Underwater Sound: Deep-Ocean Propagation

Variations of temperature and pressure have great influence on the propagation of sound in the ocean.

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The subject of underwater sound lies at the intersection of acoustics and oceanography, and the particular flavor of this branch of science arises from the necessity for continually considering the properties of the environment which provide the conditions for acoustics. Every underwater acoustician must try to be something of an oceanographer, and many oceanographers are finding a knowledge of acoustics advantageous.

An additional major influence on the subject has been its growth out of sonar, the science and engineering of the use of underwater sound in searching for submarines. This urgent practical application has generated and motivated the study of underwater sound, although recent applications to the study of the oceans have begun to broaden its uses.

Here I treat underwater sound as a body of knowledge and problems without particular reference to either its military applications or the technology used to achieve the knowledge. The emphasis is on the interaction of the physics and the environment. One major area of interest has been very long range propagation of sound in the deep ocean, and that is the subject on which I focus my attention. The major outlines of the subject are established by the sound-absorbing properties of sea water and the manner in which the sound-velocity structure of the ocean affects propagation.

Absorption

The term *absorption* refers to decrease in the energy in a sound field due to conversion into heat. In the classical mechanism, viscosity and heat conduction are involved in the transfer (*1*). Even in distilled water the classical coefficients are insufficient to account for the measured absorption, which appears to arise from a molecular relaxation phenomenon (*2*). The change in pressure as the sound wave passes produces changes in the percentages of molecules of two structural forms, a "liquid" form and an ice or "solid" form, the molecules having first one structure and then the other as the pressure oscillates. There is a characteristic time for this shift in structure, hence it will lead or lag behind the acoustic pressure change, depending on the frequency of the latter. This structural shift leads to dispersion (variation of wave velocity with frequency) and absorption of energy from the sound waves by these internal states of the liquid. The frequency (f) corresponding to this relaxation phenomenon in distilled water is much higher than any frequencies of practical concern to us (> 10 Mcy/sec), and the absorption in the frequency region of interest is proportional to the square of the frequency.

The dissociation of $MgSO_4$ present in sea water provides an additional relaxation phenomenon (*3*).

$$MgSO_4 \cdot H_2O \rightarrow MgOH^+ + HSO_4^-$$

and further dissociation

$$MgOH^+ + HSO_4^- \rightarrow Mg^{++} + OH^- + H^+ + SO_4^{--}$$

The first of these reactions adds a small additional absorption proportional to f^2 throughout the range of interest. The second dissociation has a characteristic frequency of the order of several hundred kilocycles per second. At sound frequencies above 1.5 Mcy/sec this relaxation adds nothing to the other absorptions, but below 100 kcy/sec it is responsible for greatly increased absorption of sound in sea water.

The presence of NaCl in sea water acts to depress the absorptive effect of the $MgSO_4$ dissociation slightly, because the sodium and chloride ions associate with the decomposition products and interfere slightly with the reversibility of the reactions. From the point of view of the relaxation phenomenon, there are fewer $MgSO_4$ product ions participating, and there is an effectively lower concentration of $MgSO_4$.

At frequencies below 100 kcy/sec, the absorption coefficient is well approximated by $5.29 \times 10^{-3} f^2$ down to about 1 to 2 kcy/sec (*4*), where other difficulties, discussed later, intervene. The change in pressure due to absorption alone for a plane progressive sound wave in the sea is thus proportional to

$$\exp(-5.29 \times 10^{-3} f^2 r)$$

(f is in kilocycles per second; r, or distance from the source, is in kilometers) at the surface for sea water with a salt concentration of about 3.5 percent at 20°C. The actual coefficient of f^2 varies with the pressure (and hence with depth) and temperature.

The sea thus acts as a Gaussian-shaped filter for sound. Table 1 shows the effect on propagation.

It is clear that absorption alone prevents propagation of high-frequency sound in the sea over long distances. Thus there is emphasis on frequencies below 10 kcy/sec in studies and uses of long-range propagation of sound.

The absorption values given above are based in part on laboratory measurements at high frequency and in part on *in situ* measurements using frequencies down to about 2 kcy/sec (*4*). It is difficult to make measurements at low frequencies largely because of the small value of the absorption. In spite of the difficulties, attempts to measure low-frequency absorption are being continued because of the importance of more accurate values to predictions of long-range propagation.

Several recent experiments indicate marked departures from the f^2 law below a frequency of about 1 kcy/sec. These experiments include observations

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Table 1. Effect of absorption of sound in sea water on propagation.

f(kcy/sec)	Approximate $1/e$ range(km)
1	200
5	75
10	2
50	0.75

of reception in California and Hawaii of sound generated by an underwater nuclear blast several hundred miles southwest of California (via deep-sound-channel paths reflecting from Okinawa) (*5*) (a round trip distance of approximately 20,000 km, and a time of 3 hours and 38 minutes), as well as studies of propagation of sound from detonation of 1.8-kilogram TNT bombs along paths as long as 3800 kilometers in the deep Atlantic (*6, 7*). Instead of continuing to decrease as f^2 below a frequency of 1 kcy/sec, the absorption seems to level off to an f dependence, and it may even begin to rise with decreasing frequency below 20 cy/sec. The value is still low; the $1/e$ distance at 100 cy/sec is 2600 kilometers. It has been suggested that the apparently increased absorption results from scattering of sound out of the paths connecting source and receiver by inhomogeneous patches of water of differing sound velocity, resulting from mixing processes in the deep ocean. This hypothesis appears to be consistent with measured patch characteristics (*8*). The possible rise in attenuation below 20 cy/sec can be attributed to leakage out of the deep sound channel due to diffraction effects arising from the long wavelength. (*6*).

Speed of Sound in Sea Water

The relaxation absorption at sound frequencies of several hundred kilocycles per second implies dispersion (variation of sound velocity with frequency), but the actual effect is a small one and may be entirely ignored below 50 kcy/sec (*9*).

Sound travels in sea water (of salinity of 35 parts per thousand at normal temperature and pressure) at about 1528 m/sec, a rate approximately 0.3 percent faster than its rate in distilled water (*10*). The exact speed varies with temperature, salinity (parts per thousand of dissolved salts), and pressure. The speed increases by 4.6 m/sec with each 1°C increase in temperature (T); by about 0.16 m/sec with each increase in pressure (P) of 1 kg/cm^2 ($\cong$ 1 atm.); and by 1.4 m/sec for each part per thousand increase in salinity (S) above 35 parts per thousand. There are also nonlinear terms in the empirical formula, which fits the laboratory measurements to within about $\pm$ 0.22 m/sec. Values for 95 percent of all sea water (*10*) are found in the range

$$-3°\text{C} < T < 30°\text{C};\ 1\ \text{kg/cm}^2 < P < 1000\ \text{kg/cm}^2;$$

and

$$33\ \text{per mill} < S < 37\ \text{per mill.}$$

Sound Velocity Structure of the Ocean

The major oceanographic effects on sound velocity in deep water are provided by temperature and pressure. In most circumstances the effect of variation of salinity is sufficiently small to be ignored. Here I assume that the static pressure increases linearly with depth, ignoring the small deviations arising from density variations due to variations in temperature and salinity, and from the compressibility of the water. While this model is adequate for a description of the way in which the structure of the ocean affects the sound velocity structure, the variations which I have eliminated are among those that make it possible to explain the basic structure itself (*11*).

In an isothermal, isohaline (that is, of constant salinity) ocean, the pressure, and hence the sound velocity, increase approximately linearly with depth. In sunlit areas the heat of the sun warms the surface water. Since this heating expands the water and reduces its density, the heated water remains at the surface. Consequently, there results a gradual decrease in temperature with depth. This negative temperature gradient may have a larger effect on the sound velocity than increasing pressure has, so that near the surface the sound velocity decreases with depth.

At night or in cloudy weather, when the surface water may be warmer than the air, it loses heat, cools, becomes denser than the water immediately beneath it, and sinks, mixing with the water around it until equilibrium is reached. This process is aided by wind forces tending to mix the near-surface water. These forces are sometimes strong enough to produce mixing even on sunny days, in spite of the stabilizing effect of the heat of the sun.

In the temperate zones the seasons, with successions of long sunlit days with light winds in the summer and sequences of cold, cloudy, windy days in the winter, have the effect of making one portion of the daily cycle predominant, so that the yearly cycle produces results like those described for the daily cycle, but stretched out in time to a full year.

As the winter winds and surface cooling mix the water to progressively greater depths, the contrast between the density of the newly mixed water and the density of the underlying water becomes greater and greater because of the increasing density with depth arising from decreases in temperature with depth (and also from the effect of compressibility). Thus the density gradient at the mixing boundary becomes greater and greater, until the mixing forces are insufficient to overcome it and to mix the waters to greater depths.

Only in the Arctic are the mixing forces strong enough to reach to the bottom in winter. In temperate and tropical regions a permanent main thermocline (region of decreasing temperature) occurs at a depth which depends on geography and latitude but, at

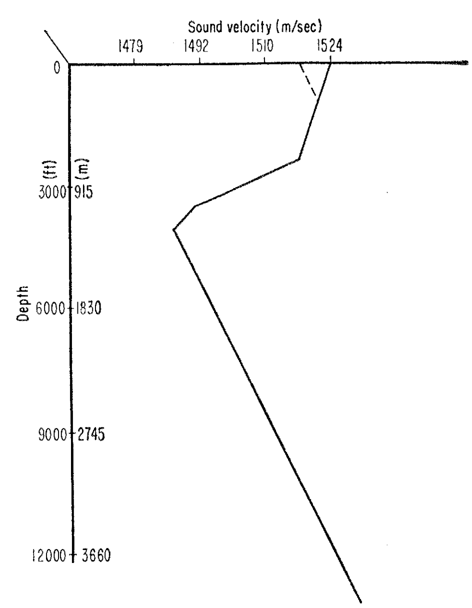

Fig. 1. Typical sound-velocity versus depth profile for the mid-latitude North Atlantic approximated by linear gradients. (Dashed line) Effect of a near-surface isothermal layer. [After M. Ewing and J. L. Worzel (*13*)]

middle latitudes in the Atlantic, is generally between 1000 and 1200 meters. Above this thermocline the temperature may increase monotonically toward the surface, or the water may be mixed, or a fairly complicated temperature-depth structure may occur (including subsidiary thermoclines), depending on the precise sequence of heating, cooling, and mixing that occurred during previous days, weeks, and months. Below the main thermocline the water is generally nearly isothermal, the temperature being determined by stability criteria for the water column.

In the heuristic sketch of the temperature structure of the deep ocean just given, a number of factors were ignored. These include the effects of evaporation and of salinity variations due to precipitation and the melting and freezing of ice. In addition, I have said nothing of the effects of the major water circulation patterns of the ocean. While these effects influence the values of depth and temperature at which the actual ocean structure stabilizes, they do not change in any major way the dynamics and structure outlined above. There are numerous exceptions to the simple model, however, including the regions of major currents (the Gulf Stream), regions of upwelling (west of Peru), and the Antarctic confluence.

I shall now describe a sound-velocity versus depth function typical for the temperate zone. Two interesting conditions of the near-surface water are the surface thermocline case and the isothermal layer case.

In the surface thermocline case the near-surface decrease of temperature with depth has a greater effect on the sound velocity than the increase of pressure with depth, and the sound velocity decreases monotonically from the surface down through the main thermocline. Below the bottom of the main thermocline the water is isothermal, or nearly so, and the sound velocity increases because of increasing pressure. Thus there is a minimum in the sound velocity at the depth of the bottom of the thermocline (Fig. 1).

In the isothermal case the water is mixed and has a constant temperature down to the depth of the main thermocline. The effect of pressure leads to an initial increase in sound velocity with depth. At the main thermocline the rapid decrease of temperature with depth leads to a decreasing sound velocity until the bottom of the thermocline is reached. Below this point the increase in sound velocity with pressure again becomes dominant.

Effects of Sound Velocity Variations

Having laid out the main oceanographic factors controlling the propagation of sound in the deep ocean, I now examine some of the interesting features of that propagation, assuming a bottomless ocean with a flat surface.

In a homogeneous ocean, sound from a point source would spread throughout the medium and the sound intensity would decrease with distance from the source (r) in accordance with the inverse square law or as $1/r^2$, absorption being neglected. There would be some effects due to the interference of surface-reflected and directly propagating waves (*12*), but, generally speaking, at large distances from the source all parts of the medium would be filled with sound. The sound-velocity variations described above, however, introduce radical variations from this situation.

The deep minimum in sound velocity deep in the ocean has a channeling effect on the sound, as does the minimum at the surface when such a minimum exists. Because sound originating on the axis of the deep sound channel (the depth of the minimum in the sound velocity) is confined to the channel, the sound intensity in the channel tends to fall off only inversely as the distance ($1/r$). Sound originating near the surface does not reach all ranges and depths, and there are pronounced "shadow zones" and peaks of intensity. In all cases the sound goes from source to receiver along many paths.

In many studies of deep-ocean propagation geometrical acoustics (ray theories) is used as an approximation to the proper solution of the wave equation (*13*), as it is in studies of propagation in other inhomogeneous media, including studies of propagation of sound in the atmosphere, electromagnetic propagation above the earth, and propagation of seismic waves in the earth.

This form of solution is satisfactory when the sound velocity does not change much in a distance equal to a wavelength of sound at the frequency in question. Thus, it is generally a good approximation at high frequencies, but difficulties may be expected at low frequencies or in regions of rapid change in velocity. Numerical methods and digital computers have been used to solve the wave equation in more correct ways, and for several special cases it has been solved exactly, but illustration of major effects is most easily made with the ray theory (*14, 15*).

In the simplest form of ray tracing we assume that variation of sound velocity occurs only vertically, and that the sound ray is governed by Snell's law:

$$\frac{\sin\theta}{v} = \text{constant } (k)$$

where θ is the local angle of the ray with the vertical and v is the local sound velocity. One important consequence of this law is that a ray has the same angle with the vertical at all depths where the sound velocity is the same. The constant k is set by the sound velocity at the source and the initial angle of the ray with the vertical. As a consequence, we may write:

$$\tan\theta = \frac{d\rho}{dz} = \frac{kv}{(1-k^2v^2)^{\frac{1}{2}}}$$

where z is the depth measured from the surface and ρ is the horizontal range from the source.

If v is a function of z, this differential equation may be integrated to give ρ as a function of z. In the case where the dependence of velocity on depth may be taken to be linear, the result is particularly simple.

The rays are arcs of circles concave in the direction of lower velocity (that is, the rays are bent away from regions of higher velocity and toward regions of lower velocity, as is usual in geometric optics).

A ray starting downward at angle θ_0 with the vertical from a point with velocity v_0 into a region of monotonically increasing sound velocity will be horizontal at a depth, with velocity v_1 such that

$$\sin\theta_0/v_0 = 1/v_1$$

or

$$v_1 = v_0/\sin\theta_0.$$

The portion of the path at greater range than ρ_1 is the mirror image of that between the source and ρ_1. The steeper the initial angle (the smaller θ_0 and $\sin\theta_0$), the deeper the deepest point (the turn-around point) of the ray.

If a more complicated velocity versus

depth function (velocity profile) is approximated by linear sections, the ray-tracing equation may be solved for each such section, and appropriate circular ray segments may be joined at the boundaries of the section to piece out complete rays. This is the commonest form of ray tracing, although more complicated functions have been used for $v(z)$ (*15*). An interesting article by Cooper suggests that linear gradients may be a good approximation in many cases, since deep water may consist of a sequence of isothermal layers (*16*).

Propagation in the Deep Sound Channel

Propagation in the deep sound channel is referred to as SOFAR propagation. SOFAR is an acronym for Sound Fixing and Ranging (*18*). We may now trace a ray starting at some angle with the horizontal at the axis of the deep-ocean sound channel (minimum velocity point). Increasing the velocity below or above the axis turns the ray toward the axis until it becomes horizontal. It then starts back toward the axis, eventually crossing it at the same angle with the horizontal that it had initially. This process continues, with the ray repeatedly crossing the axis.

The ray with the largest initial angle with the horizontal has the largest excursion away from the axis and the longest range from axis crossing to axis crossing. The shallower the initial angle of the ray with the horizontal the greater the number of axis crossings it makes in a given horizontal distance (Fig. 2). In this ray formulation the number of axis crossings of the initially horizontal ray is infinite. For any finite frequency (and thus wavelength) the ray approximation within several wavelengths of the axis is no longer a reasonable approximation to a correct solution of the problem. It is, however, convenient to ignore this fact, since useful heuristic results can be obtained by doing so.

Only rays starting from the source at certain discrete angles will arrive at the axis precisely at the position of a given receiver. The simplest of these is the ray that leaves the source at just such an angle as to return to the axis at the receiver. The next discrete ray satsifying the requirements starts at a shallower angle with the horizontal, so that it has a loop on one side of the axis, then one loop on the other side which it completes just at the receiver. Other rays are found with 2, 3, 4, . . . axis crossings between source and receiver (Fig. 2). The number of rays per unit initial angle increases as the initial angle becomes shallower relative to the horizontal, finally becoming infinite as the initial angle approaches zero. For long ranges in an ocean of finite depth, solutions with only a few loops may be lacking due to the presence of bottom and surface.

With the assistance of Snell's law the travel time along each ray may be determined by integrating the quantity (increment of ray path length divided by the sound velocity) along the ray. The rays that make the largest excursions away from the axis travel in the regions of highest sound velocity and, in spite of their longer paths, have the shortest times for travel of sound from source to receiver. The sound traveling along the axis itself takes the longest time, as its entire path is in the region of lowest velocity. The many nearly horizontal paths have almost the same travel time.

Thus a short pulse of sound at the source will arrive at the receiver as a sequence of pulses separated by continually decreasing times. The final pulse arrives at the time appropriate for travel along the axis. Since the number of pulses per unit time (and unit initial angle) diverges near the final time, for any real pulse of finite duration we expect an increase in sound level until the final arrival, after which there will be silence. The final sound level is predicted to be infinitely high, since the number of pulses overlapping is infinite, but, as previously noted, we need better approximations to the exact solution if more correct results are to be obtained for this portion of the problem.

The multiple arrivals, finally merging into a crash of sound followed by silence, can be observed at distances of thousands of kilometers when small charges of explosive (1.8 kg of TNT) are detonated at the axis of the deep sound channel. The greatest distance at which they have been observed experimentally is 19,200 kilometers, from a point near Australia to Bermuda (*6, 13, 17, 18*).

Because of the difference in effective horizontal velocity of the initial and final arrivals, the composite pulse grows longer with range. The pulse lengthens about 9.4 seconds in the first 1600 kilometers (*6*).

Convergence Zone Propagation

Let us now examine what happens to sound from a source near the surface in the surface thermocline model. The ray that leaves the source horizontally will be refracted downward at steeper and steeper angles until it crosses the axis of the deep sound channel, after which it will be refracted upward until it is horizontal, finally turning upward and crossing the axis, arriving at the surface once again horizontally. Rays which start out more vertically go deeper before being refracted back toward the surface.

For many possible profiles, including those characteristic of the true oceans, the rays with initial angles slightly steeper than horizontal have shorter surface-to-surface ranges; those with deeper initial angles have longer surface-to-surface ranges. Thus there is an initial angle which gives minimum surface-to-surface ranges (Fig. 3).

A suitable measure of sound level in ray theory is the number of rays arriving at a given range and depth. This measure, proportional to $1/(d\rho/d\theta_0)$, describes the way in which source energy, originally radiated equally in all directions, is concentrated by the refraction at particular ranges and depths. Because of the minimum in range as a function of θ_0 there is a zero in the expression $d\rho/d\theta_0$, hence a region in which rays overlap, and the ray theory predicts an infinite density of rays and an infinite sound level. The theory requires correction for diffraction effects in this region, but these corrections merely modify the prediction from infinity to a high sound level.

The sound level near the surface from a surface source decreases rapidly in a horizontal direction away from the source, since all the energy is refracted downward. There is a complete "shadow zone," then a caustic zone (a narrow region of very high sound level), then a somewhat wider region of lower sound level. This pattern is repeated over and over.

After the rays reach the surface in the caustic zone the horizontal ray is once again refracted downward. The steeper rays reflect downward from the surface of the ocean (when smooth, the ocean surface is an excellent reflector), and each ray proceeds through its downward and upward cycle again. The range from the source to the first caustic zone is 48 to 56 kilometers. The width of the first high-intensity zone, or "convergence zone" (which includes the caustic zone and the region of lower sound intensity, at the end of which the last ray is reflected downward again), is about 8 kilometers. The range from the first caustic zone to the second is, again, about 56 kilometers, but the width of the second high-intensity zone is approximately twice that of the first. Thus there are regions of high sound intensity at 56-kilometer intervals, the width of these regions doubling with each "skip"; low-intensity intervals separate the high intensity regions (*19*).

The sound intensity at the first convergence-zone peak (about 56 kilometers from the source) is about 10^{-7} that 1 meter from the source. Spreading at 56 kilometers, as calculated from the inverse square law, would give an intensity of $(5 \times 10^9)^{-1}$ that at 0.9 meter from the source. Thus, the actual peak level is about 500 times the level expected on the basis of inverse square spreading.

The pattern of convergence zones has been observed in both the Atlantic and the Pacific to distances of more than 650 kilometers. It is rather more pronounced in the Pacific than in the Atlantic, largely because the sound-velocity minimum is at a depth of about 600 meters in the Pacific, while it is at about 1250 meters in the Atlantic. For a given depth of ocean the bottom is more of an interference in the Atlantic than in the Pacific.

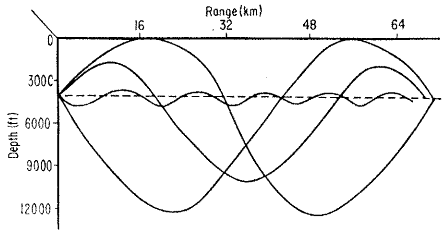

Fig. 2. Some rays originating at the axis of the deep sound channel. Metric equivalents for depth are given in Fig. 1.

Effect of the Bottom

Now let us consider the effect of the bottom of the ocean in the case where the sound velocity decreases from the surface to the axis of the sound channel. Although so far we have ignored it, the function of the bottom as a reflector (and scatterer) of sound is important. As we have seen, the convergence-zone peaking depends upon the properties of the refracted, surface-reflected rays (RSR rays), those that reflect at the surface and are turned upward at depth by the refractive effects of the pressure-induced increase in sound velocity below the main thermocline. If we insert a flat reflecting bottom at a depth shallower than the depth at which the horizontal ray at the surface again becomes horizontal (that is, the depth at which the sound velocity at depth equals the surface velocity), then all rays will reflect before they become horizontal. Thus they will return to the surface at ranges shorter than would be the case if they were refracted back, and the whole pattern of shadow zones and caustic peaks generated by the deep refraction will be destroyed. This is the fully "bottom-limited" case: no convergence zones and no marked shadow zones.

If we put our flat reflecting bottom at a depth below the horizontal point of some of the RSR rays we have a mixed case. There will be peaking due to the remaining RSR rays, perhaps a caustic zone if the bottom is deeper than the depth where $d\rho/d\theta_0$ becomes zero, and the "shadow zones" will be partially filled with bottom-reflected sound.

All of these situations can occur, depending on the precise details of velocity profiles and ocean depths. When seasonal increases in the near-surface velocity profile occur, the deep sound velocity which equals the surface sound velocity occurs at a greater depth, perhaps at the bottom, and the pattern may be changed from strong convergence-zone propagation to bottom-limited propagation.

The extent to which the "shadow zones" are filled with sound is dependent on the effectiveness of the bottom as a reflector. It is frequently a reasonable approximation to the facts to assume that sound incident on the bottom at angles steeper than some critical angle penetrates into the bottom (that is, less than 10 percent of the energy is reflected), while sound incident at more grazing angles is wholly reflected. Actually the bottom is frequently rough, and the scattering of sound other than specularly results in an apparent loss in bottom reflection, which increases with increasing frequencies. It is customary to use values deduced from measurements for losses in bottom reflection (*9*), and the geographical variability in values appears to be very high.

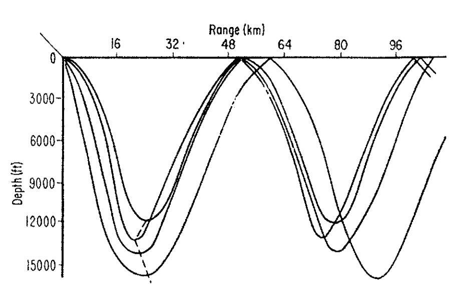

Fig. 3. Some rays originating at the surface, showing the origin of the 48- to 56-kilometer (30- to 35-mile) peaking phenomenon. (Dashed curve) Locus of ray turn-around points. Metric equivalents for depth are given in Fig. 1.

Flat areas of the bottom frequently appear to be covered by very soft sediments with sound velocities close to, or less than, the sound velocity of water; these areas can be very poor reflectors except at extremely low frequencies.

Roughness in the surface of the ocean can also lead to scattering of sound out of the paths connecting source and receiver, and thus to an apparent increase in absorption losses (*20*).

Effect of the Isothermal Layer

Now let us examine the case of a near-surface source where there is an isothermal layer (surface channel) above the thermocline. Starting at the surface, the sound velocity increases to the top of the thermocline, then decreases to the axis of the deep sound channel, and then increases.

We may trace a limiting ray (or split ray) which starts down from the surface at just such an angle that it becomes horizontal at the velocity maximum at the top of the thermocline. It may then be considered to be refracted either up or down. In the former case it goes up until it is reflected from the surface and repeats the previous cycle. In the latter case it is refracted down through the deep sound channel until the upward refraction below the axis bends it upward again and it returns to the bottom of the isothermal layer, again horizontal.

Rays starting from the source at shallower angles to the horizontal than the limiting ray will be refracted back to the surface before reaching the bottom of the isothermal layer and will act like RSR rays confined to the surface channel (Fig. 4). Rays starting out more steeply than the limiting ray will behave as deep RSR rays. Thus a portion of the energy will be confined to the surface sound channel and the remainder

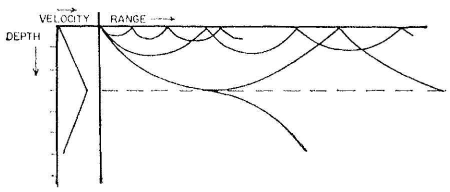

Fig. 4. The limiting ray and the trapped rays of the surface isothermal channel.

will have a typical convergence-zone character (*21*).

Sound of suitable frequency can propagate in the surface channel for many tens of kilometers. If surface heating warms the upper layer so that the sound velocity decreases from the surface to the top of the thermocline, instead of increasing, this channeling of the energy disappears, all rays become deep RSR rays, and no energy travels to the surface except by way of the convergence-zone mechanism.

It is not uncommon for near-surface channeling existing after several days of cloud cover and strong winds to disappear in the course of a single day of light winds and strong sun. In this situation excellent near-surface propagation to 15 or 30 kilometers will also disappear, to be replaced by the RSR near-source shadow zone. Since this is a refraction effect, increases in source strength do not overcome it. This effect became famous during World War II as the "afternoon effect." Excellent sound-ranging conditions in the morning after a cool night disappear by afternoon, to be replaced by unbelievably poor conditions, ranges of several kilometers being reduced effectively to zero.

It should be noted that the extreme shadow-zone, convergence-zone case occurs only for the surface source. For sources deep in the ocean the sound is much more evenly distributed, and for sources deeper than 3500 meters there is complete "insonification" (a jargon term, analogous to "illumination") of the near-surface region by upward-going rays out to about 32 kilometers in most cases. It is easy to see that this is so by considering the ray that starts horizontally from a source at a depth of 3500 meters (where the sound velocity equals that at the surface for the mid-latitude North Atlantic). It will arrive horizontally at the surface at a range of about 32 kilometers. Any ray that starts up from the source at a steeper angle will arrive at the surface at a shorter range; the steeper the angle, the shorter the range. Thus the whole region inside the "limiting ray" will be filled with sound. If the bottom is deeper than 3500 meters (say 5100 meters), the 32 kilometers may be extended to about 40 kilometers for rays that start down from the source but are refracted upward before hitting the bottom.

Summary

The absorption of sound in sea water varies markedly with frequency, being much greater at high than at low frequencies. It is sufficiently small at frequencies below several kilocycles per second, however, to permit propagation to thousands of miles.

Oceanographic factors produce variations in sound velocity with depth, and these variations have a strong influence on long-range propagation. The deep ocean is characterized by a strong channel, generally at a depth of 500 to 1500 meters. In addition to guided propagation in this channel, the velocity structure gives rise to strongly peaked propagation from surface sources to surface receivers 48 to 56 kilometers away, with strong shadow zones of weak intensity in between. The near-surface shadow zone, in the latter case, may be filled in by bottom reflections or near-surface guided propagation due to a surface isothermal layer. The near-surface shadow zones can be avoided with certainty only through locating sources and receivers deep in the ocean.

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