Arctic Hydroacoustics

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INTRODUCTION

The two features peculiar to the polar environment that most strongly influence underwater sound are the permanent ice cover and the velocity structure in the water. Ice movement generates background noise and the ice modifies propagation, particularly at high frequencies, by scattering waves from the rough ice boundaries. Sound velocity is a function of temperature, salinity, and pressure. The relationship between these variables in the central Arctic Ocean is such that sound velocity is generally an increasing function of depth from the surface to the bottom. Such a velocity profile is found only in polar waters. The sound velocity structure is remarkably uniform both as a function of location and time of year. Sounds are transmitted to great ranges in this natural arctic waveguide or sound channel by upward refraction in the water and repeated reflection from the ice canopy. A two-pound explosion of TNT has been heard at ranges exceeding 1,100 km. (700 miles). The surface sound channel of the Arctic is the polar extension of the deep sound channel or SOFAR channel of the nonpolar oceans (Ewing and Worzel 1948), but the arctic signals are often quite different from those observed in the deep channel, largely because of the predominance of low-frequency waves in the Arctic. The arctic sound channel is of considerable importance to the Navy because of the possibility of long-range detection and communication. That ocean also provides an ideal test area for new concepts of signal detection and processing because of the easy access to the sound channel and the permanence of installations located on ice islands.

The purpose of this paper is to review our present knowledge of underwater sound obtained from experiments made aboard drifting ice stations in the central Arctic Ocean and to recommend future research in this field. I shall present a summary of the results of experiments made by Lamont-Doherty Geological Observatory of Columbia University; these results have been published by Kutschale (1961), Hunkins and Kutschale (1963), Hunkins (1965), Hunkins (1966), and Kutschale (1968). Many of our experiments were conducted in cooperation with the U.S. Navy Underwater Sound Laboratory, the Pacific Naval Laboratory of Canada, and AC Electronics Defense Research Laboratories of General Motors Corporation. Results by workers from these laboratories have been published by Marsh and Mellen (1963), Mellen and Marsh (1963), Milne (1964), Buck and Green (1964), and Buck (1968).

Drifting ice stations provide an ideal platform for research on underwater sound. These stable platforms over deep or shallow water are far removed from ship traffic and they provide a large surface area for detector arrays. Detectors

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may be either seismometers mounted on the ice or hydrophones suspended at shallow depths in the water. Measurements of background noise and scattering layers may be made over periods of many years as the station moves slowly under the influence of winds and currents. Experiments on propagation are commonly made between two drifting stations, or between a station and an icebreaker or an aircraft. The latter type of experiment is particularly suited to measure the range dependence of the sound field and to determine the effects of bottom topography on the propagation.

High explosives have been the principal sound sources for transmission experiments. These sources radiate high sound intensity over a broad frequency range and they are easy to launch. Offsetting these advantages are the change of source spectrum with shot depth at constant charge size, occasional partial detonations, and some variation of firing depth for pressure-activated charges dropped from an aircraft. Also, detailed comparison of theoretical computations with measurements are often far more difficult than for constant frequency sound sources.

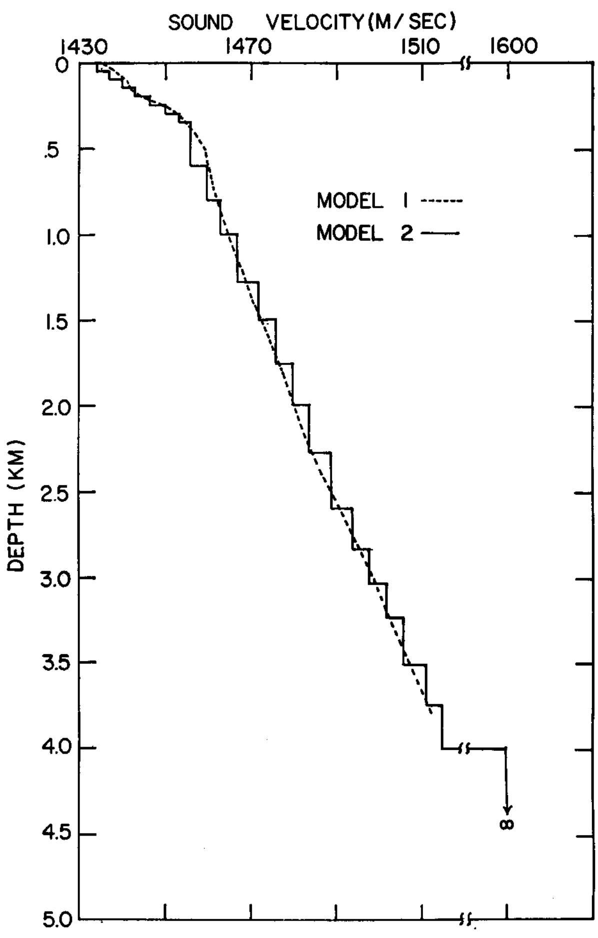

FIG. 1. Models representing the arctic sound-velocity profile.

Propagation

Many aspects of sound propagation in the Arctic Ocean may be understood in terms of ray theory, but at long ranges where low-frequency waves predominate, the solution of the wave equation in terms of normal modes is a powerful method for describing the propagation in detail. Fig. 1 shows sound-velocity profiles which closely follow those observed. Model 1 consists of a sequence of plane parallel layers, each layer having a constant velocity gradient. Such a model is convenient for numerical computations by ray theory. Model 2 represents the continuous variation of velocity with depth by a series of flat-lying layers of constant velocity and density. This representation is extremely useful for solving the wave equation when solid layers as well as liquid layers must be considered. The ice sheet is represented by a layer 3 m. thick, with the appropriate compressional velocity, shear velocity, and density. In the deep ocean the bottom sediments are represented by a liquid half-space, but in shallow water it may be necessary to represent the bottom by a layered solid.

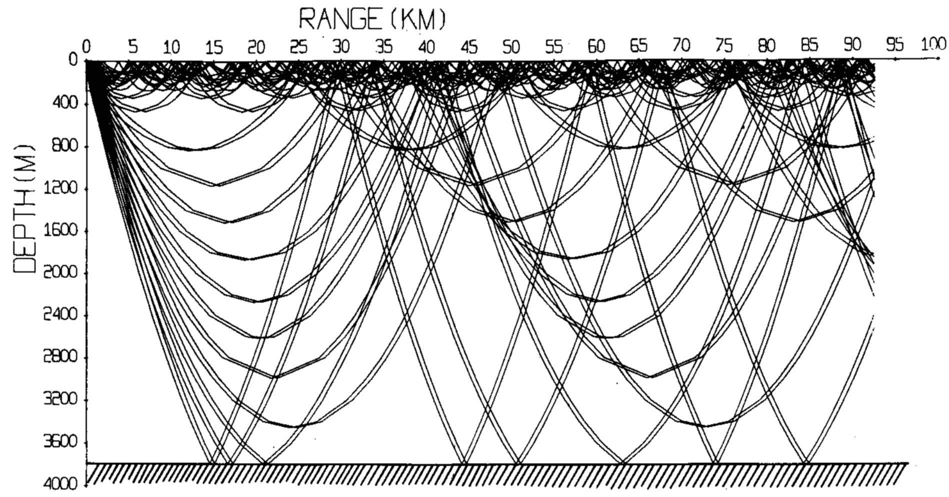

FIG. 2. Ray paths for model 1 of Fig. 1. Source depth 100 m.

Fig. 2 shows ray paths from a source 100 m. deep for Model 1 over the Canada Abyssal Plain in 3,800 m. of water. The concentration of rays near the axis of the channel is apparent. The paths were computed by high-speed digital computer employing a program supplied by the Naval Ordnance Laboratory (Urick 1965). The first refracted and surface-reflected (RSR) sound to arrive at a detector corresponds to the ray which has penetrated to the greatest depth into the channel. The RSR sounds arrive with increasing frequency until they terminate with the arrival of the sound which leaves the source in a horizontal direction. If the detector is deeper than the source, the last RSR sound is the one which arrives from a horizontal direction. The bottom-reflected sounds are generally interspersed between the RSR sounds and they may continue long after the last RSR sound has passed. Except for signals travelling over abyssal plains, the bottom-

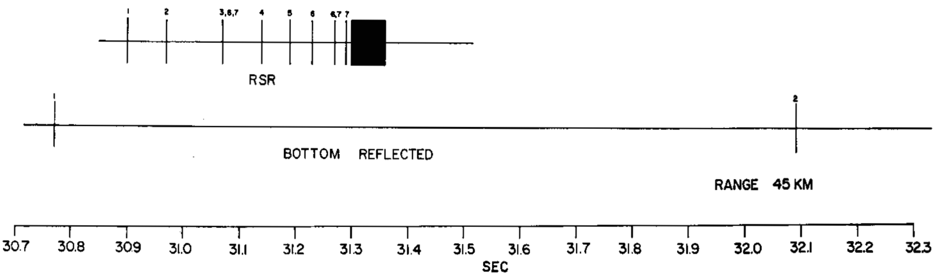

FIG. 3. Time sequence of arrivals for Model 1. Surface source. Surface detector.

reflected sounds have a noncoherent character and they are weak compared with sounds travelling by RSR paths. The first strong sound generally corresponds to the ray which has passed over all bottom topography without striking the bottom.

Any two rays that make the same angle with the axis of the sound channel differ only by a horizontal displacement. The time sequence of sounds travelling by different paths may be determined graphically. An example of the sequence of arrivals is given in Fig. 3 for a surface source and a surface detector separated by a range of 45 km. The number of cycles a ray has made is shown in the figure. There is a duplication or triplication of travel times for rays departing a surface source at angles of between 4 and 10 degrees. At these angles the signal strength is enhanced because of the focusing of sounds by the relatively strong changes in the velocity gradient in the upper 400 m. of water. At long ranges and low frequencies a regular oscillatory wave train is the result of interference of sounds

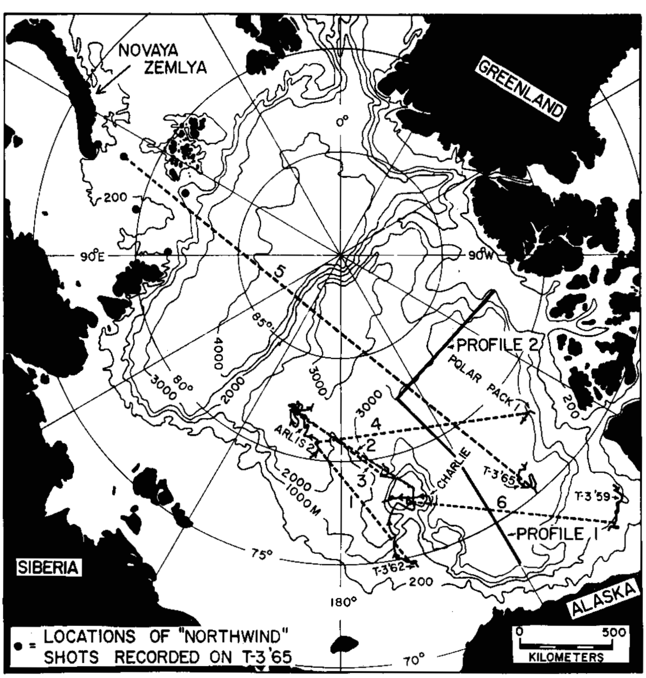

FIG. 4. Bathymetric map of the central Arctic Ocean. Propagation paths are numbered 1 to 6. Transmission profiles 1 and 2 recorded on Fletcher's Ice Island, T-3, during May 1968. Contours based on Geologic Map of the Arctic (1960).

travelling along the various paths in Fig. 2. These are the signals that are conveniently described in detail by normal-mode theory.

Fig. 4 shows the major bathymetric features of the central Arctic Ocean and the locations of drifting stations Fletcher's Ice Island: T-3, ARLIS II, Polar Pack I, and Charlie during experimental periods. Also shown in the figure are shot points occupied by the U.S. Coast Guard icebreaker *Northwind* and Profiles 1 and 2 recorded on T-3 from small TNT charges dropped by a Navy aircraft. Over four hundred shots were recorded and analysed. Ranges extend from 1 km. up to 2,860 km. The bottom topography is variable along paths 1 to 6, but along Profile 1 and part of Profile 2 in the Canada Abyssal Plain the bottom was flat. These transmission runs were made to measure the range dependence of the sound field without any of the effects caused by changes in bottom topography.

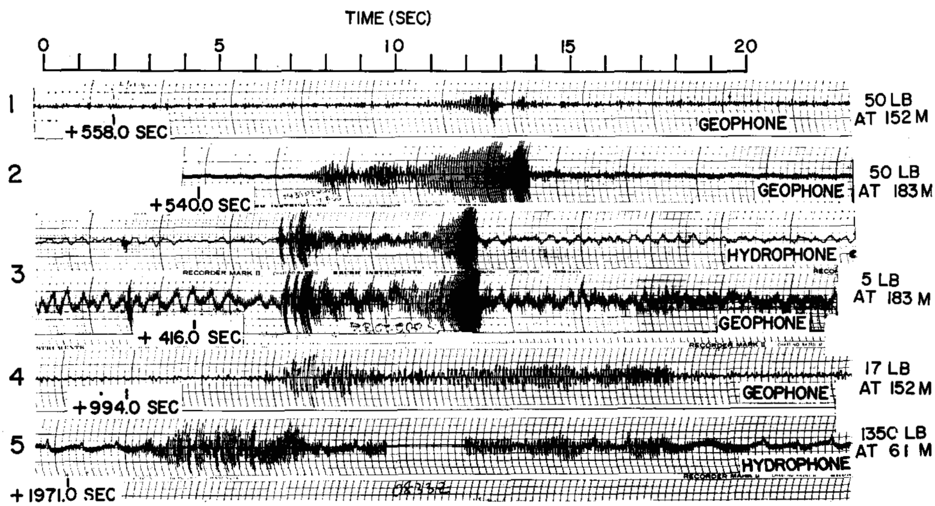

FIG. 5. Typical signals transmitted along paths 1 to 5 of Fig. 4.

Fig. 5 shows typical signals transmitted along paths 1 to 5 of Fig. 4. The variation in amplitudes between the signals is principally caused by the variable bottom topography along the five profiles and the variable ranges to which the signals travelled. The signals transmitted along a deep-water path, such as path 3, begin with a sequence of sounds at arrival times in close agreement with those predicted by ray theory. Following these sounds a regular oscillatory wave train is observed in which frequency increases with time. This wave train terminates with the last RSR sound and is followed by incoherent waves reflected from the ocean floor. The sound spectrogram of Fig. 6 shows that the signals consist of a superposition of many normal modes of oscillation. Waves corresponding to each normal mode exhibit normal dispersion. At ranges greater than 1,000 km. only the first 2 or 3 normal modes are generally observed because of attenuation of waves corresponding to higher modes by the boundaries of the channel. The oscillogram of Fig. 7 shows clearly the regular oscillatory appearance of waves corresponding to the first two normal modes.

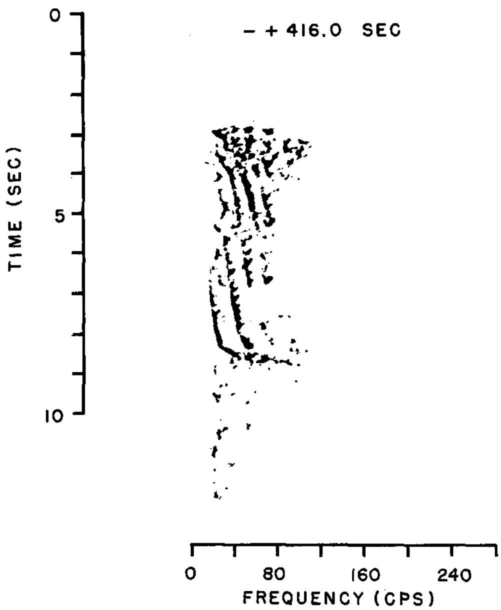

FIG. 6. Sound spectrogram of 5-lb TNT charge fired at a depth of 122 m. Hydrophone at a depth of 46 m. Waves travelled a distance of 609.4 km. along path 3 of Fig. 4.

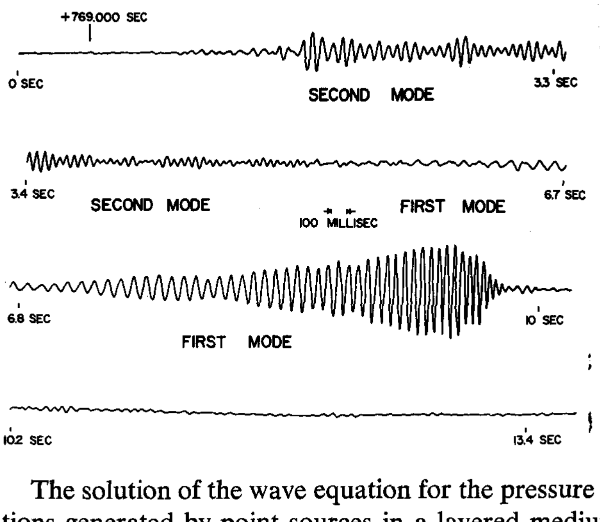

FIG. 7. Oscillogram of signal transmitted along path 6 of Fig. 4. Range 1118.2 km. 9-lb. TNT fired at a depth of 152 m. Hydrophone at a depth of 61 m. Passband of listening system 10 to 21 cps.

The solution of the wave equation for the pressure or particle velocity perturbations generated by point sources in a layered medium makes possible a detailed comparison of observations with normal mode theory. This has been shown in a convincing way by Pekeris (1948) and by Tolstoy (1955, 1958) for acoustic-wave propagation in shallow water. The formulas for an n-layered, interbedded liquid-solid half-space bounded above by a rough layer are very complex and will not be given here. Our analysis, based on the Thomson-Haskell matrix method (Thomson 1950; Haskell 1953), follows Harkrider (1964) for harmonic Rayleigh waves in an n-layered solid half-space. Layer matrices given by Dorman (1962) for computing dispersion in an n-layered liquid-solid half-space are used for the liquid layers, and they are modified at high frequencies to improve numerical precision. The solution for harmonic point sources is extended to explosive sources in the usual way and the Fourier integral for each mode is evaluated by the principle of stationary phase. The model of the underwater explosion at low frequencies and high frequencies for three bubble pulses is given by Weston (1960). Attenuation by the rough ice boundaries is taken into account by multiplying the expression for pressure or particle velocity for each mode by a modified form of the formula of Marsh (1961), Marsh *et al.* (1961), and Mellen and Marsh (1965). The attenuation factor for each mode is an exponential term which is a function of the root-mean-square (rms) ice roughness below sea level, wave frequency, phase velocity dispersion, range for one cycle of a ray as a function of frequency, surface sound velocity, and the distance between source and detector. The additional formulas and subroutines required for the solution of the wave equation in terms of normal modes are incorporated into Dorman's dispersion program for the IBM 7094 or 360 digital computers in either single or double precision arithmetic.

We shall now present some computations for Model 2 and show that the normal mode theory for the layered models predicts quite reliably the frequency and amplitude characteristics of the observed acoustic signatures for Model 2. Fig. 8 shows the range dependence of waves corresponding to the first mode when the surface is bounded by a rough ice layer. Curves of this type show that

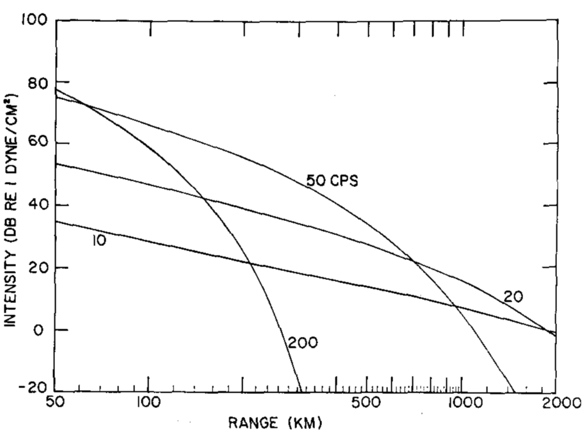

FIG. 8. Range dependence of waves of the first normal mode. Computations for Model 2 with 3 m. root-mean-square (rms) ice roughness. 5-lb. TNT charge at 150 m. Hydrophone at 50 m.

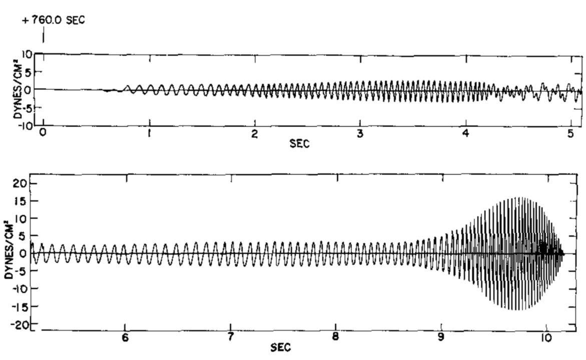

FIG. 9. Computed oscillogram of pressure variations for Model 2. The rms ice roughness 3 m. 5-lb TNT at 150 m. Hydrophone at 50 m. Range 1106.0 km.

low-frequency waves will predominate at long ranges. Fig. 9 shows a computed oscillogram of pressure variations in dynes/cm.[2] at the hydrophone for the same parameters used for computing the curves of Fig. 8. Waves corresponding to the third and higher modes are neglected since they are weak at a range of 1,106 km. compared with waves corresponding to the first two modes. Although the oscillogram was computed for a charge about half as large as the one corresponding to the signal of Fig. 7, the similarity of the two waveforms is nevertheless striking. Computations just completed specifically for the signal of Fig. 7 are in close agreement with field data. These computations include the response of the listening system and the bathymetry along the propagation path.

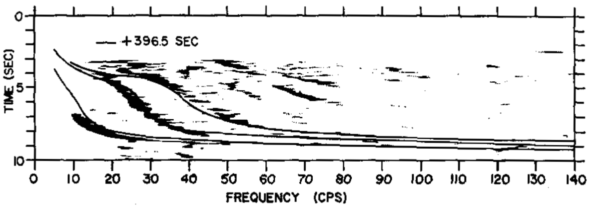

FIG. 10. Comparison of observed and computed dispersion for first three modes. Computations for model similar to Model 2 but with a water depth of 2800 m.

In Fig. 10 an observed sound spectrogram is compared with a computed one for a model similar to Model 2, but for a water depth of 2,800 m. The signal travelled approximately along the deep-water path 3 of Fig. 4. The agreement between theory and experiment is extremely good.

Fig. 11 shows peak signal intensities and peak intensities of waves corresponding to the first normal mode in the band from 25 to 70 cps as a function of range along Profiles 1 and 2 of Fig. 4. The more rapid decay of peak signal intensities along Profile 2 than along Profile 1 is probably due both to a rougher ice surface along Profile 2 and to the bottom topography on the Alpha Cordillera and continental margin. The peak signal intensity corresponds to the deep-penetrating RSR sounds which may have been weakened by reflection from the Cordillera and continental margin. On the other hand, the more rapid decay of waves of the first normal mode along Profile 2 than along Profile 1 is apparently due to the greater ice roughness along Profile 2. Fig. 11 shows computed peak signal

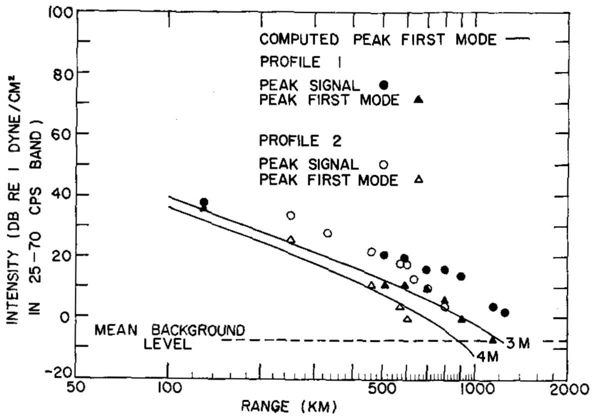

FIG. 11. Peak signal intensities and peak intensities of waves corresponding to the first normal mode as a function of range. Computations for Model 2 with rms ice roughnesses of 3 m. and 4 m. Shots 1.8-lb. TNT at 274 m. Hydrophone at 30 m.

intensities for the first normal mode in the band from 25 to 70 cps for an rms ice roughness of 3 and 4 m. The 3 m. ice roughness fits the data from Profile 1 quite nicely, while at long ranges the 4 m. ice roughness fits the data from Profile 2 reasonably well. The rms ice roughness of 3 to 4 m. is in close agreement with the analysis by Mellen (1966) of Lyon's (1961) under-ice echograms made aboard a nuclear submarine. Our data are also consistent with the transmission loss data of Mellen and Marsh (1965) analysed in terms of energy flux. These workers found that an rms ice roughness of 2.5 m. was indicated by their measurements made largely during the summer and fall months when the pack ice is often broken by large patches of open water.

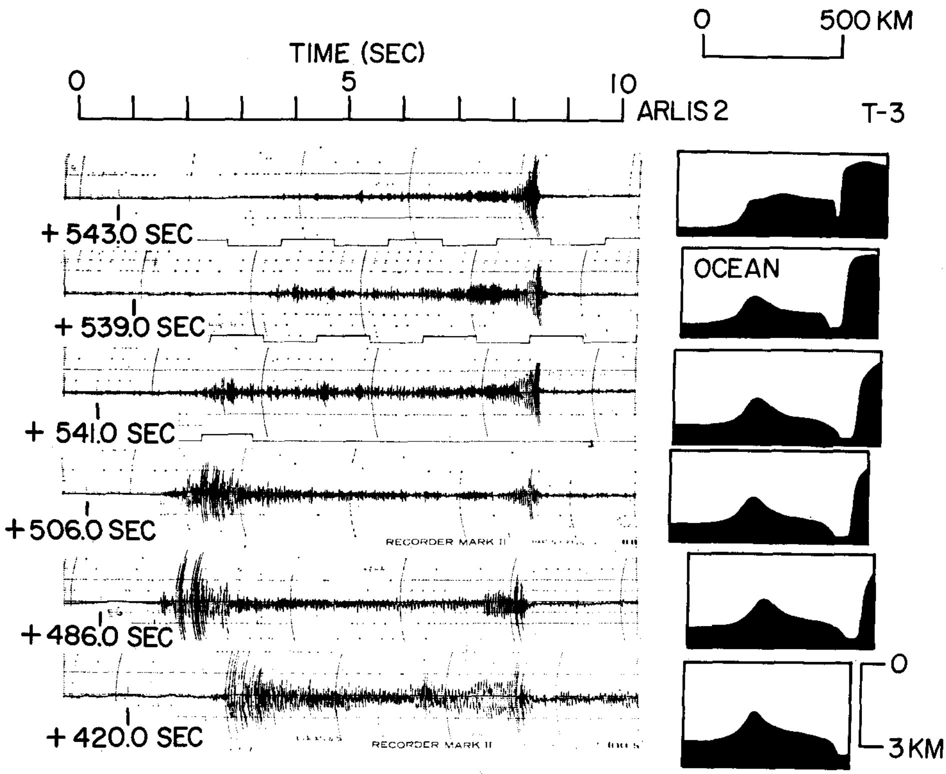

FIG. 12. Oscillograms showing effect of bottom topography on the amplitudes of the waves. Shots 1-lb. TNT at a depth of 71 m. Hydrophone at 46 m. Propagation paths between paths 2 and 3 of Fig. 4.

Fig. 12 shows the effect of bottom topography along the propagation path on the amplitudes of the signals. The sound sources were 1-lb TNT charges fired at a depth of 71 m. The hydrophone was at a depth of 46 m. The propagation paths lay between paths 2 and 3 of Fig. 4. This experiment shows that the first strong sound corresponds to an RSR ray which has passed over all bottom topography without suffering a bottom reflection. For this sequence of shots, the shallowest point along the paths is about 350 m. This corresponds to a speed of sound or phase velocity of 1,454 m. sec.$^{-1}$, and to a group velocity or mean horizontal velocity of 1,445 m. sec.$^{-1}$, which is in good agreement with the measurements.

In shallow water, sounds may be propagated to moderately long ranges by repeated reflections from the surface and bottom. The propagation is not as efficient as in deep water because of the absorption of sound in the sediments and scattering of sound from the ocean bottom. The propagation is generally quite variable depending on the water depth and the nature of the bottom. Water waves from a 1,100-lb TNT charge exploded at a depth of 16 m. were recorded at ARLIS II at a range of 163.5 km. and showed the inverse dispersion of the first normal mode in contrast to the normal dispersion observed in the deep ocean. Fig. 13 shows that this dispersion is in good agreement with that computed for the layered model given in Table 1. Fig. 14 shows diagrammatically the decrease of peak amplitude with range for three shots recorded on ARLIS II.

TABLE 1. Parameters for Computing Shallow Water Dispersion.

Layer	*Longitudinal Velocity* km. sec.$^{-1}$	*Transverse Velocity* km. sec.$^{-1}$	*Density* gm. cm.$^{-3}$	*Layer Thickness* m.
1	1.435	—	1.025	230
2	1.75	—	1.6	200
3	2.7	—	2.08	—

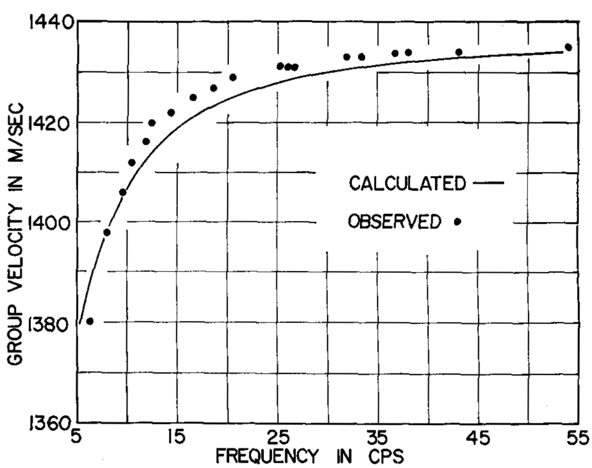

FIG. 13. Observed dispersion of waves of the first mode from several shots compared with first mode computed for model given in Table 1. From Hunkins and Kutschale (1963).

The amplitudes have been normalized to a 800-lb TNT charge. The peak amplitude, which corresponds to a frequency of 18 cps, decreases as the -1.85 power of range in the range interval from 75 to 275 km. rather than the inverse first power of range at this frequency and in this range interval which is observed in deep water. Signals from large charges fired in shallow water have been recorded aboard listening stations in deep water, and also for the reverse situation. In these cases when the length of the shallow-water path is a sizable fraction of the deep-water path, computations are made for each segment of the path separately and they are then combined for the total path.

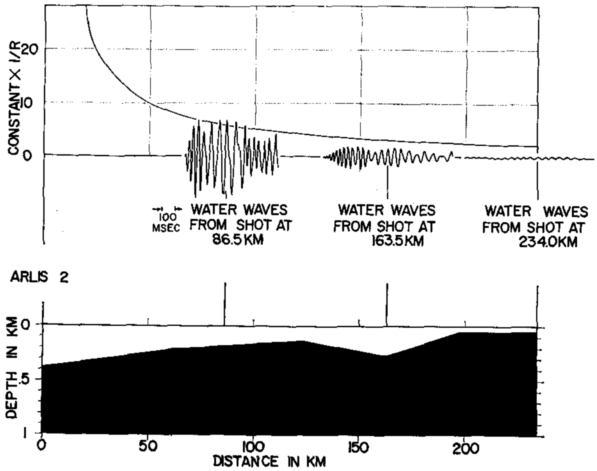

FIG. 14. Water-wave transmission loss and bathymetric profile for shots recorded on ARLIS II. From Hunkins and Kutschale (1963).

Background Noise and Reverberation

An important acoustical parameter of the ocean is the natural background noise. It does not affect sound propagation, but it is extremely important to all aspects of signal detection and processing. An important problem is to isolate the principal sources of noise, measure their strengths, and determine how the noise is propagated away from the sources. Measurements over long periods of time at many locations are necessary to determine dependence of the noise on time, location, and direction.

The principal source of noise in the Arctic Ocean is the ice cover. This ice is in continual motion under the influence of winds and currents and thousands of tons of ice may be displaced vertically and horizontally when a large pressure ridge is formed or a floe breaks up. The air-borne sounds generated by this ice-movement are often heard by ear up to 1 km. from the active area. The sound is commonly a low-frequency rumble. Ice vibrations may also be felt under foot if one is standing on the floe which is breaking up or where a pressure ridge is being formed. Besides these large-scale ice movements, the ice may be under sufficient stress to induce small ice quakes. This is particularly common when the ice is under thermal stress during periods of rapid temperature drop in the spring and fall. The air waves generated by these ice quakes have a snapping sound and their frequency of occurrence may be over one per second. Other sources of noise that may be heard at times are wind-blown snow moving over the ice and gravity waves splashing in open leads.

The natural background noise on the ice and at depth is quite variable in strength. This is to be expected since the noise level depends largely on the relative motion of the pack ice in the immediate area under investigation, and this motion may be from practically zero when all floes are moving with a uniform velocity to highly variable velocities of neighbouring floes during break-up and pressure ridging. The strength of the background noise does not always correlate with the local wind speed, but there is a higher probability of high noise levels during storms than during periods of prolonged calm. The noise may also have a directional character, depending on the locations of the principal ice activity. When local ice activity is low and the noise level is correspondingly low, sounds may arrive from considerable distances travelling in the sound channel. Under these conditions the noise level falls off rapidly with increasing frequency because of attenuation of the high-frequency waves by scattering.

The background noise on the ice and at depth in the ocean is measured by single detectors and vertical and horizontal arrays of detectors. Recordings from a single detector made over long periods of time yield information on the time variation of the noise. Recordings from arrays provide information on the directional properties of the noise and permit an identification of the types of waves present in the noise. Hydrophones at depth pick up sounds travelling along paths like those of Fig. 2. On a typical day, the scraping and grinding of ice may be heard on a loud speaker interrupted occasionally by explosion-like sounds from ice quakes. At night in the spring and fall thermal cracking may be so strong and frequent that noise from other sources above 20 cps is blocked out. The hissing sound of wind-driven snow is heard during storms in the cold months and during the warm months sounds from marine mammals are sometimes heard. On a very quiet day, even the splashing of waves in a nearby lead may be audible.

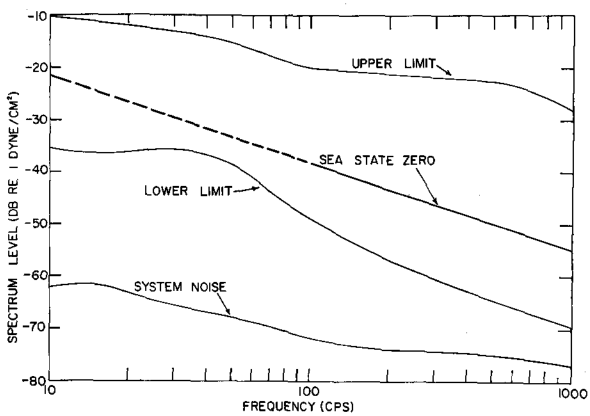

FIG. 15. Upper and lower limits of spectrum levels of water noise. From Mellen and Marsh (1965) and Buck (1968).

Fig. 15 shows the range of spectrum levels measured by Mellen and Marsh (1965) and Buck (1968) with a hydrophone at depths between 30 and 61 m. For comparison, the sea state zero curve of Knudsen *et al.* (1948) extrapolated to 10 cps is also shown. The range of variation of noise levels is more than 25 db, although the average level appears to lie about 6 db above the Knudsen zero sea state curve.

Noise levels in the water generally build up and decay over periods of at least a day, but the ice vibrations may fluctuate by more than 40 db over periods of less than one hour. The waterborne sounds come from many active floes, while the strong ice vibrations are generally confined to the floe on which the seismometer is located. The dominant ice vibrations correspond to flexural waves generated by ice movement at the boundaries of the floe and as this movement increases in magnitude and then weakens so do the flexural waves. The flexural waves are surface waves travelling in the ice sheet and, therefore, pressure perturbations decay exponentially with depth. Only seismometers on the ice and hydrophones directly beneath the ice detect these waves; they are identified by their characteristic inverse dispersion and by the particle motion of the waves. When the movements at the boundaries of the floe are small, then the principal source of noise may be sound transmitted through the water from other active floes, either near or distant. Fig. 16 shows vertical particle motion measured in octave bands from data taken on ARLIS II (Prentiss *et al.* 1966). For comparison, the curves of Brune and Oliver (1959) for land noise and Hunkins (1962) for the Arctic Ocean are shown in the figure. The noise curves of Prentiss *et al.* are for average levels, not for bursts of noise occurring during particularly active times. Even so, the variation of noise levels is more than 30 db, corresponding to levels ranging from very quiet land sites up to noisy land sites.

The total scattering effect from inhomogeneities in the ocean is called reverberation. Reverberation may be subdivided into surface reverberation, volume reverberation, and bottom reverberation. Some aspects of scattering from the surface and bottom were discussed under Propagation; following is a brief description of volume reverberation from the Arctic scattering layer.

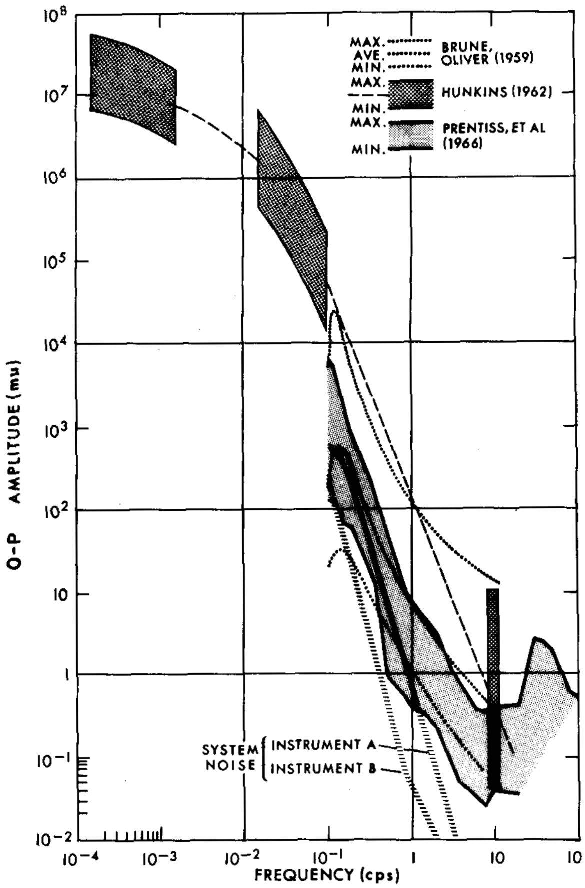

FIG. 16. Ambient ice vibrations recorded on ARLIS II analysed in octave bands. System B was used to record during low noise levels.

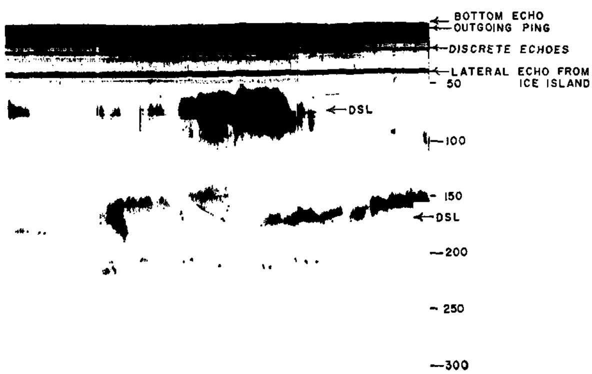

FIG. 17. Precision depth recording of arctic deep scattering layer (DSL). From Hunkins (1966).

The deep scattering layer (DSL) in the central Arctic Ocean was discovered by Hunkins (1965, 1966) in June 1963, aboard Fletcher's Ice Island, T-3. Fig. 17 shows a typical recording of the layer. Precision depth recordings of the layer made up to the present time aboard T-3 at a sound frequency of 12 kc have revealed two important features which distinguish this layer from that observed in the non-polar oceans. The arctic layer occurs at a moderately shallow depth of between 50 and 200 m., and it exhibits an annual rather than a diurnal cycle. This character is apparently a response of the scattering organisms to the unique light conditions present in the Arctic. Light is relatively weak under the ice so that the organisms can find the safety of darkness at moderate depths and the cycle of light and darkness is annual. Consequently the layer is present during the summer and disappears during the winter. At times the arctic scattering layer divides into two or three layers which is similar to what occurs in the other oceans. In addition to the scattering layers, discrete echoes from shallow depths above the layers are commonly observed throughout the year, although they are particularly frequent during the winter months. Presumably, these reflectors correspond to fish or seals. To date, the organisms producing the scattering layers have not been identified but there is some evidence that they may be siphonophores.

FUTURE RESEARCH

Research Platforms

Research in arctic hydroacoustics during the next twenty years will probably continue to be carried out aboard drifting ice stations. Consideration must be given to improving these platforms as listening sites. A major problem of the past has been man-made noise at the stations. This noise comes from heavy equipment used in normal camp operations and from apparatus used by investigators working in other fields of research. Two solutions of this noise problem are possible. One is to move the listening site far enough from the areas of activity so that the noise is negligible. The other is to establish a drifting station which is designed especially for quiet operation. In both cases maintenance of a camp is required. The latter possibility is particularly attractive for future work, since a station in addition to Fletcher's Ice Island, T-3 is required for detailed experiments on sound transmission. The new station should preferably be on pack ice to provide acoustic data under typical ice conditions in the central Arctic Ocean. A suitable initial location for the station would be near the centre of the Canada Abyssal Plain. This station would remain over the Plain for a reasonable period of time whereas T-3 might pass over other bathymetric features. Every effort should be made aboard the new station to minimize man-made noise. Generators should be quiet, but provide reliable power for the equipment. The camp should be small and mobile so that in the event of breakup it could be moved to another suitable floe in the area. In addition to serving as a listening platform for long-range transmission work, the new station would be particularly suited for investigations of ambient noise, reverberation from the underside of the pack ice, short-range transmission in the ice and water, and as a test platform for new hydroacoustic apparatus. For sound transmission experiments, explosives could be launched from both stations, but high-power harmonic sound sources should be installed aboard T-3 or the successor to this ice island to avoid noise at the quiet station. Satellite navigation would provide precise positions of both stations. Listening could continue aboard T-3 at sites sufficiently far removed from the main areas of man-made noise.

Oceanography

Of basic importance to the interpretation of hydroacoustic experiments are the physical properties of the medium in which the waves travel. Hydrographic stations provide the data to determine in detail the seasonal and regional variations of sound velocity with depth. Although a considerable body of data has been obtained in the past from drifting stations, more area can be covered by aircraft landings on the ice. A worthwhile project for the future is to measure sound velocity profiles at many locations with a portable velocimeter carried to the station by aircraft. Ocean currents and internal waves in the upper layers may have important effects on sound intensities at high frequencies. Continuous measurements of currents and temperatures at depth should be made simultaneously with a transmission experiment between a fixed transducer and hydrophone separated by perhaps a kilometre or two. Drifting stations provide a unique opportunity for such a detailed experiment.

More data on bottom and under-ice topography are required for investigating effects of variations in this topography on sound transmission. Ice roughness varies both seasonally and regionally and therefore significant differences in signal strength at different seasons and locations are expected. For computations of bottom-reflected sounds by ray or mode theory, the bottom topography along the transmission path must be known. Only in a few cases are there sufficient data to model the bottom topography even approximately. Although ice stations will continue to supply high-quality precision depth recordings, only measurements from nuclear submarines can provide the regional coverage required and data on both the upper and lower boundaries of the ocean. Under-ice echograms obtained from submarines should be analysed in detail to determine the roughness spectra of the underside of the ice as a function of location and season. These data might be supplemented with transmission profiles made along the submarine track either by launching small charges from the submarine or from an aircraft.

The elastic constants of sea ice are established from seismic experiments (Hunkins 1960), but our knowledge of the elastic properties of the bottom sediments in the central Arctic Ocean is meagre. Most of the information we have comes from sound velocity and density measurements made on bottom cores. This work should be supplemented with wide-angle reflection and refraction profiles made periodically aboard a drifting station to measure the velocity distribution in the sediments.

At phase velocities greater than the speed of sound in water, sounds travel in the ice by repeated reflections from the upper and lower boundaries of the ice. High-frequency waves are attenuated strongly not only by scattering from the boundaries, but also by inhomogeneities in the sea ice. Laboratory experiments on attenuation in ice have been made and these data should be supplemented by measurements in the field at various frequencies and at different times of the year.

Propagation

Experiments on long-range explosive sound transmission should be designed to investigate the effects of variations of ice roughness and bottom topography on signal strength. These experiments can be efficiently carried out with aircraft dropping charges into open leads. The location of the listening station is important and it should be over an abyssal plain so that some profiles are over a flat bottom and others show the effects of bottom topography at one end of the profile only. Propagation experiments made on a year-round basis between two drifting stations might reveal significant variations of signal strength which could be explained in terms of ice conditions, bathymetry, and small variations in the velocity structure in the upper layers of the ocean. More data should also be obtained between two drifting stations for the variation of pressure level with source or detector depth. These measurements are best made by keeping the shot depth constant and varying the hydrophone depth to at least 800 m.

Computations by normal-mode and ray theory should be compared with measurements of sound fields from harmonic sound sources at frequencies from 10 cps upward. Measurements at various frequencies and detector depths made between two drifting stations positioned by satellite navigation would provide detailed data on the variation of pressure with depth and frequency and on attenuation of waves by the ice boundaries as a function of frequency. The measurements could be repeated periodically to obtain the range dependence of the sound field at the operating frequencies. These data might be supplemented by a transmission run by submarine. This type of profile has the great advantage of a continuous record of sound pressure level as a function of range, together with the important data on the shape of the surface and bottom.

Background Noise

Long-term measurements of background noise are being carried out aboard Fletcher's Ice Island, T-3 (Buck 1968). Measurements of this type should also be made aboard a quiet pack ice station under typical ice conditions. The noise levels must be carefully examined in terms of environmental conditions, such as local winds, ice movement, and air and ice temperatures. Measurements employing horizontal arrays of hydrophones and seismometers provide data on the directional properties of the noise. The interpretation of these data is greatly assisted by air photographs made periodically over the area under investigation to determine the active areas of ice movement. A promising new tool for investigating the regional variation of noise is the IRLS system. This remote sensing platform transmits data via satellite to a distant manned station. An experiment in the spring of 1969 aboard T-3 was designed to establish the possibility of using this platform to gather noise data at unattended sites in the Arctic Ocean.

A problem of basic importance to computing theoretical noise spectra in terms of propagation models is to know the spectral characteristics of the noise sources. Portable listening equipment installed at active pressure ridges and leads would provide such data for the large-scale ice movements. Milne (1966) has computed the spectral characteristics of thermal cracking for a model of the sources and he has obtained reasonably good agreement with measurements made under shore-fast ice. Thermal cracking of sea ice investigated in the laboratory under controlled conditions might provide useful data to determine the critical temperature gradient required for the onset of cracking and the peak amplitude distribution of the ice tremors.

Reverberation

A hydrophone near an underwater explosion detects strong reverberations from the underside of the ice. It is expected that these reverberations are strongly dependent on local under-ice topography. This appears to be the case for the data of Mellen and Marsh (1963), Milne (1964), and Brown (1964). The data of Milne and Brown, although not in agreement, show an increase of scattering strength with frequency and grazing angle, while the data of Mellen and Marsh indicate an absence of frequency dependence. Marsh and Mellen have derived the ice roughness spectrum from their data, but this spectrum is in poor agreement with the spectrum computed by Mellen (1966) from Lyon's under-ice echograms. More measurements should be made at various locations and seasons of the year for comparison with predictions made by theory for models of the under-ice topography.

Drifting stations provide ideal platforms for experiments on volume reverberation. The seasonal and regional characteristics of the DSL may be measured along the drift path from recordings made over periods of years. Biologists can probe the layers for samples of the scattering organisms. Present sounders aboard T-3 operate at 12 kc. and 100 kc. These measurements should be continued and extended to other frequencies. Reverberation levels as a function of depth at a number of frequencies should be measured periodically aboard the stations.

Strong bottom-reflected sounds are commonly observed over abyssal plains at ranges up to 500 km. from an explosion. In almost all cases over rough topography, bottom reverberation is observed at long ranges beginning with the onset of the RSR sounds and continuing many seconds after the last RSR sound has passed. The sound spectrogram of Fig. 6 shows this reverberation clearly. Future experiments should measure the level and spectral character of this reverberation as a function of range, bathymetry, charge size, and charge depth. The data at long ranges should be supplemented with bottom reverberation measured at the ice stations from local explosions as the stations move over different bottom topography. Precision depth recordings and ocean-bottom cores will aid in the interpretation of the reverberation data.

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