface reflections are not possible as may be seen from the ray diagram. Energy can reach a receiver in the shadow zone only by diffraction or reflection at one of the interfaces, and these processes are not efficient.

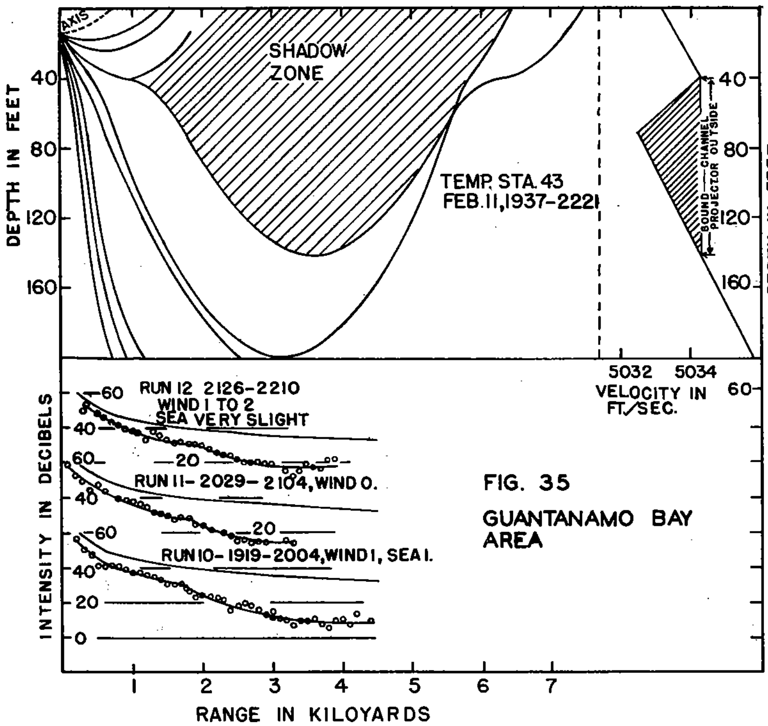

FIG. 35

GUANTANAMO BAY AREA

Figure 35. Feb. 11, 1937. Guantanamo Bay Area. The limiting range for direct rays is about 1900 yards as seen from the ray diagram. This agrees with a break in the curve of observed intensities which appear on all three runs. The ranges were not extended far enough to show the effect of direct rays returning to the surface at about 7000 yards. The surface has a positive gradient, which is the situation favorable for surface reflections, and the wind and sea were slight, providing best conditions for surface reflections. Thus the surface-reflected waves carry well into the shadow zone with usable intensity.

Figure 36. Guantanamo Bay Area. Feb. 11, 1937. The range of the direct beam is only about 700 yards. The positive gradient in the surface layer is favorable for surface reflections but this layer is so thin that it does not carry the sound effectively. If a ray is deflected slightly by wave action during reflection it may escape the surface layer

and become lost. The speed at all points below the projector level is less than at the projector, with the consequence that every ray which starts downward fails to return to the surface.

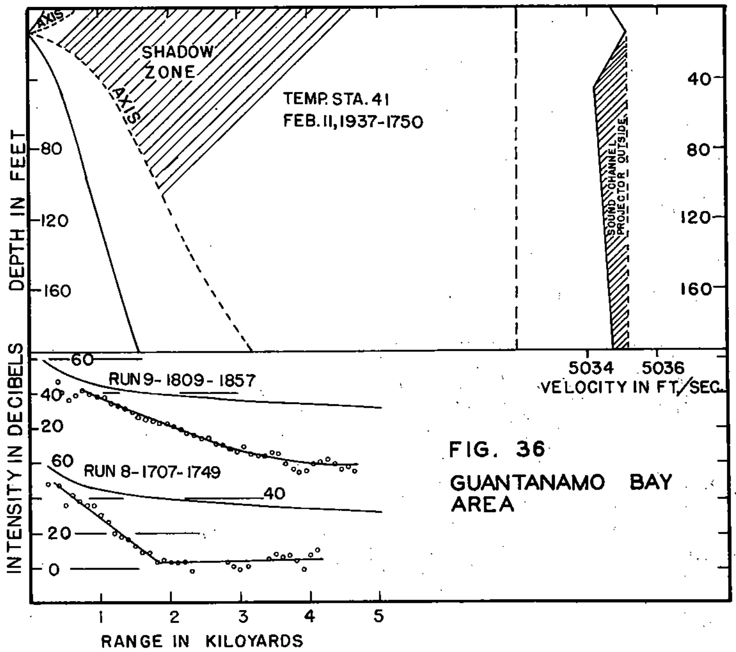

FIG. 36
GUANTANAMO BAY AREA

Figure 37. South Atlantic-Caribbean Area, March 9-10, 1938. These are ideal sound conditions, the wind having thoroughly mixed the water to the maximum depth involved in the transmission of sound. The observed intensities parallel the theoretical ones (based on the assumption that geometrical divergence is the sole factor influencing intensity) in practically every case. As seen from the ray diagram, the direct rays at range 5000 yd. penetrate to a maximum depth of 120 ft. which is the greatest depth to which temperatures were measured.

The fact that the theoretical curves are uniformly somewhat above the observed ones is not significant, for the choice of the level at which the theoretical curves were drawn was arbitrarily selected for each expedition.

Figure 38. Mar. 10, 1938. South Atlantic-Caribbean Area. Here the wind has fallen and the sun has heated the surface water, giving a stable surface layer 30 ft. deep which disturbs the good sound conditions existing shortly before, as shown in Figure 36. There is a shadow zone extending from about 400 yards to about 4500 yards shown in the ray diagram. This diagram was drawn from data about the water conditions as measured between the runs 32 and 33, and the predicted ranges would probably be slightly different for runs 31 and 34.

Energy does not spread into this shadow zone by surface reflection because the negative velocity gradient at the surface prevents effective surface reflections. The

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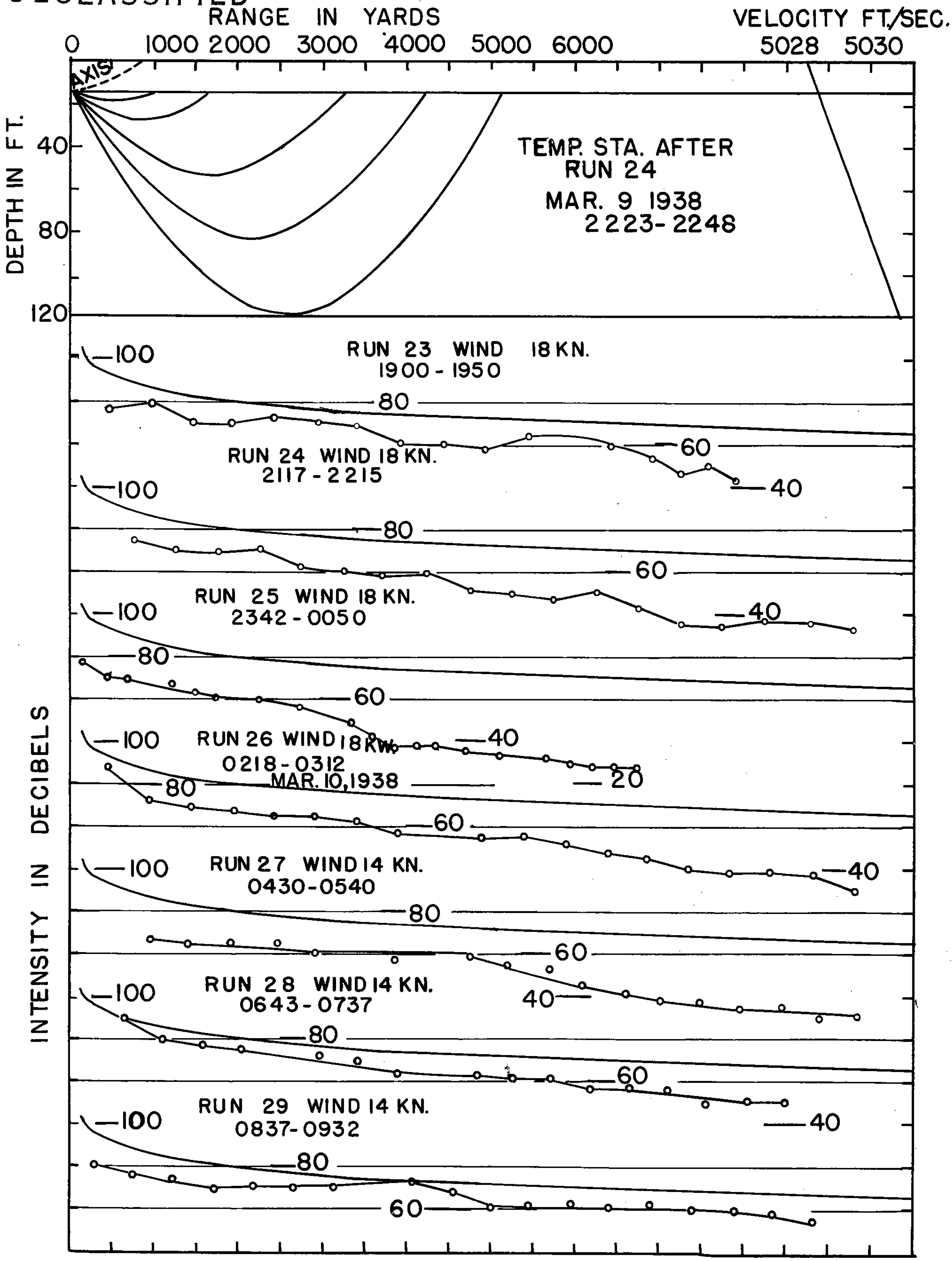

Fig. 37. SOUTH ATLANTIC-CARIBBEAN AREA.

observed intensity thus falls sharply in the shadow zone but returns to the theoretical value for ranges beyond it. The fact that the observed intensity rises higher just beyond the shadow zone than the intensity at equal ranges under the good sound conditions indicates reenforcement by the energy deflected away from the shadow zone, as can be seen from the ray diagram.

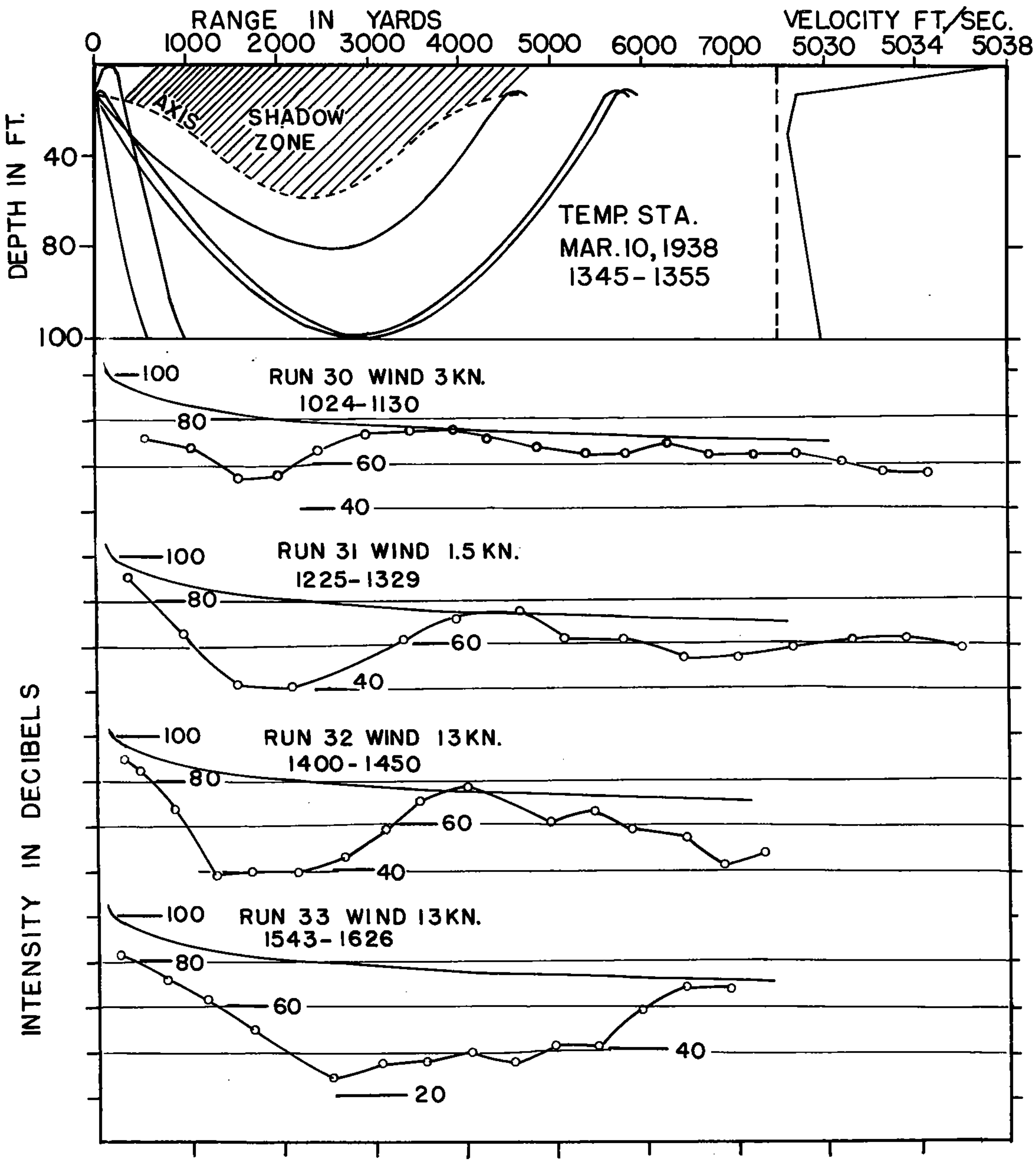

Fig. 38. SOUTH ATLANTIC-CARIBBEAN AREA

<u>Figure 39, Feb. 28-Mar. 9, 1939. Pacific Areas A, B and C.</u> Twenty six intensity runs and temperature measurements were made in the Pacific. These were located in the trade wind belt a short distance beyond the 100 fathom curve. These intensity runs, as shown in Figure 39, are divided into 5 groups, indicated by the letters A to E, according to the type of thermal conditions present. A typical thermal curve is shown with each

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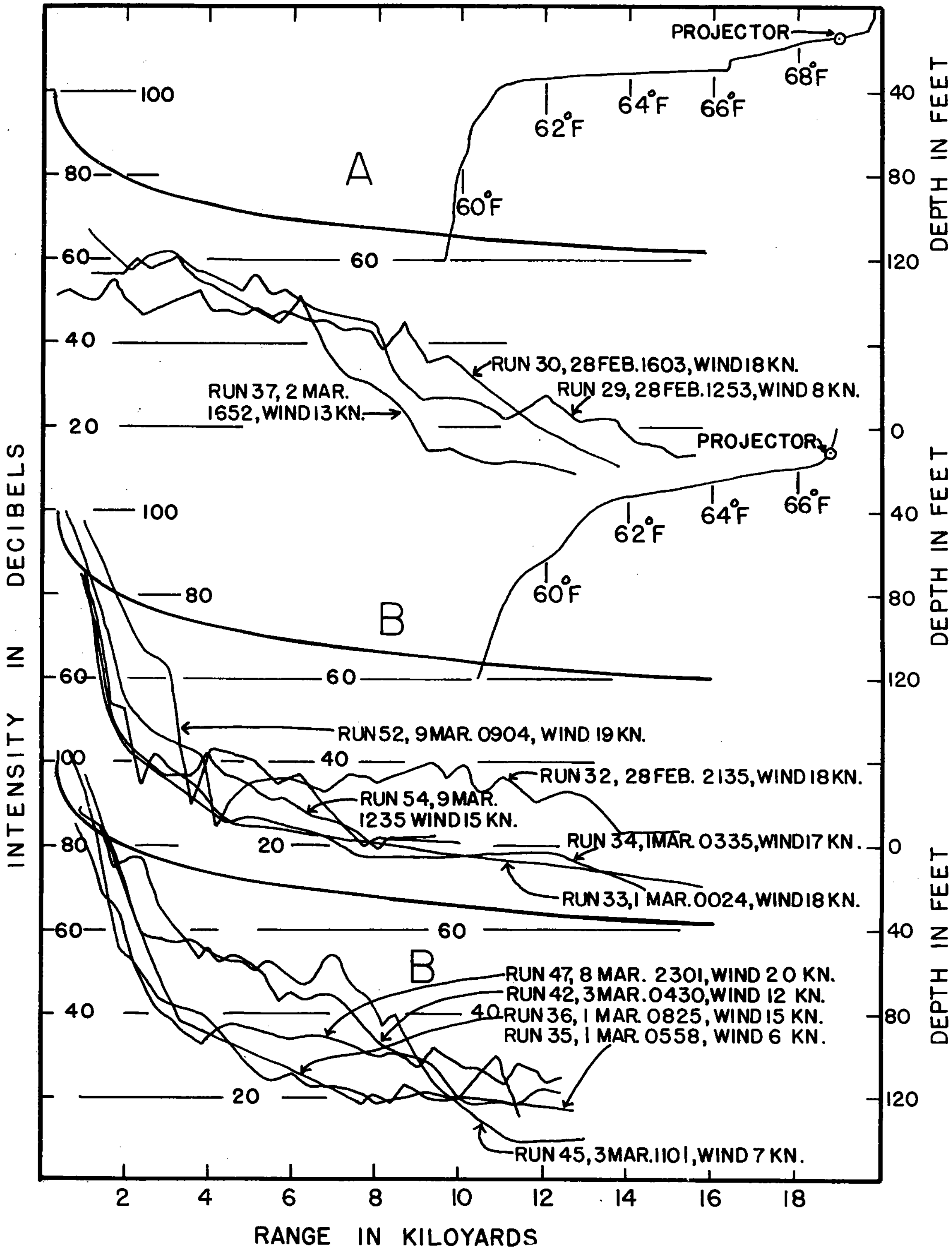

Fig. 39 A and B. PACIFIC AREA

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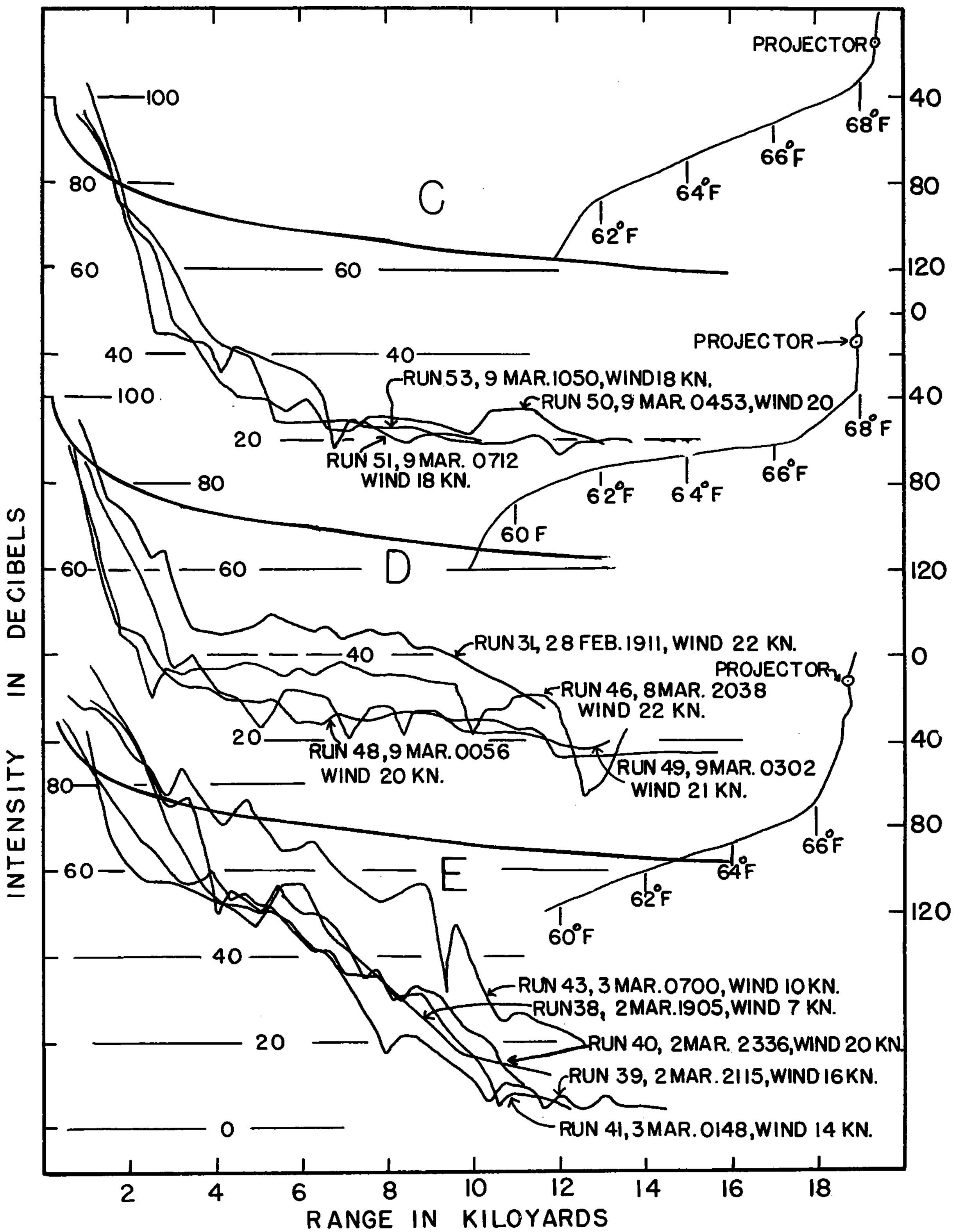

Fig. 39 C, D, and E. PACIFIC AREA

group of intensity curves. The hour and the wind speed are indicated for each intensity curve.

All thermal curves of Figure 39 are notable for the great drop in temperature near the surface. It sometimes amounts to 10°F in 80 feet, which would give a change in sound speed of about 50 ft/sec. The magnitude of this change will be apparent when it is recalled that the speed changes in all cases previously described in this report were usually about 4 ft/sec or less. Another unusual feature of the thermal condition of the water in this area is that the trade wind, which averaged about 15 knots during the tests, did not produce mixing of the water to appreciable depths. As explained in Chapter IV, page 51, upwelling of the cold deep water induced by the offshore wind accounts for all features of this thermal situation. In other seasons of the year the thermal conditions would be radically different.

As one would expect, these violent changes in sound velocity with depth are very unfavorable for horizontal sound transmission. Any ray which gets into the region of steep thermal gradient is bent downward so sharply that it never returns to the surface. In each thermal curve there appears a small surface layer in which the temperature is approximately constant, and this layer is the only part of the water available for sound ranging on a surface target. The intensity runs have been divided into groups A to E according to the depth of the mixed surface layer. All observed intensity curves fall far below the theoretical curve but it will be seen that group A, in which the mixed surface layer is thinnest, is worst, and that the results generally improve toward group E in which the mixed layer is thickest.

In group A, the mixed layer does not even extend to projector depth, and the recorded intensity does not rise to the theoretical curve even at the shortest ranges. Groups B to E show an increase in thickness of the mixed layer, and gradual increase in the range of signals for a given intensity. The range at which the level falls to 40 db. is in gradual agreement with the predictions made by the methods outlined earlier in the report. It is not clear how the signals at levels below 30 db. can continue to come in at such great ranges in spite of the large thermal gradients. This point is not of immediate importance because these signals are too weak to give useable echoes.

7. PREDICTION OF THE MAXIMUM RANGE

To predict the maximum range at which echoes may be obtained, it is necessary that the following items be known:

a) The variation of velocity of sound with depth. In most cases the salinity is sufficiently constant to permit determination of the velocity distribution from thermal data alone. The thermal data may be obtained by use of the bathythermograph.

b) The power of the projector and the maximum sensitivity at which the receiving circuit may be operated. The power of the projector is perhaps best specified as the level of sonic energy in decibels dbo in the direct beam at some distance Ro small enough that refraction, surface reflection, etc. have not influenced the signal strength.

c) A graph showing the variation of signal strength with distance, constructed from the formula

$$db_1 = (dbo + 20 \log Ro) - 20 \log R_1,$$

which assumes that geometrical divergence is the only factor influencing the signal strength. This curve will be similar to the smooth intensity curves in Figures 33 to 39.

d) Knowledge of the signal strength which must exist at the target in order that a measurable echo may be obtained for it. It is stated by Stephenson (NRL Report No. 3-1670, page 11) that "for the equipment and technique used by the SEMMES and the S-20, the echo intensity from the S-20 on the surface was 60±5 db. below the level of the direct signal. At speeds under 12 knots, 0 db. was the threshold level for 17 kc. echoes on the SEMMES, so that at any time the direct signal received on the S-20 was 60 db. or more, echoes from the S-20 on the surface could be heard on the SEMMES".

It seems likely that experiments with projectors of different power and with targets of different effective reflecting areas would show that this limiting value is different from 60 db. in some cases, but until further data are available this value will be adopted. A given target may be considered to act as a source of sound whose power depends on the product of the effective reflecting area times the strength of the direct signal at the target.

e) A ray diagram constructed from the thermal data according to the methods given in this Chapter.

The procedure for predicting the range is as follows:

a) Note the range at which the 60 db.level (or a revised value of the limiting level in case this is available) is reached on the geometrical divergence curve. If no shadow zones are shown on the ray diagram inside this range, take it as the limiting range.

b) If there is a shadow zone, shown on the ray diagram, take range to be the edge of the shadow zone (for the target depth in question) plus an increase of a certain percentage of this range which is added to take account of the spread of sound into the shadow zone through diffraction, surface reflection, and bottom reflection. These percentages are as yet undetermined, but every test of the maximum range of echoes will help to determine them. As explained in Chapter III the sound will spread further into the shadow zone in situations like Figure 5 than in those like Figure 6.

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It is hoped that experiments will be made in the near future which will make it possible to give more definite statements about the points which are left indefinite in the above paragraphs.

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8. SUMMARY AND CONCLUSIONS

1. By use of Figures 29 and 30, according to the method outlined on pages 54 to 58, it is possible to draw a diagram of the sound beam from a projector at any depth. This diagram will show what parts of the potential target area lie in shadow zones.

2. The signal strength at any target in the direct beam may be calculated by assuming that geometrical divergence is the sole determining factor, which leads to the equation

$$db = (dbo + 20 \log Ro) - 20 \log R,$$

where db and dbo are the intensities in decibels at ranges R and Ro.

~~apparently wrong at the latter~~ in error by not including absorption for freq. F.R.

3. The spread of the sound beam into a shadow zone may be estimated qualitatively on the basis of the discussions on surface reflection, scattering, and diffraction.

4. Calculations based on the principles outlined in the four preceding points have been made for all data on signal intensity vs. range which were available for this report. The excellent agreement between the calculated and experimental results forms the basis for the viewpoint that refraction due to vertical gradients of sound speed is the (major?) controlling factor in sound ranging work.

5. Although seasonal and regional charts of water temperature provide very useful information, it is desirable to have temperature measurements made on the spot if detailed predictions of effective range of sound apparatus are required.

6. The bathythermograph is the instrument most suitable for measuring water temperatures for this purpose. The instrument will function at speeds up to 14 knots, and the predictions can be made within 30 minutes after the instrument is operated.

7. A method is put forward for predicting the maximum distance at which echoes may be obtained. Some details of this method need further experiments and extensive checking against the results of actual tests, but it is hoped that these will be forthcoming in the near future.

CHAPTER VI

VARIATIONS IN THICKNESS OF THE WIND STIRRED LAYER

1. GENERAL STATEMENT

The need for charts showing the thickness of the wind stirred layer at the surface is only too clear from the preceding discussion. Of course such charts could only show the mean monthly conditions, nevertheless the seasonal and geographical changes are sufficiently marked so that from the tactical point of view they far outweigh the minor day to day variations imposed by the local weather. Were such charts available, both surface and subsurface craft could plan their courses to remain as far as possible in good or bad sound water as the need might be. Ships employed as a sound screen would know in just which areas a submarine would be most likely to slip through undetected. On the other hand, a submarine, when attacking sound protected surface craft, could plan the place and time of her approach so as to remain as far as possible in the shadow zone. In short, both from the defensive and the offensive standpoints a knowledge of the thickness of the wind stirred layer at the surface would be of great advantage.

It seems likely that for tactical purposes two sets of charts will suffice: one group showing the average depth of the homogeneous surface layer during the 4 winter months (Nov. 15-March 15) when the range is at a maximum and another for the summer months (May 15-Sept. 15) when the range of the direct ray is greatly reduced by the secondary thermocline. For the transition months of spring (March 15-May 15) and autumn (Sept. 15-Nov. 15) the range should be intermediate and roughly predictable through a knowledge of the seasonal temperature cycle in the surface layer. These dates of course apply only to the northern hemisphere.

Such charts would show the average thickness of the wind stirred layer from region to region when a steady wind prevails. In calm weather or with a light variable wind the diurnal effect might greatly reduce the expected range during the middle part of the day, but this factor will perhaps be predictable from daily weather maps when it has been further examined. In short, from our present knowledge it does not seem out of reason to imagine that before long a commander will be able to estimate the sound transmitting qualities of distant waters towards which he is approaching.

In the hope of constructing some preliminary charts all published oceanographic observations have been examined, but for the most part these data are unsatisfactory. In offshore waters it has been the custom in oceanography to observe the temperature only at the surface and at 50 meter depth intervals. At perhaps half the stations an additional reading is available at a depth of 25 meters. Except in winter and in the heart of the trade wind belts, such wide depth intervals give little information concerning the thickness of the homogeneous layer,for in many circumstances only 5-10 fathoms of mixed water can be expected to overlie the secondary thermocline. The relatively few summer stations where temperature has been observed at sufficiently close depth intervals do not permit the charting of any considerable offshore area.

Unfortunately, most oceanographic expeditions have been at sea during the summer months. Winter observations have been made at only a few hundred points and these are practically all from the North Atlantic. A few South Atlantic data are available for winter, but almost none from the central Pacific. The result is that only a single preliminary chart can be prepared for this report and although its shortcomings are only too evident, it does give an indication of what can be expected once bathythermographs are put into use on a large scale.

2. NORTH ATLANTIC WINTER CONDITIONS

The distribution of the available observations is shown by dots on the chart in Figure 40. In areas where the data are too scattered to justify the drawing of contours, the observed depth of the wind stirred surface layer has been entered beside the station or in less doubtful cases the contour lines have been broken.

It will be seen that over the continental shelf between Cape Hatteras and Nova Scotia the water is mixed from surface to bottom with the exception of the central part of the Gulf of Maine where depths of more than 100 fathoms prevail. Just beyond the 100 fathom curve at the edge of the continental shelf the relatively fresh and cold coastal water is in contact with the slope water (see Fig. 21, Chapter IV). As already explained, at this slanting boundary the coastal water usually overlies the slope water, resulting in an increase of temperature with depth. This situation is of course favorable to horizontal sound transmission and in winter the same structure no doubt continues northeastward along the edge of the continental shelf off Nova Scotia and Newfoundland.

On the other hand, the contact between the slope water and the Gulf Stream is such that the warmer waters of the current usually overlie the colder waters north of it and this produces a narrow band of poor sound water all along the northern edge of the Gulf Stream. In all probability this condition continues northeastward (as shown by the broken line marked "north limit of Gulf Stream") as far as the current has a northward component.

At the southern edge of the Gulf Stream, where for a short distance its warm waters overlie the slightly colder surface layer of the Sargasso Sea, there is also a narrow band with some thermal stability near the surface. It is probable that this does not extend east of the Grand Banks.

In the northwestern Sargasso Sea the strong winter gales keep the water stirred to a very considerable depth. Several of the stations inside the contour line marked 100 on the diagram show mixed water down to 200 fathoms. This is probably the largest area of deeply mixed water in the Atlantic, except in latitudes higher than 55° where the winter mixing goes to depths of 500 fathoms or more.

South of Bermuda a considerable area is shown where less than 25 fathoms of homogeneous water are indicated. At first sight this is surprising unless one takes into account the horse latitude belt of moderate, variable winds which separates the westerlies from the trade winds. In this belt, even in winter, wind stirring is at a minimum and consequently it is possible that only a relatively shallow layer of mixed water is maintained at the surface.

Along the northern edge of the trades about 50-60 fathoms of mixed water are found. This good sound water gradually diminishes in thickness in a southerly direction. Two factors seem to explain this situation. In the first place, as already shown (Fig. 11B, Chapter IV) the main thermocline approaches the surface along the southern edge of the Northern Equatorial Current and secondly, the high humidity of the trade winds in low latitudes reduces evaporation (and thus surface cooling) to a minimum. Consequently there is little tendency for thermal instability at the surface to assist the winds in maintaining a deep mixed layer.

The effect of prolonged offshore winds in reducing the thickness of the surface layer can be clearly seen off the African coast, but the relatively poor sound water in the Gulf of Guinea is also explainable by the high rainfall of the doldrum belt. Equatorial rains and a northwestward flowing current seem to account for the relatively shallow mixed layer along the northeast coast of South America.

The conditions in the Caribbean illustrate another general oceanographic phenomenon. The transport of the surface water by the wind does not coincide with the wind direction. In the northern hemisphere the movement is somewhat to the right of the wind direction, while in the southern hemisphere it is to the left. Thus the easterly winds in the Caribbean give the surface layer a slight northward component and this, added to the stability resulting from the relatively fresh water which enters the southern Caribbean from the Guiana Current (see Fig. 20, Chapter IV), produces poor sound conditions in the southern part of the area.

Off the Panamanian coast the relatively poor sound water is probably caused by the pocketing of land drainage and since this area is strategically important it should be investigated in detail. Other areas with a shallow mixed layer, such as that along the south coast of Cuba, probably result from insufficient winds. In short, the general features shown by Figure 40 seem to be explainable by meteorological and oceanographic factors, although it is clear that a much closer network of observations is needed in several areas.

What can be said about the winter sound conditions in the greater part of the eastern North Atlantic where no oceanographic observations have been made? It is believed that for the most part where westerly winds prevail by midwinter at least 50 fathoms of mixed water will be found at the surface. In the northeastern area the currents are broad and weak. Thus few sharp horizontal transition zones will be found such as are present in the west. As the westerly winds blow the waters towards the coast of northern Europe the conditions should be excellent for deep mixing during the winter months. On the continental shelf off Ireland and England, and in the North Sea, the available oceanographic data show that the stirring is complete, surface to bottom. A number of stations in shallow water along the coast of Portugal indicate similar conditions further south. One station in deep water just off the Straits of Gibraltar shows at least 50 fathoms of mixed water. In short, in the latitudes of prevailing westerly winds favorable sound conditions in winter should be more widespread in the eastern half of the ocean than they are in the west.

3. SOUTH ATLANTIC WINTER CONDITIONS

The South Atlantic was thoroughly explored in 1925-27 by the German naval vessel "Meteor". The published temperature observations of this expedition have been examined, but they reveal little that is of special significance to this report. The winter stations (May 15-Sept.15) from the southern hemisphere show much the same conditions as prevail in the North Atlantic. In the central part of the ocean these observations give the following mean values for the thickness of the wind stirred surface layer

Latitude	Depth of homogeneous layer
40°S	50-75 fathoms
30°S	40-70 "
20°S	45-60 "
10°S	20-40 "

In the east along the African coast the offshore trade winds greatly reduce the depth of the mixed layer, while in the west along the South American coast relatively poor conditions are also indicated by a few scattered observations. But nowhere are the available data in disagreement with the principals arrived at in the analysis of North Atlantic observations. In short, in all probability if we had a thorough knowledge of the sound transmitting conditions in one ocean, we could apply these results to the rest of the world. The current systems of all the oceans are closely comparable, just as their general wind systems are very similar.

4. PACIFIC WINTER CONDITIONS

Until actual observations are available it is believed that during the four winter months the pilot charts can to some extent be interpreted in terms of sound transmission, provided one has in mind the various oceanographic factors discussed in this report. Let us take for example, the North Pacific ocean in winter and consider how the depth of the wind stirred layer should vary from one region to another.

Starting in the east we find prevailing northwest winds blowing parallel with the coast from off San Francisco to the southern tip of the peninsula of Lower California. Since this is the northern hemisphere such winds will produce an offshore component to the surface layer with upwelling along the coast. Relatively poor sound conditions can be expected inshore, but the wind stirred layer will increase in thickness gradually in an offshore direction so that at a distance of perhaps 200 miles from the coast 25 fathoms of homogeneous water might be found.

There is a further factor, unfavorable to sound ranging, in such a situation as the southward flowing current off California. Since the water is moving from the north it is approaching warmer latitudes, and even in winter this may produce some warming near the surface. However, such an effect has not yet been investigated.

Continuing southward, offshore winds along the coast of Mexico and Central America should make this a poor region for sound ranging, comparable to the west coast of Africa in similar latitudes.

Further westward in the northeast trade wind belt the conditions should likewise be similar to the corresponding region of the North Atlantic. The doldrum belt is at about Lat. 5°N in midwinter and here the surface layer should have considerable stability due to high rainfall and minimum evaporation. Northward across the trades the homogeneous layer will increase in thickness from only a few fathoms in the doldrum belt to perhaps 50 fathoms or more at the southern edge of the horse latitude belt. This explains the relatively good reputation for sound transmission which the region of the Hawaiian Islands enjoys. Conditions should be even better further to the westward.

The area just east of the Philippines is entirely comparable with the waters off the Bahamas, while from there northward towards Japan conditions must be very similar to the western North Atlantic in corresponding latitudes. Unfortunately it will be noted that the oceanographic data on the North Atlantic chart (Fig. 40) are very scanty in this area.

The continental shelf off Japan is much narrower than along our eastern coast (except near Cape Hatteras), consequently the warm Japan Current flows northeastward, close to the shore. However, north of Lat. 35° it leaves the coast just as the Gulf Stream does, and poor sound conditions must prevail all along the northern boundary of the warm current. Further offshore in the latitudes of westerly winds a very deep mixed layer can be expected in winter, just as in the northwestern Sargasso Sea.

South of the Aleutian Islands, as is the case south of Greenland, the wind stirring in winter must go very deep indeed, perhaps all the way to the bottom. Sound conditions should also be excellent at this season in the Alaskan Gulf. Further southward off the Canadian coast the winter gales also should maintain a relatively deep mixed layer as is believed present off the coast of northern Europe.

Such speculation concerning the thickness of the homogeneous layer during the winter months in the North Pacific cannot be backed up by any trustworthy oceanographic data, except along the southern California coast. However, it is to some extent justified on the basis that from the oceanographic standpoint, provided we compare geographically analogous parts, the oceans are very similar one with the other. Thus, as previously emphasized, a thorough knowledge of the structure of the surface layer in one ocean should give a good idea of the sound transmitting qualities in all other parts of the sea.

5. SUMMER CONDITIONS

When we turn to the four summer months (May 15-Sept. 15) we can get almost no help from the oceanographic data because of the wide depth intervals at which temperature has generally been observed. It is probable that in the latitudes of westerly winds only 5-10 fathoms of mixed water can be expected. Moreover, the diurnal thermocline during periods of moderate winds probably often produces thermally stable water at depths of between 2 and 3 fathoms.

In the latitudes of easterly winds the summer sound conditions should be somewhat more favorable. In the heart of the North Atlantic trades, for example, a number of summer oceanographic stations show 25 to 40 fathoms of mixed water at the surface. Nearer the equator the effect of the secondary thermocline vanishes, so that it is probably safe to say that between Lat. 10°N and 10°S there is no difference between winter and summer, except as the wind system moves north and south with the sun. This will shift the latitude of the doldrum belt and its stable surface layer. But where the trades blow steadily in low latitudes the season probably has little influence on the success of echo ranging.

CHAPTER VII

RECENT DEVELOPMENTS

This preliminary analysis of problems involved in the horizontal transmission of sound in the surface layer, no doubt leaves much to be desired from the naval point of view. Not only are the pertinent oceanographic observations scattered and inadequate, but there exist very few cases where sound transmission has been measured in areas free from the disturbing effects of coastal water. However, this report is designed merely to serve as a stop-gap and as a guide for investigations now in progress.

The National Defense Research Committee having provided the necessary funds, beginning in October 1940 the Woods Hole Oceanographic Institution undertook a two year program of cooperative work on the oceanographic and meteorological aspects of the sound transmission problem. As rapidly as results are obtained further reports will be issued. The oceanographic program can be summarized as follows:

1) Construction of a considerable number of cheap, rugged bathythermographs for use from commercial and naval vessels.

2) Training of observers to use this equipment from as many cooperating vessels as can be arranged for.

3) Study of the daily and seasonal heat exchange at the surface and its effects on sound transmission.

4) Investigation of wind transport and wind turbulence in the surface layer.

5) Preparation of seasonal charts showing the average conditions in the surface layer.

Up to February 1st, 1941 (the time of writing) four cruises in the western North Atlantic have been carried out by the research vessel "Atlantis". Using the bathythermograph in conjunction with the Nansen bottle observations in the surface layer, it has been found that at least during the autumn and early winter months conditions are essentially as predicted in this report. The work will be continued and the range of the surveys greatly enlarged as the necessary new instruments are constructed.

The practical value of this program will depend largely on securing controlled sound ranging data under a wide variety of accurately known water conditions. Carefully planned experiments must be made to find how far outside the limits of the direct beam useable echoes may be obtained. Similar experiments can determine the effectiveness of sound which undergoes one or more reflections at the surface (or bottom) between the projector and the target. This phase of the work should obviously be supervised by the sound experts at the Naval Research Laboratory, but all vessels having high frequency sound equipment can contribute important supplementary data. The following summary outlines the sort of observations which are needed:

1) Date, time and position.

2) Wind direction and strength, and if possible the wind of the preceding 24 hours.

3) Air temperature (preferably both wet and dry bulb) and sea surface temperature.

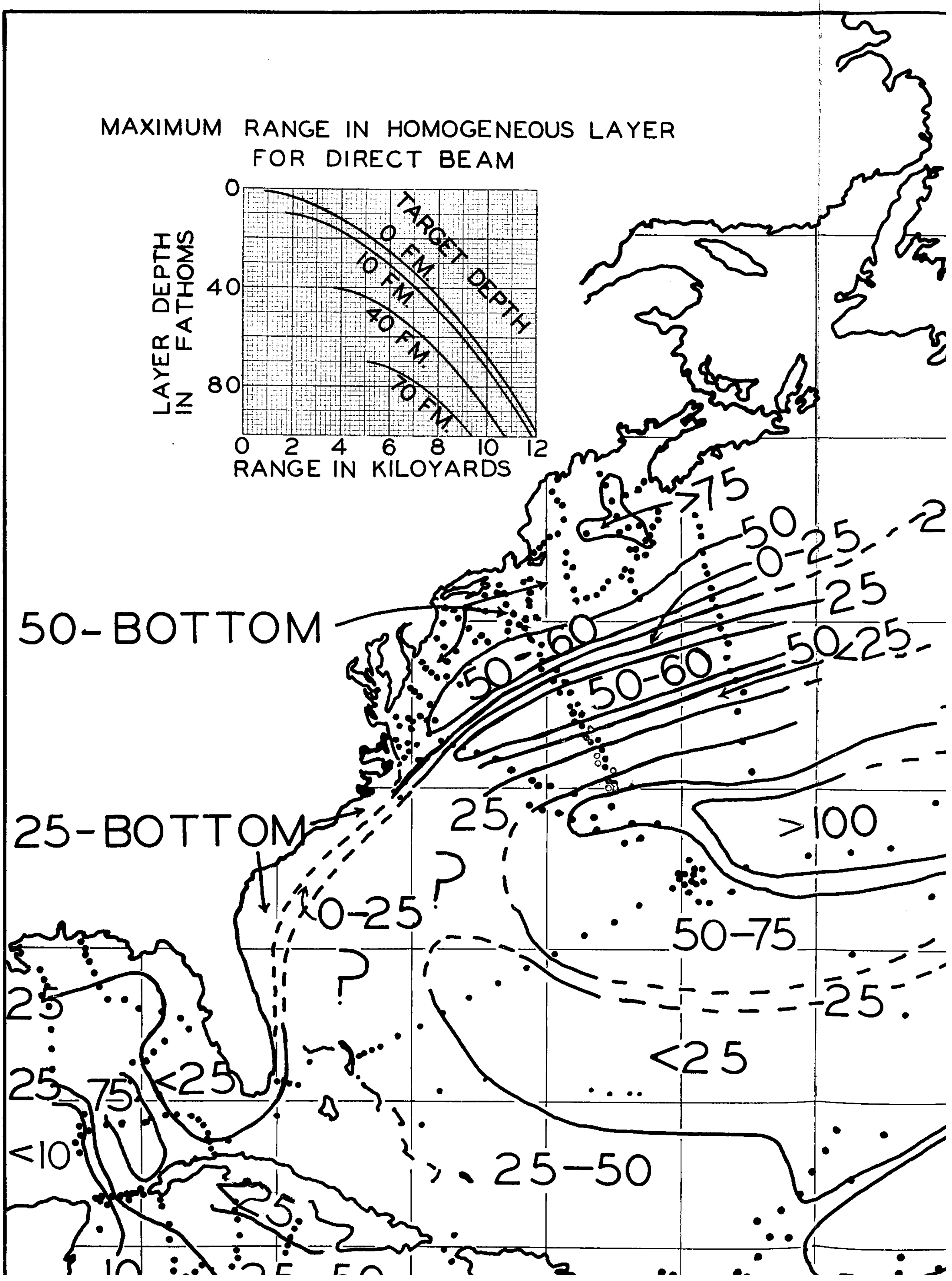
4) Maximum range and direction of echo, with a description of the target and, in the case of a submarine, its depth.

5) Remarks concerning the clearness of the echo and the presence or absence of other disturbing sound.

6) If in addition a bathythermograph record can be obtained at the same time, the value of such data will be greatly increased. As instruments are completed for use by destroyers, full instructions will be provided.

If our present conclusions are essentially correct and future investigation does not uncover additional complications, it seems likely that the local range of horizontal signals will soon be predictable in a routine way from bathythermograph records. The most recent instruments are light enough to be handled at speed up to 15 knots from a standard wire sounding machine. Their design has been improved so that they are rugged and free from vibrational difficulties. Moreover, they are inexpensive so that several can be carried by each vessel. At the present time difficulty is being experienced in using the bathythermograph from a destroyer. When lowering from a short boom amidships, the convergence of surface water near the stern causes the wire to foul the screws. To overcome this difficulty the instrument is being fitted with a small rudder so that while being hauled in it will tow clear of the ship's side.

The second phase of the problem, that of predicting the sound transmitting qualities of the waters towards which a convoy or fleet is moving, will probably involve close cooperation with meteorologists, at least during the summer season. Such charts of the wind-mixed layer as are now being prepared will only be able to show the expected range with a steady wind and under average weather conditions. But it is by no means improbable that the investigations in progress will enable the diurnal effect to be forecast from daily weather maps. In the same way, the various temporary effects of the wind should also soon be predictable.


MAXIMUM RANGE IN HOMOGENEOUS LAYER
FOR DIRECT BEAM
LAYER DEPTH IN FATHOMS
0
40
80
TARGET DEPTH
0 FM.
10 FM.
40 FM.
70 FM.
0 2 4 6 8 10 12
RANGE IN KILOYARDS
>75
50
0-25
25
50-BOTTOM
50-60
50-60
50
<25
25
>100
25-BOTTOM
0-25
50-75
25
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<25
25
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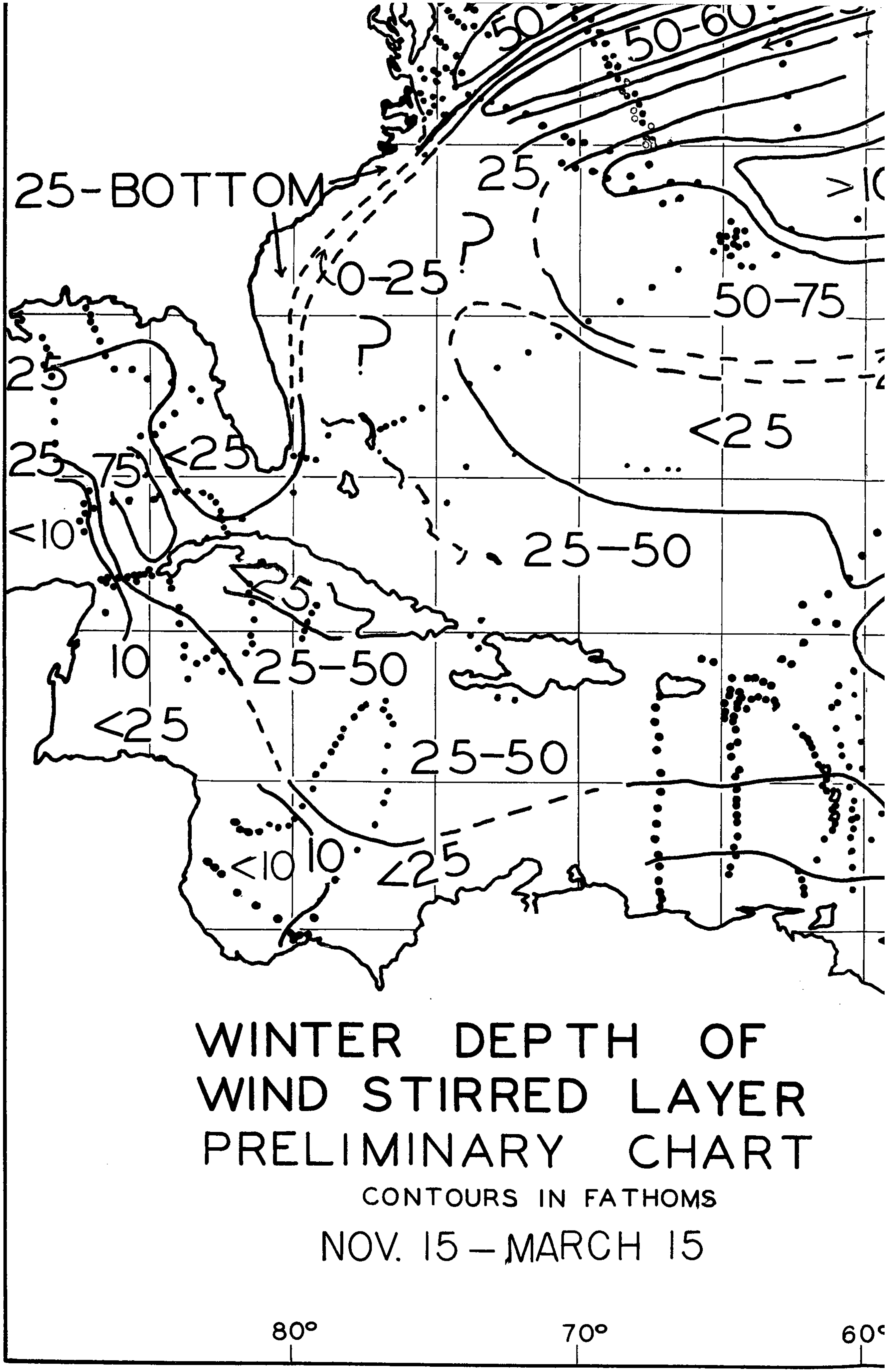
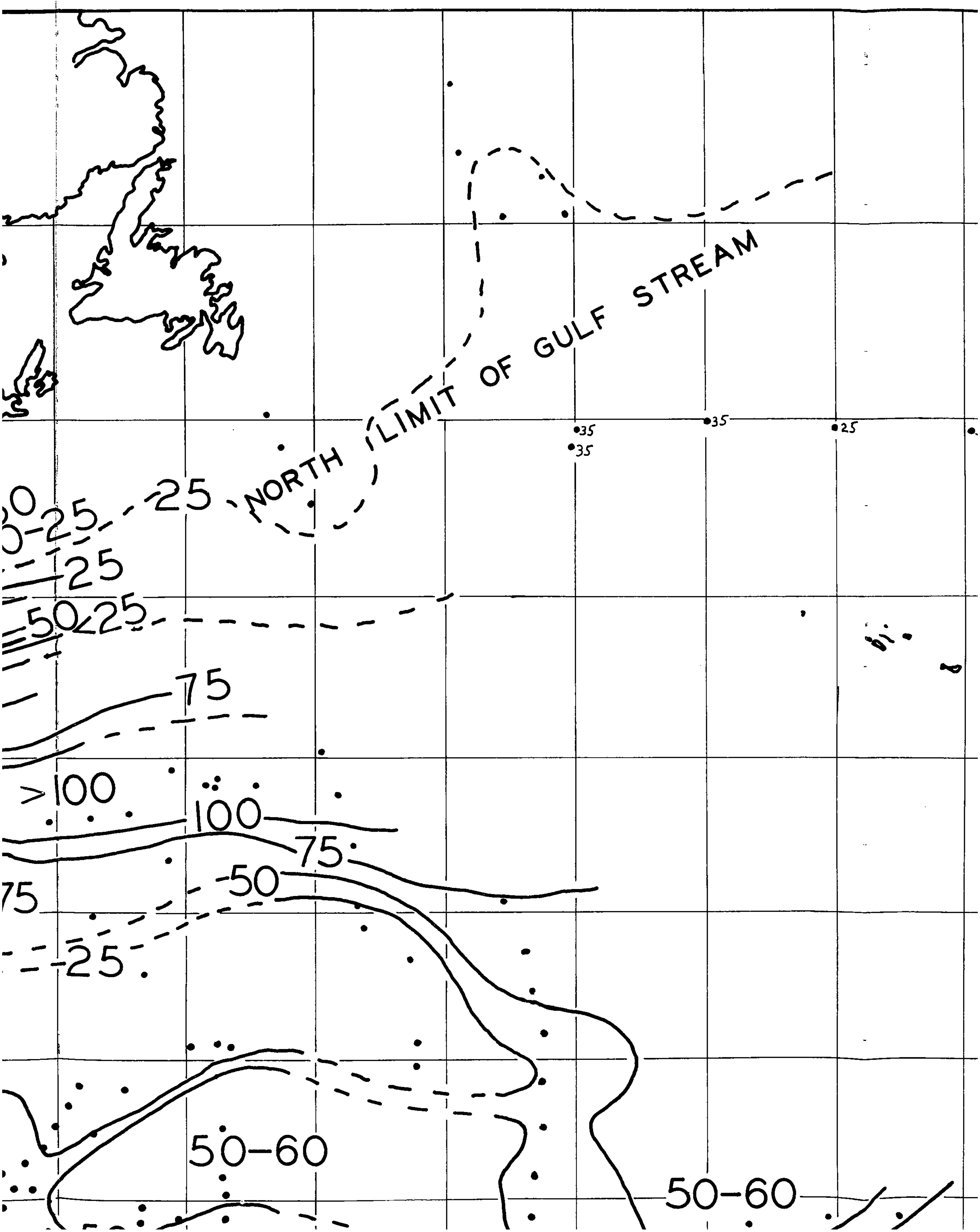
75
<25
<10
25-50
25



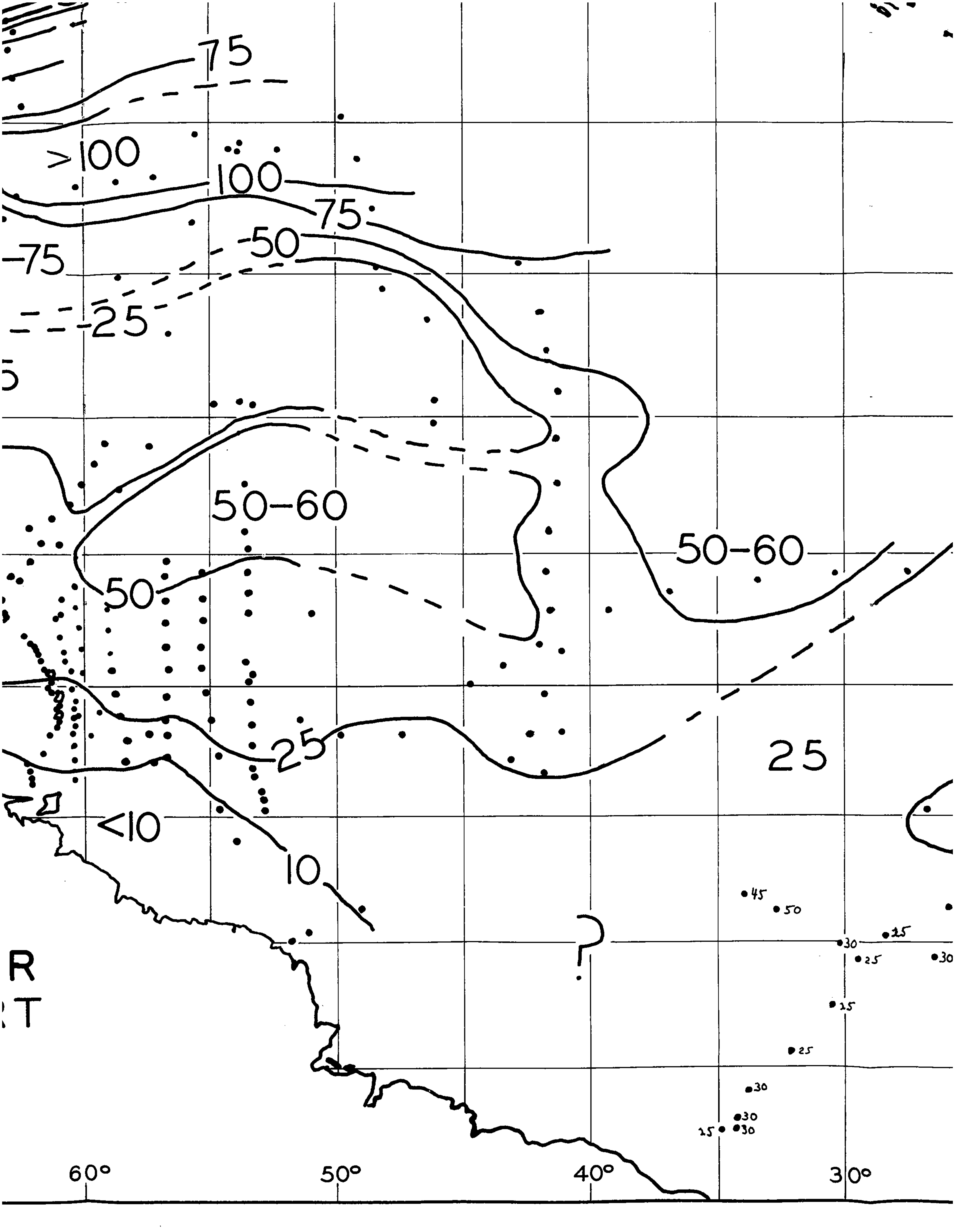
WINTER DEPTH OF
WIND STIRRED LAYER
PRELIMINARY CHART

CONTOURS IN FATHOMS

NOV. 15 – MARCH 15



NORTH LIMIT OF GULF STREAM
35
35
35
25
25
20-25
25
50
<25
75
>100
100
75
50
75
25
50-60
50-60



75
>100
100
75
-75
50
25
50-60
50
50-60
25
25
<10
10
45
50
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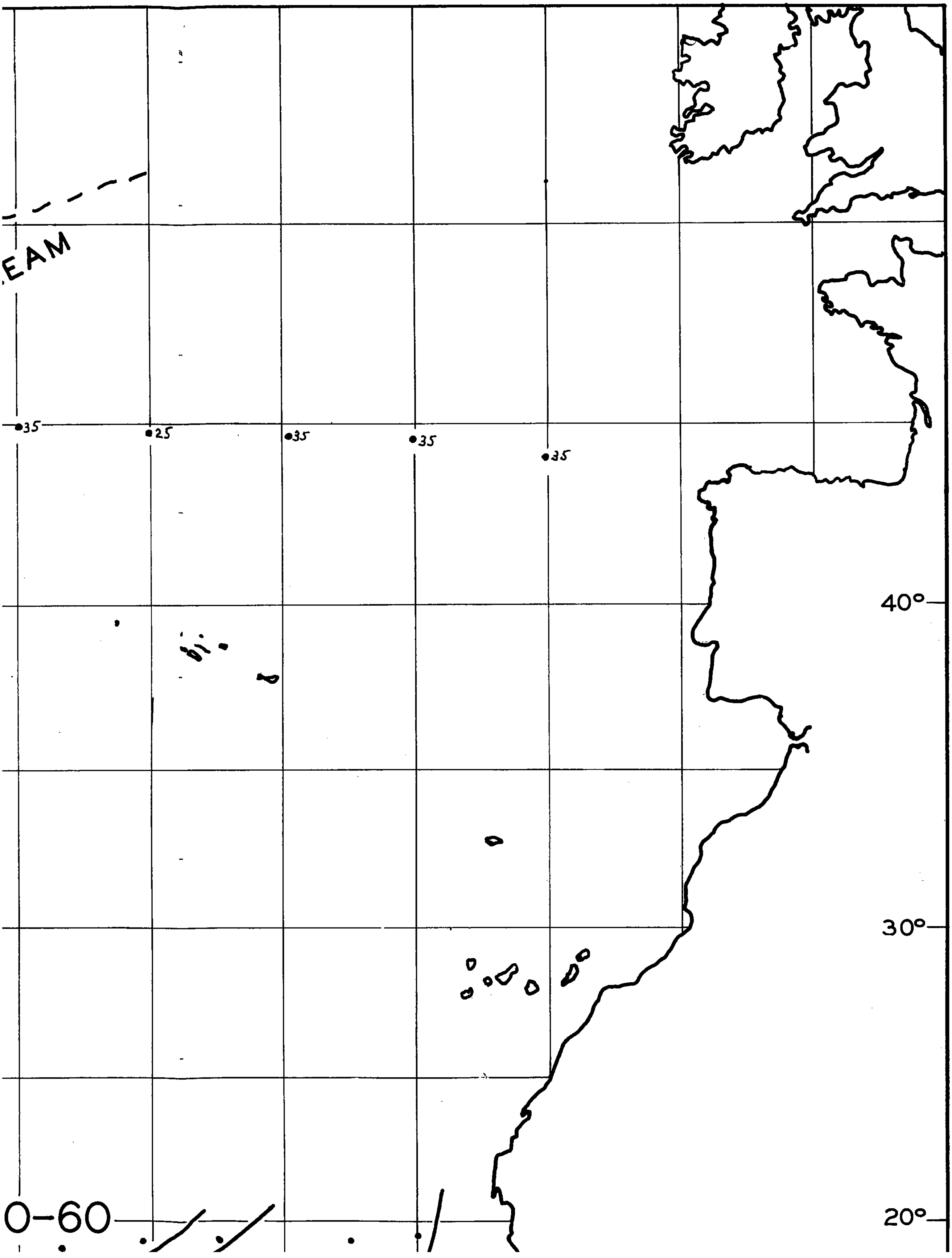
60°
50°
40°
30°



EAM
35
25
35
35
35
40°
30°
20°
0-60

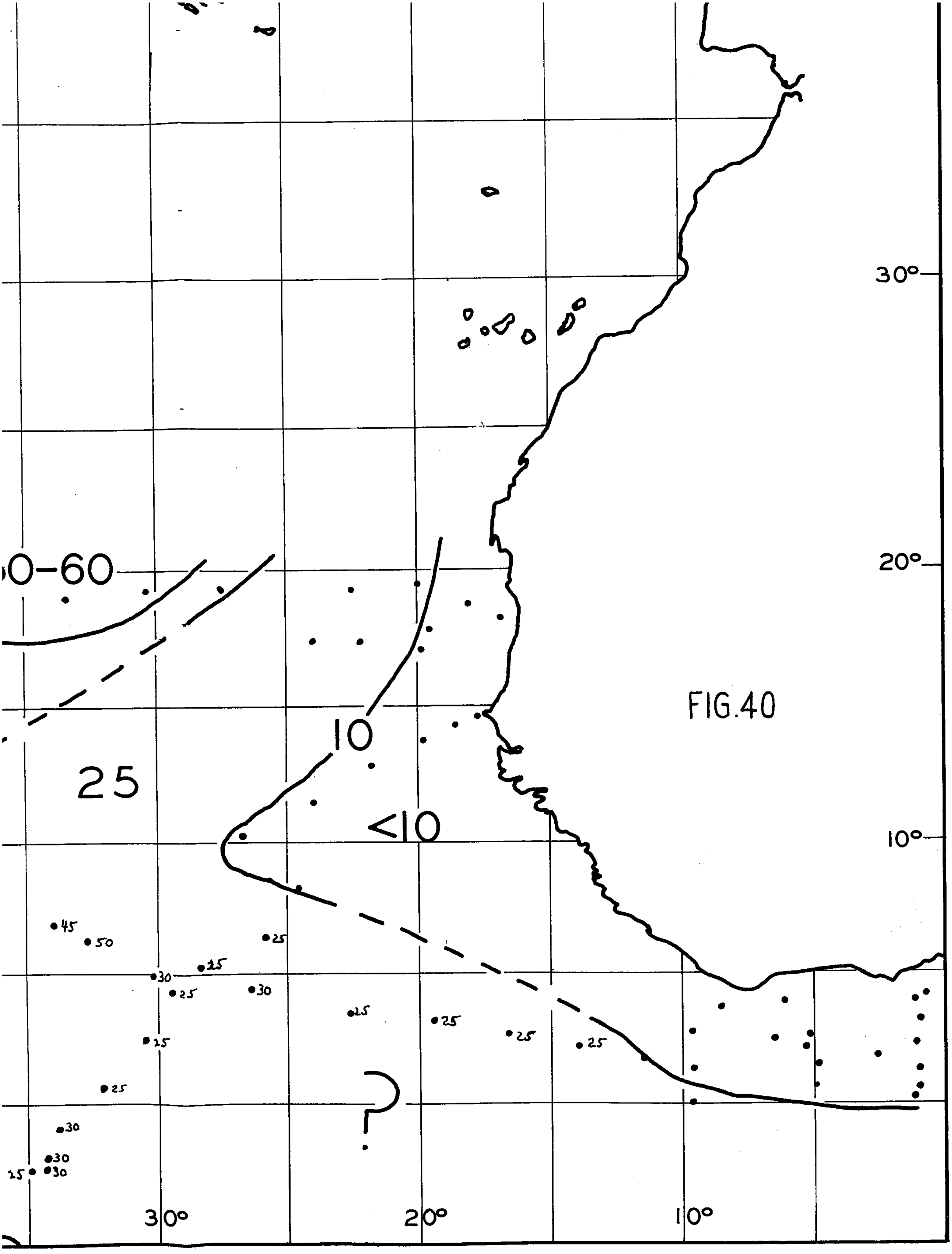

30°
20°
10°
50-60
25
10
<10
45
50
25
30
25
25
30
25
25
25
25
25
25
25
30
30
30
25
?
30°
20°
10°

FIG.40