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Scientific session of the Division of General Physics and Astronomy of the Russian Academy of Sciences (25 January 1995)

A scientific session of the Division of General Physics and Astronomy of the Russian Academy of Sciences was held on 25 January 1995 at the P L Kapitza Institute of Physical Problems. The following papers were presented at this session:

(1) **K D Sabinin** (N N Andreeva Acoustics Institute, Moscow) "Oceanological aspects of acoustic thermometry of the Arctic Ocean";

(2) **A N Gavrilov** ('Ocean Acoustics and Information' Scientific Company; N N Andreev Acoustics Institute, Moscow), **M M Slavinskii** (Institute of Applied Physics, Russian Academy of Sciences, Moscow), and **A Yu Shmelev** ('Ocean Acoustics and Information' Scientific Company; Institute of General Physics, Russian Academy of Sciences, Moscow) "Theoretical and experimental investigations of the feasibility of acoustic thermometry of climatic changes in the Arctic Ocean".

Summaries of these papers are given below.

PACS numbers: 92:10.Bf; 93.30.Li; 93.85.+q; 43.30.Pc

Oceanological aspects of acoustic thermometry of the Arctic Ocean

K D Sabinin

The first experiment on very-long-range propagation of sound in oceans was carried out in 1960: underwater explosions of 300 lb charges off the coast of Western Australia were recorded by hydrophones off the Bermuda Islands, i.e. after the acoustic signal travelled half way round the Earth, passing through the Indian Ocean and a large part of the Atlantic [1]. The ability of the ocean to concentrate the acoustic energy in an underwater acoustic channel, which makes possible very-long-range propagation of acoustic signals, has led to the suggestion that the travel time of acoustic pulses through oceans should be used to measure the average temperature and to detect global warming of the Earth because of the greenhouse effect [2]. The changes in the velocity of sound in oceans are almost entirely governed by changes in temperature, and the average temperature of water along transoceanic paths (particularly if there are several) provides an excellent measure of the thermal state of the oceans, reflecting the thermal state of the planet Earth as a whole.

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The first trial was carried out in 1990 as part of the ATOC (Acoustic Thermometry of Ocean Climate) programme. Signals generated near the Heard Island (in the southern part of the Indian Ocean) were received at various parts of the World Ocean, including both coasts of the USA: Atlantic and Pacific [3]. The acoustic power of the source operating at 57 Hz was 209 dB, relative to 1 μPa at a distance of 1 m. A study was made of the various forms of signals with a sufficiently wide spectrum, ensuring accurate measurement of the travel times of the signals along the path. If $\overline{C}$ is the average velocity of sound along a path of length L, then the relative error in the velocity measurements is $\Delta C/\overline{C} \approx \Delta\tau/\tau$, where τ is the travel time of the signal measured to within $\Delta\tau$. If $L = 10^4$ km and $\Delta\tau = \pm 10^{-2}$, then $\Delta C = \pm 2 \times 10^{-3}$ m s^{-1}, which corresponds to measurements of the average (along the path) temperature to within $\pm 5 \times 10^{-4}$ °C. Many countries participated in this experiment, which confirmed the feasibility of global acoustic thermometry of the oceans. These countries included also the Soviet Union (the vessel of the Acoustics Institute 'Akademik Nikolai Andreev' received the Heard Island signals in the northern tropics of the Atlantic Ocean).

An important and successfully developing part of the international ATOC programme had become its Arctic component: Arctic ATOC. The special nature of the Arctic Ocean, justly known as the 'weather kitchen', makes it particularly attractive for acoustic monitoring because:

— the Arctic has an influence on the climate and weather over the whole Earth;

— simulation results (see, for example, Ref. [4]) predict the greenhouse warming of the Earth to be strongest in the Arctic;

— deep-water inhomogeneities (fronts, vortices, internal waves) in the Arctic Basin are much weaker than the temperature zones;

— the permanent ice cover of the Arctic can be regarded as an independent and very informative object for acoustic monitoring.

On the other hand, some characteristic features of the Arctic Ocean make it difficult to detect the greenhouse effect and require a special approach both in respect of organisation of acoustic monitoring and in the interpretation of its results. First, the natural variations of the Arctic Ocean over a period of many years are much stronger than the corresponding variations of the oceans in the temperate zones. This means that the advantages over the simulation-predicted enhancement of the greenhouse signal may be lost completely because of the 'noise' of natural variations.

In addition to its main distinguishing feature, which is a permanent ice cover of many metres in thickness, the Arctic

Ocean has the following specific characteristics: strong freshening (desalination) of the upper layer because of the large amounts of water arriving from rivers (about 5×10^3 km^3 per year); a relatively warm and saline layer of waters of Atlantic origin, located between the upper and deep layers and characterised by a negative temperature. The Atlantic waters arrive mainly through the Fram Strait with the West Spitsbergen Current. These waters pass West Spitsbergen on the north side, 'dive' under the surface waters of the Arctic Basin, and fill the whole of this Basin at moderate depths from several hundred metres to one kilometre (Fig. 1). They spread in the form of a giant cyclonic circulation with numerous branches: in this way the Atlantic waters lose some of their heat and become transformed before they leave the Arctic Basin. The total loss of heat by the Atlantic waters as a result of mixing with the upper layer is prevented by a sharp density discontinuity at the upper boundary above which the water is freshened by river drainage. This density discontinuity in the Arctic Basin is responsible for the permanent ice cover, because without the screening influence of this layer there would be sufficient heat to melt ice completely [5].

Let us now consider a possible reaction of the Arctic Ocean to the greenhouse warming and how this can affect the travel time of acoustic signals in this Ocean.

The most obvious warming effect is the melting of ice, which reduces its thickness and redistributes the surface layer of the Ocean. The change in the ice thickness does not affect directly the signal propagation time τ, since the freshening of the upper layer of the 'ocean' associated with the melting of ice reduces the velocity of sound C, i.e. increases τ. In fact, since the temperature of the upper layer of water in the Arctic Basin is always close to freezing point, which increases only slightly because of a reduction in the salinity, the changes in the velocity of sound in this layer are almost entirely due to changes in the salinity S and not of the temperature T, as is true of ice-free oceans. The total derivative, representing the change in the velocity of sound C with the freezing point Θ, is

$$\frac{\mathrm{d}C}{\mathrm{d}\Theta} = \frac{\partial C}{\partial T} + \frac{\partial C}{\partial S}\frac{\mathrm{d}S}{\mathrm{d}\Theta}. \tag{1}$$

Let us substitute here the values $\mathrm{d}C/\partial T = +4.7$ m s^{-1} °C^{-1}; $\partial C/\partial S = +1.3$ m s^{-1} (‰)$^{-1}$, and $\mathrm{d}S/\mathrm{d}\Theta = -18$ ‰ °C^{-1}, where ‰ represents thousandths parts used to measure the salinity of seawater; °C denotes degrees Celsius. This gives $\mathrm{d}C/\mathrm{d}\Theta = -18.7$ m s^{-1} °C^{-1}; i.e. instead of the usual (for ice-free oceans) increase in C by 4–5 m s^{-1} as a result of warming by 1 °C, we have here a much stronger reduction in the velocity of sound.

It follows that acoustic consequences of the warming of the upper layer of ice-free and Arctic oceans are directly opposite; in the former case there is the usual increase in C, whereas in the latter case the velocity C is reduced by the freshening of water caused by the melting of ice.

If the greenhouse effect produces a flow of heat to the surface of the Arctic Ocean at a rate of 2 W cm^{-2} [2], this amount is sufficient to melt a metre-thick ice layer in 4.7 years, which will result in the freshening of the top 50-m layer of the Ocean by 0.5‰ and a corresponding reduction in the velocity of sound in this layer by 0.6 m s^{-1}. The same heat arriving at the ice-free ocean surface can heat a 200-m layer of water by 0.35 °C. Let us assume that this is the increase in temperature of the core of the Atlantic waters in the European part of the Arctic Basin. The temperature of the Atlantic waters falls as these waters move away from the point of intrusion to the north of Spitsbergen. Therefore, the influence of the thermal impulse which these Atlantic waters have acquired during their contact with the atmosphere at Spitsbergen decreases in the

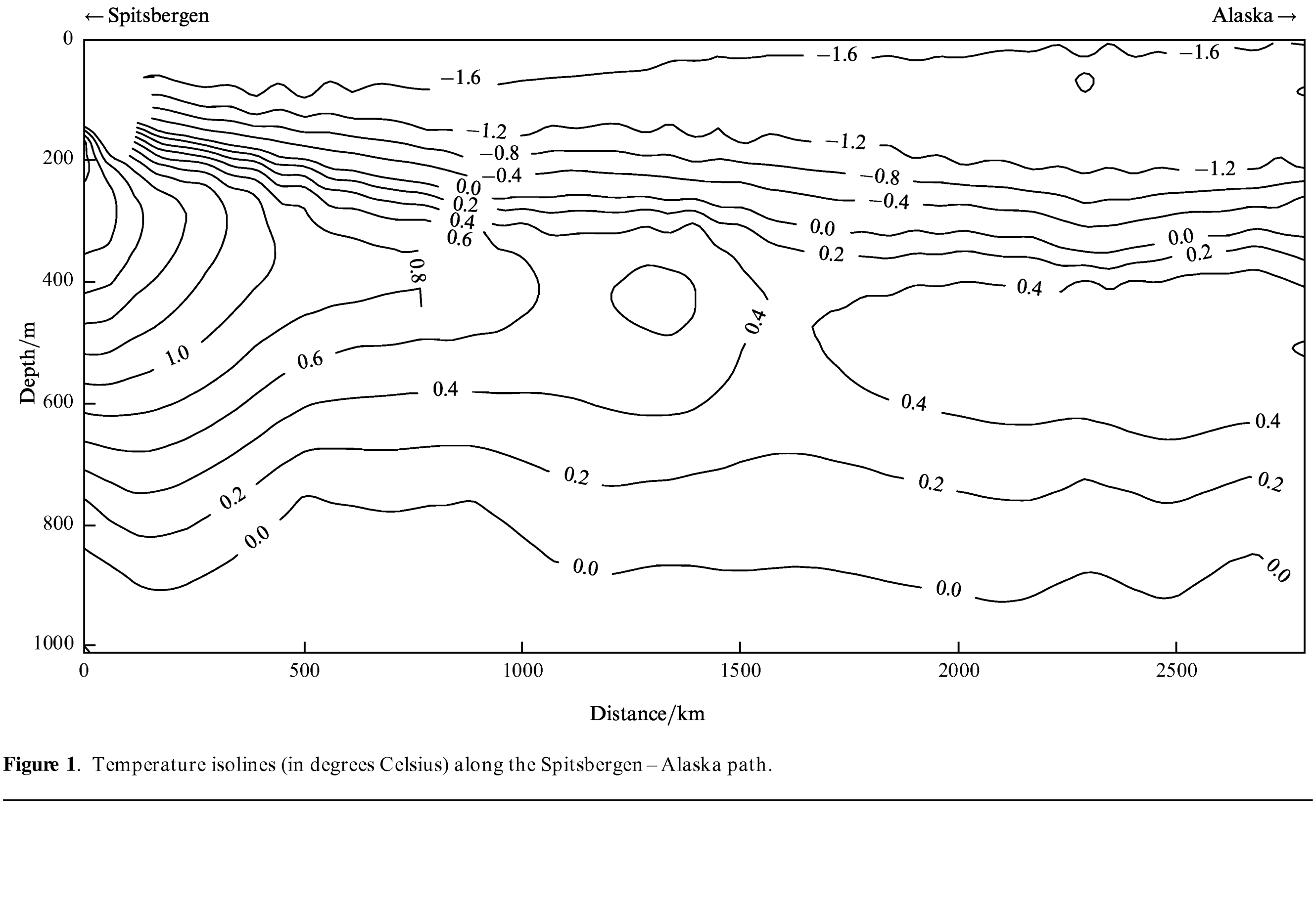

Figure 1. Temperature isolines (in degrees Celsius) along the Spitsbergen – Alaska path.

same proportion. In view of the above ideas and some of the data on the thermal conditions in the Arctic Basin on both sides of the Lomonosov ridge, separating the main proposed Spitsbergen – Alaska acoustic monitoring path into the European and Canadian parts, we can assume that the warming of the core of the Atlantic layer by +0.35 °C in the European part should be followed by a somewhat smaller warming of the core by +0.14 °C in the Canadian part. Such warming corresponds to an increase in the velocity of sound by 1.5 m s^{-1} in the European part and by 0.6 m s^{-1} in the Canadian part of the path.

Let us consider as our base model the simplest approximation of the field of the velocity of sound C along the path by $C(z)$ profiles of two types, of which the Canadian type differs from the European type simply by a greater depth of the core of the Atlantic waters (first two columns in Table 1, where the values for the Canadian part of the path are given in the parentheses). Let us now estimate how the travel time of an acoustic signal varies along the path if the following events take place:

— the melting of a metre-thick ice layer accompanied by freshening of the upper layer of water (model 1);

— melting of ice and warming of the core of the Atlantic layer in the European part of the path by +0.35 °C (model 2);

— melting of ice and warming of the whole path by +0.35 °C and +0.14 °C in the European and Canadian parts (model 3).

Table 1. Model profiles of the velocity of sound $C(z)$ in the European and Canadian (the latter are the values in parentheses) parts of the Spitsbergen – Alaska path.

Base model		Change in the velocity of sound (ΔC/m s^{-1}) predicted by models		
z/m	C/m s^{-1}	1(melting of ice)	2 (melting and warming of European part)	3 (melting and warming of whole part)
0	1440	−0.6 (−0.6)	−0.6 (−0.6)	−0.6 (−0.6)
50	1440	−0.6 (−0.6)	−0.6 (−0.6)	−0.6 (−0.6)
300 (400)	1459		+1.5 (0)	+1.5 (+0.6)
1000	1462			
3000	1500			

Calculations based on these models were carried out in accordance with the Gavrilov – Kudryashov programme [6] for three modes of an acoustic signal of 30 Hz frequency. The results of the calculations (Table 2) demonstrate that the changes in the travel time of this signal along the Spitsbergen – Alaska path (which is about 2700 km long), associated with the melting of ice and warming of the Atlantic layer, can be very considerable. The sign of these changes for the first mode is opposite to the sign for the other modes: the lowest mode, travelling mainly in the surface layers, is very sensitive to freshening of these layers, whereas the higher modes react mainly to the warming of the Atlantic layer.

Table 2. Changes in the travel time (in seconds) of an acoustic signal of frequency 30 Hz along the Spitsbergen – Alaska path, calculated on the basis of different models and compared with the base model.

Mode	Model		
	1	2	3
I	+0.8s	+0.8s	+0.8s
II	+0.6s	−0.5s	−0.5s
III	+0.2s	+0.2s	−1.0s

The heat content of the Atlantic layer is thus a very important factor in acoustic thermometry of the Arctic Ocean. In view of the main task of the Arctic ATOC programme, it is important to estimate the effect of these changes on the greenhouse warming of the Arctic. This can be done on the basis of the following simple model.

Let us assume that $q_1 = c\rho T_1 w_1$ is the heat flux contributed to the Arctic Basin by the West Spitsbergen Current. Here, c, ρ, and T_1 are, respectively, the specific heat, density, and temperature of water; w_1 is the flow rate of water, i.e., it is the volume of the Atlantic waters transported by the Current per unit time. The heat carried away by the transformed Atlantic waters from the Arctic Basin will be identified by the index 2: $q_2 = c\rho T_2 w_2$. If the interest is only in the mean (over the whole Arctic Basin) temperature of the Atlantic layer $\overline{T}$, the loss of heat from this layer can be attributed to the flow of heat to the colder higher and lower layers. Such flow is proportional to the vertical gradients of the water temperature T:

$$q_3 = c\rho\nu\sigma\left|\frac{\partial T}{\partial z}\right|_3 \quad \text{and} \quad q_4 = c\rho\nu\sigma\left|\frac{\partial T}{\partial z}\right|_4 , \qquad (2)$$

where ν is the turbulent diffusion coefficient, assumed to be independent of the depth z; σ is the area of the Atlantic layer in the Arctic Basin; the indices 3 and 4 identify the upwards and downwards flow of heat, respectively. The heat stored in the Atlantic layer is $Q = c\rho\sigma\Delta z\overline{T}$, where Δz is the layer thickness. The change in this heat is determined by all the four types of heat flow mentioned above:

$$\frac{dQ}{dt} = q_1 - q_2 - q_3 - q_4 , \qquad (3)$$

where t is time.

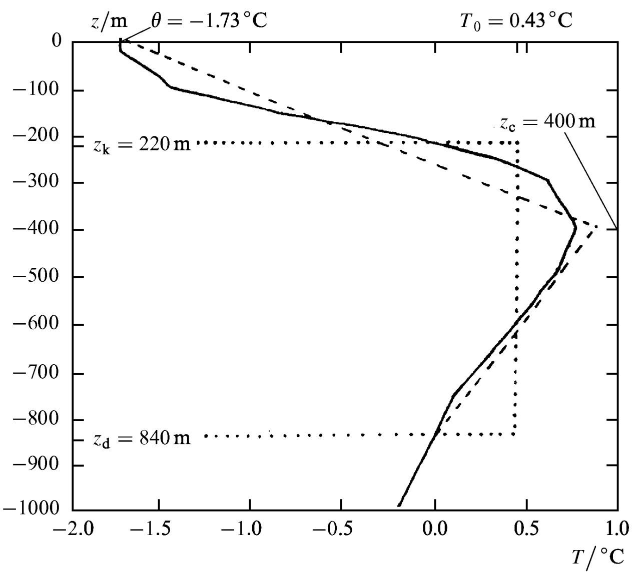

Figure 2. Temperature profile averaged over five proposed acoustic paths.

Fig. 2 shows the average (for the proposed acoustic monitoring paths) temperature profile in the Arctic Basin found by averaging the results of oceanographic measurements obtained in the course of the POLEKS programme [7]. This figure gives the mean positions of the upper ($z_u = 220$ m) and lower or deep ($z_d = 840$ m) boundaries of the layer of the Atlantic waters in the Arctic Basin. Fig. 2 identifies also the maximum temperature horizon ($z_c = 400$ m) where the temperature of the Atlantic waters also reaches its highest value. We can see that the highest temperature of the Atlantic waters is almost twice as large as their average temperature ($\overline{T}_0 = 0.43$ °C). The dashed lines are used to estimate the temperature gradients by approximation of the profile, which gives

$$\left|\frac{\partial T}{\partial z}\right|_3 = \frac{2\overline{T}\Theta}{z_c} \quad \text{and} \quad \left|\frac{\partial T}{\partial z}\right|_4 = \frac{2\overline{T}}{z_d - z_c}\,,$$

where Θ is the freezing point in the upper layer.

Substitution of these expressions in Eqn (3) on the assumption that the rate of flow of the Atlantic waters is the same at the entry and exit from the Arctic Basin, i.e. that $w_1 = w_2 = w$, finally yields

$$\frac{\mathrm{d}\overline{T}}{\mathrm{d}t} + a\overline{T} = g(t)\,, \tag{4}$$

where

$$a = \frac{2\nu z_d}{z_c(z_d - z_c)\Delta z} \quad \text{and} \quad g(t) = \frac{w}{\sigma\Delta z}\,\Delta T(t) + \frac{\nu\Theta}{z_c\Delta z}, \tag{5}$$

and $\Delta T = T_1 - T_2$.

The solution of Eqn (4) is known:

$$\overline{T}(t) = \exp(-at)\left[c + \int g(t)\exp(at)\,\mathrm{d}t\right]. \tag{6}$$

The expressions for a and $g(t)$ include a parameter difficult to measure, which is the vertical diffusion coefficient ν. We can determine it from the average parameters of the temperature profile and from the heat and water exchange in the Arctic Basin on the assumption that they all correspond to an equlibrium state when $\mathrm{d}\overline{T}/\mathrm{d}t = 0$. It then follows from Eqn (3) that

$$\nu = \frac{\Delta T_0\Delta z}{t_r\left(2\overline{T}_0 - \Theta/z_c + 2\overline{T}_0/z_d - z_c\right)}, \tag{7}$$

where $t_r = \sigma\Delta z/w$ is the time needed for complete renewal of the Atlantic layer and the index 0 represents the initial (equilibrium) conditions.

If we assume that the temperature T_1 of the Atlantic waters increases linearly at the entry to the Arctic Basin, i.e. at the shores of Spitsbergen where these waters are still in contact with the atmosphere, then the difference $\Delta T = T_1 - T_2$ will also increase linearly with temperature during the time interval $t < t_r$ until the water-warming front traverses the whole path inside the Arctic Basin and the temperature T begins to rise at the exit from the Basin. Substitution of $\Delta T(t) = \Delta T_0 + \alpha t$, where ΔT_0 is the initial temperature difference $\alpha = \partial T_1/\partial t$, in expression (6) yields (after some simple transformations) the following expression for the increase in the average temperature of the Atlantic layer in the Arctic Basin:

$$\overline{T}(t) = \frac{\Theta}{2}\left(1 - \frac{z_c}{z_d}\right) + \frac{\Delta T_0}{at_r} + \frac{\alpha}{a^2 t_r}(at - 1)$$
$$+ \left[\overline{T}_0 - \frac{\Theta}{2}\left(1 - \frac{z_c}{z_d}\right) - \frac{\Delta T_0}{at_r} + \frac{\alpha}{a^2 t_r}\right]\exp(-at)\,. \tag{8}$$

Let us now substitute specific values of the quantities that occur in formula (8). It follows from the average equilibrium temperature profile of the Atlantic layer (Fig. 2) that $\overline{T}_0 = 0.43$ °C, $z_c = 400$ m, $z_d = 840$ m, $\Delta z = 620$ m, and $\Theta = -1.73$ °C. Various estimates have been put forward for the rate of flow of the Atlantic waters into the Arctic Basin: they range from 2×10^6 to 6×10^6 m^3 s^{-1} [8]. If we assume that $w = 2 \times 10^6$ m^3 s^{-1}, we obtain the following estimate for the renewal time of the Atlantic layer: $t_r = \sigma\Delta z/w = 46$ years, if σ is the area of the Arctic Basin above depths in excess of 200 m, i.e. $\sigma = 4.5 \times 10^6$ km^2. The rate of temperature rise $\alpha = \partial T_1/\partial t$ can be estimated on the assumption of the postulated acquisition of heat from the atmosphere because of the greenhouse effect: $\Delta q \approx 2$ W m^{-2} [2]. Such heat flow increases the temperature in a 600-m layer of the water in the West Spitsbergen Current (i.e. the Atlantic Waters entering the Arctic Basin) at the rate $\alpha = 0.025$ °C year^{-1}. Observations show that the Atlantic waters are cooled by 1.5 °C – 2 °C during the time they spend in the Arctic Basin. If we assume that $\Delta t_0 = 1.7$ °C and use the above values of the other parameters which occur in formula (8), we find that in ten years the average temperature of the Atlantic layer should rise by 0.024 °C (Fig. 3).

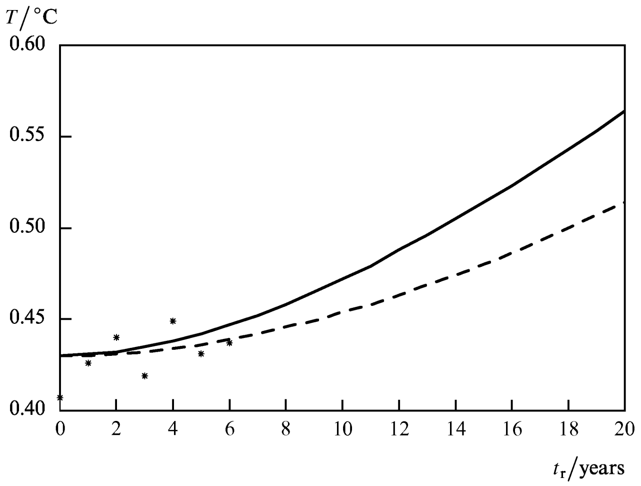

Figure 3. Warming of the Atlantic layer calculated for renewal times $t_r = 23$ years (continous curve) and 46 years (dashed curve), compared with the average temperature along acoustic paths recorded during each year of observations in 1973 – 1979 (asterisks).

An investigation of the dependence of the solution on the values of the parameters employed, carried out by S V Pisarev, has shown that the characteristics of the temperature profile and the value of ΔT_0 have hardly any effect on the results of calculations, since the acceleration of water exchange (reduction in t_r) and heating at the shores of Spitsbergen (increase in α) increase considerably the rate of warming of the Atlantic layer in the Arctic Basin (Fig. 4).

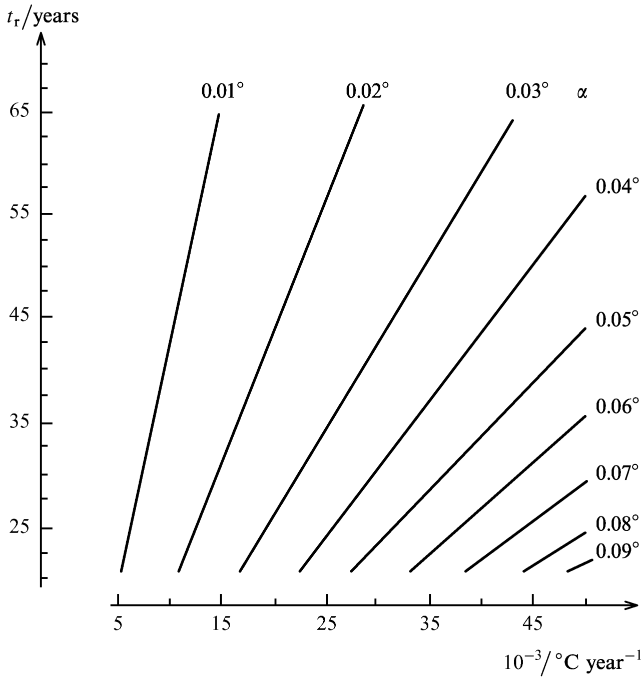

Figure 4. Dependences of the model warming of the Atlantic layer in 10 years on α and t_r (the warming is given in degrees Celsius).

A radioisotope analysis of water samples taken from different parts of the Arctic Basin shows that the 'age' of the Atlantic layer waters does not exceed 20–30 years [9] (the 'age' is understood to be the period from the time when these waters have been last in contact with the atmosphere before plunging into the deep layers). If we assume that $w = 4 \times 10^6$ m^3 s^{-1}, we find that $t_r = 23$ years, which is in agreement with the results of radioisotopic analysis. The corresponding temperature rise curve for the Atlantic layer, obtained for the values of the other parameters given above (Fig. 3), shows that temperature should rise by 0.04 °C in ten years. If we bear in mind the enhancement of the greenhouse warming effects in the Arctic, predicted by the models, compared with the other regions of the Earth, we can assume also a greater rate of heating of the West Spitsbergen Current (α). For example, if $\alpha = +0.05$ °C year^{-1} and t_r amounts to 1 year, the increase in the average temperature of the Atlantic layer in 10 years reaches 0.1 °C (Fig. 4).

Such warming can definitely be detected acoustically during the propagation of a signal along extended paths, but the main problem is whether this consequence of the greenhouse effect is detectable against the background of natural fluctuations from year to year that occur in the Arctic. The asterisks in Fig. 3 give the mean (averaged over all the paths) temperature of the Atlantic layer for each year from 1973 to 1979, obtained in the POLEKS programme [7]. There are not only strong fluctuations from one year to another, but also evidence of a positive trend, but this trend can hardly be related to the greenhouse effect, because during a different time interval from 1955 to 1975 a strong fall of the temperature of the Atlantic waters in the Arctic Basin has been noted: −0.35 °C in the Atlantic area and 0.14 °C in the Pacific area [5].

The strong natural variations of the oceanographic characteristics of the Arctic Ocean represent the main obstacle facing the chief task of the Arctic ATOC programme, which is identification of a possible 'greenhouse signal' against the background of the existing 'noise'. Considerable inventiveness will be needed both at the stage of planning the experiments and during analysis of the results, if success is to be achieved.

However, success may come from completely different quarters. Acoustic monitoring of the Arctic Basin not only provides extremely important information on changes in its heat content, which are important irrespective of whether the greenhouse effect exists or not, but it also makes it possible to follow changes in the salinity of its upper layer. This may be even more important than thermometry, because variations of the salinity of the upper layer of oceans have a fundamental influence on heat and mass transfer between the surface and deep layers [10]. In the light of this, one should call the Arctic ATOC programme the programme of acoustic monitoring of the ocean climate of the Arctic, i.e. Arctic AMOC.

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Theoretical and experimental investigations of the feasibility of acoustic thermometry of climatic changes in the Arctic Ocean

A N Gavrilov, M M Slavinskii, A Yu Shmelev

1. Introduction

Changes in the Earth's climate are attracting growing interest irrespective of whether they are anthropogenic or are due to natural processes. Atmospheric climatological observations are greatly complicated by strong synoptic, seasonal, and year-to-year variations of the temperature of air, as well as by thermal pollution of large cities and industrial objects. It is postulated that possible trends in climatic changes, including those associated with the greenhouse effect, should be reflected in changes of the temperature distribution of seawater. This hypothesis led Munk and Forbes [1] to the proposal that observations should be made of changes in the average temperature of seawater by measuring the travel time of acoustic signals over very long transoceanic paths. The proposal is based on the existence of an approximately linear relationship between the temperature of seawater and the velocity of sound (the constant of proportionality is approximately $5 \text{ m s}^{-1} \, ^{\circ}\text{C}^{-1}$). According to a scientific programme proposed by Mikhalevsky et al. [2], a study should be made of the feasibility of long-term acoustic monitoring of the temperature of water in the Arctic Basin and of the average thickness of the ice cover, which is an equally important indicator of the state of the climate in the Northern Hemisphere. Scientific studies forming part of the programme known as Arctic ATOC (Arctic Acoustic Thermometry of Ocean Climate) began in 1993 with the participations of scientists from the USA, Russian, Canada, and Norway.

The theoretical part of the programme includes simulation of the propagation of lf acoustic signals in the Arctic waveguide and an analysis of the reaction of the amplitude – time characteristics of signals to large-scale changes in the temperature of water and in the parameters of the ice cover, including possible climatic trends. With this in mind, a model of the reflection and scattering of sound by an uneven elastic ice cover and an algorithm for the calculation of acoustic fields of lf sources in the Arctic waveguide have been developed. The fullest description of this model and algorithm can be found in Ref. [3]. It is also shown in Ref. [3] that the optimal frequencies for signal transmission over trans-Arctic paths lie in the range 15 – 25 Hz. At frequencies below 15 Hz the absorption of sound by the sea bottom becomes significant. At higher frequencies the scattering of sound by the uneven ice cover leads to a strong attenuation of acoustic signals and to considerable energy losses, which are unacceptable for the trans-Arctic signal transmission. The results of such simulation were checked and the scientific and technical viability of the proposed method of observations of the climate of the Arctic Ocean was verified by the first experiment on trans-Arctic transmission of lf acoustic signals carried out in the spring of 1994.

2. Simulation of the reaction of an acoustic signal to large-scale variations in the seawater and ice in the Arctic ocean

Both tonal and pulsed signals are used in acoustic monitoring of media. The measured parameters of a tonal signal are its amplitude and phase and in the case of a pulsed signal, they are the amplitude and travel time. In view of the limited power of a source, the signal/noise ratio can be improved if pulsed signals are replaced with broadband frequency-modulated or phase-modulated signals which have a transient characteristic and, consequently, permit time resolution of the corresponding pulsed signal. In acoustic tomography it is frequently found that signals are phase-modulated by pseudorandom sequences of 'maximum' length, which are known as M sequences [4].

Changes in the distribution of the temperature of water in an oceanic waveguide influence the travel time and the phase of the signals, alter the interference structure of the field, but have no effect on the energy characteristics of signals such as the amplitudes of the waveguide modes. In simulation of the influence of the variations of the waveguide on the signal we shall consider the response of individual acoustic modes. Monitoring of the waveguide medium on the basis of individual modes or rays, on condition that they can be separated in the space or time domain at the receiving end of a path, can provide additional and very important information on the vertical distribution of changes in the waveguide. Mesoscale (tides, vortices) and large-scale (seasonal, year-to-year) variations of seawater form a natural background which hinders identification of climatic trends. In the Arctic Ocean the scale of such variations associated with internal waves, tides, and synoptic vortices is considerably less than in the middle lattitudes. The influence of the ice cover, which insulates the Arctic Ocean from the atmosphere, is the reason why seasonal fluctuations of the temperature of seawater are also small and occur only in a narrow (amounting to several tens of metres) surface layer [5]. The slower year-to-year fluctuations of the temperature of seawater in the Arctic Ocean are associated mainly with the variations in the temperature and volume of the Atlantic waters entering this region. The latter fluctuations are the most important. Information on the year-to-year variations in the temperature of the Arctic Ocean along the proposed acoustic monitoring path has been deduced from the results of seven-year hydrological observations made by the 'Sever' expedition carried out by the Arctic and Antarctic Scientific-Research Institute every spring from 1973 to 1979. This expedition involved measurements of the vertical distribution of the temperature and salinity of water over most of the Arctic Basin. Fig. 1a shows the year-to-year variations of the mean temperature of water over the Spitsbergen – Barrow Point path, relative to the 1974 temperature. The measurements were carried out in the following layers: (*1*) the upper cold layer where mixing takes place; (*2*) the layer of a positive salinity jump; (*3*) the layer with a strong positive temperature gradient; (*4*) the core of the Atlantic waters; (*5*) the deep-water part of the Atlantic layer. It is evident from this figure that the year-to-

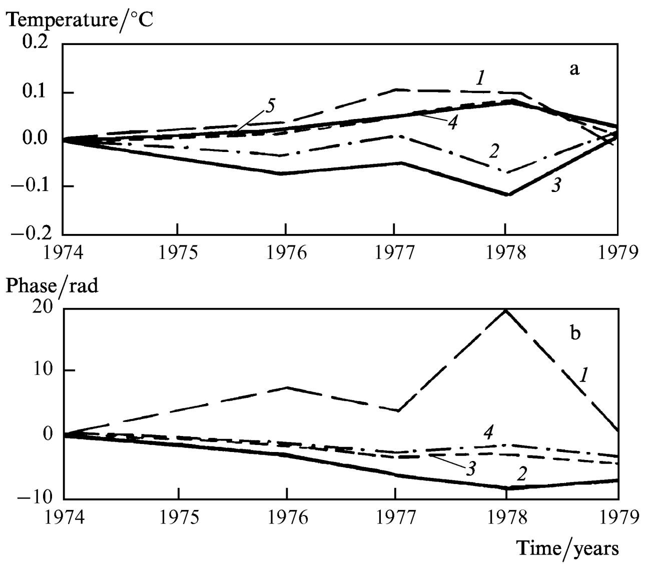

Figure 1. (a) Relative year-to-year variations of the average temperature of the Arctic Ocean water along the Spitsbergen – Barrow Point path in the following layers (*1*) 0 – 50 m; (*2*) 50 – 150 m; (*3*) 150 – 300 m; (*4*) 300 – 500 m; (*5*) 500 – 1000 m. (b) Relative changes in the phase of modes *1* – *4* of a 20 Hz signal over a path 2200 km long, caused by year-to-year fluctuations of the temperature and salinity of water and by the warming of the Atlantic layer at the rate of 0.01 °C per year.

year fluctuations of the average temperature of water in some layers can reach 0.1 °C. Moreover, these changes in temperature can have opposite signs in the higher and lower layers of seawater.

The acoustic response to these climatic changes was simulated by adding to these year-to-year temperature fluctuations a hypothetical positive trend of 0.01 °C per year in a column of the Atlantic water layer (300 – 500 m). This trend was taken from an estimate [6] of the reaction of the temperature of this layer to the possible greenhouse warming in the atmosphere of the North Atlantic. Fig. 1b shows the calculated relative changes in the phase of the first four acoustic modes of a signal of 20 Hz frequency over a 2200 km section of the Spitsbergen – Barrow path in the presence of this hypothetical temperature trend in the Atlantic water layer. A comparison of Figs 1a and 1b shows that the variations of the phases of the individual modes are strongly correlated (but with the opposite sign) with the variations in the temperature in specific water layers. The first mode 'senses' the changes in the temperature of the upper layers, whereas the second mode reacts mainly to variations in the Atlantic waters layer and is not affected by the influence of the strong year-to-year fluctuations of the temperature of the upper layers. The positive trend of the temperature of the Atlantic layer is revealed quite clearly by changes of the phase of the second mode against the background of the natural year-to-year fluctuations. The higher modes are much less affected by variations in the Atlantic layer. The sensitivity of the phase and the travel time of the modes to the temperatures of the individual layers depends on the signal frequency. For example, at 30 Hz the changes in the Atlantic layer have the greatest influence on the travel time of the third mode.

Let us now consider the variations in the ice cover of the Arctic Ocean on the characteristics of acoustic signals propagating in the Arctic waveguide. In contrast to the water layer, the ice cover in the Arctic is subject to large seasonal variations. There are not only changes in the total area of the ice cover, but also in the average thickness and height of the irregularities of the ice cover. These parameters have the strongest influence on the propagation of sound under ice. Moreover, the acoustic properties of sea ice are governed by the velocities of the longitudinal and shear waves, which are also characterised by a seasonal dependence. The phase of the ice-reflected coherent (not scattered) component of the acoustic signal depends on the average thickness of the ice plate. However, at low acoustic frequencies the changes in the ice thickness have little influence on the phase and group velocities of acoustic modes, because: first, at low frequencies an ice cover 3 – 4 m thick becomes acoustically almost transparent; second, changes in the thickness of ice immersed in water to the extent of 90% alter only slightly the position of the upper ice – air reflecting boundary of the waveguide. Fig. 2 shows the dependences of the phases of the first three modes on the thickness of ice over a path 2200 km long. A comparison of the scales of changes in the phase in Figs 1b and 2 shows that the maximum possible changes in the thickness of ice have an influence on the signal travel time which is two orders of magnitude weaker than the effects of natural fluctuations of the temperature of water.

The thickness and height of the irregularities of the ice cover affect significantly the attenuation of acoustic signals. Therefore, measurements of the energy characteristics of

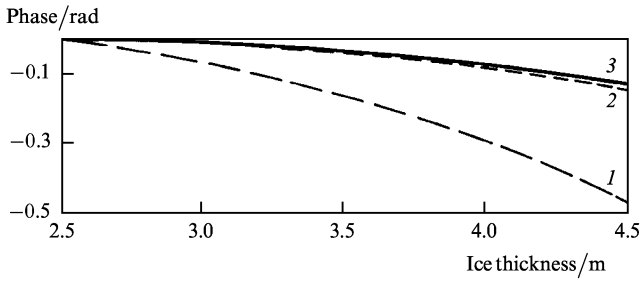

Figure 2. Relative changes in the phase of modes *1 – 3* of a signal of 20 Hz frequency over a path 2200 km long, plotted as a function of the average thickness of the ice cover.

such signals can provide additional information on variations of the Arctic ice cover and on possible climatic trends. We simulated the process of acoustic monitoring of the ice cover on the basis of the data on the seasonal variations of the ice thickness and of the height of the irregularities [7], and of the physical properties (density of ice, velocity of sound, etc.) [8]. A hypothetical climatic reduction in the thickness and height of the irregularities of ice, postulated to be due to global warming (−10 cm and −6 cm per year, respectively), was added to the seasonal variations.

Fig. 3a shows a model of variations of the characteristics of ice over a period of 10 years. The calculated changes in the integral attenuation of the first three modes over a path 2200 km long deduced from this model are plotted in Fig. 3b. It is clear from these results that the seasonal variations of the ice cover result in strong fluctuations of the signal level, particularly of the first mode which interacts most strongly with ice. However, against the background of these fluctuations in a ten-year series of acoustic data we can identify the climatic trend reducing the thickness of ice at the rate of 10 cm per year. The adopted model ignores the year-to-year variations of the parameters of ice because there are at present insuffi-

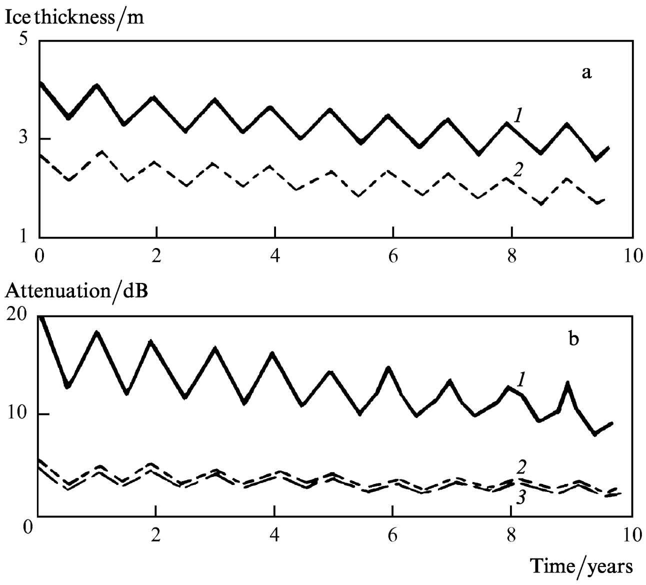

Figure 3. (a) Model of the temporal variations of the average thickness of the Arctic ice cover (*1*) and of the rms height of the irregularities of this cover (2) with a superimposed climatic trend reducing the ice thickness at −10 cm per year. (b) Changes in the integral attenuation of modes *1 – 3* of a 20 Hz signal over a path 2200 km long.

cient data for simulation of such variations. In the presence of strong year-to-year variations the period of time needed to identify stable climatic trends could be considerably longer than 10 years.

3. Experiments on trans-Arctic transmission of acoustic signals

The Russian – American cooperative experiments on the transmission of acoustic signals across the whole Arctic Ocean, carried out in the spring of 1994, had the following main aims: (1) determination of the attenuation of lf acoustic signals over trans-Arctic paths; (2) determination of the stability of the amplitude and time characteristics of signals over different time periods; (3) investigation of the mode structure of signals and variations of this structure with time; (4) gathering of experimental data for the verification of the proposed theoretical models; (5) testing of technical means for the emission and reception of signals, and of algorithms for the processing of these signals.

Over a period of 6 days an lf (19.6 Hz) signal was transmitted from the Russian Scientific Stations 'Turpan', drifting 300 km to the north of Spitsbergen. This signal was received by American drifting 'SIMI' and 'Narwhal' Stations in the Beaufort and Lincoln Seas, respectively (Fig. 4). The transmitted signal was of two types: tonal and phase-modulated (by $\pm\pi/4$) periodic M sequences of different lengths (M-127, 255, 511, and 1023). Signals lasting 1 h were emitted periodically over the 3 hours. The length of each bit of the modulating M sequence was 12.5 periods of the 19.6 Hz carrier. The carrier was synchronised with the aid of a highly stable rubidium frequency standard accurate to within 1×10^{-11}. The source of the signal was at a depth of 60 m. The acoustic power of the emitted signal (195 dB, relative to 1 μPa at

Figure 4. Acoustic paths in the reported experiments.

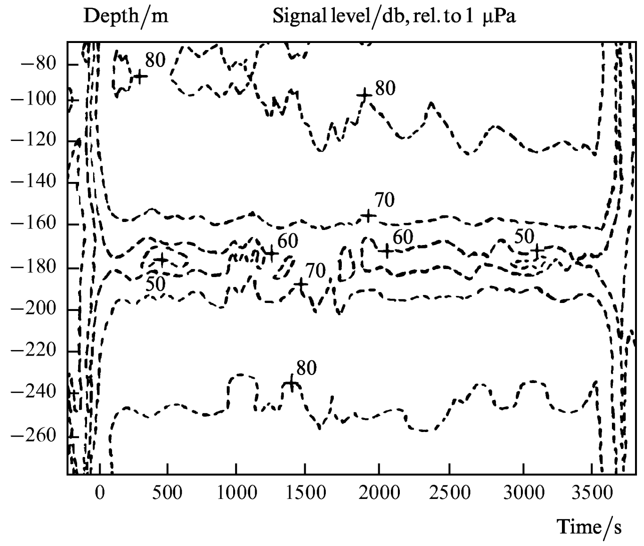

Figure 5. Time dependence of the vertical distribution of the acoustic field level obtained for the tonal signal over a period of 1 h.

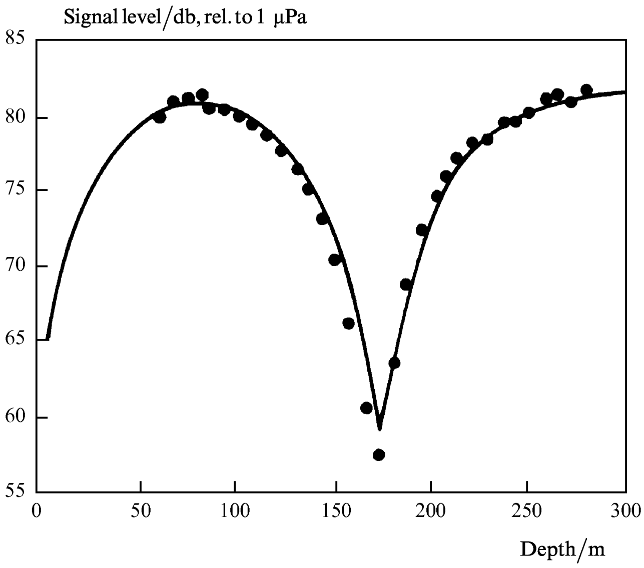

Figure 6. Distribution of the tonal signal levels (averaged over 1 h) recorded by the vertical antenna hydrophones () at a distance of 2635 km and the calculated vertical distribution of the acoustic field level (continuous curve).

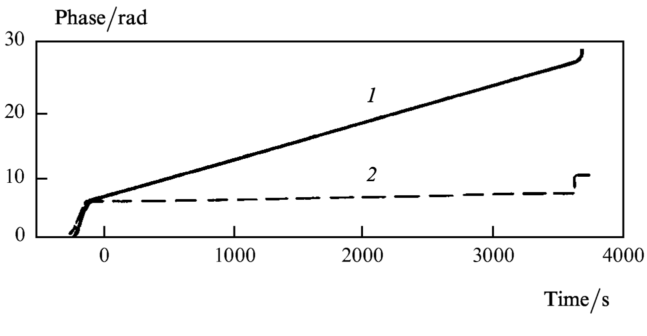

Figure 7. Changes in the phase of the tonal signal during one run before (*1*) and after (*2*) corrections for the Doppler frequency shift based on the GPS navigational satellite data.

1 m) was monitored with a special hydrophone located at 100 m from the source. At the 'SIMI' Station the signals were received by a vertical 32-component equidistant antenna 217 m long, placed at depths from 62 to 279 m, and by a 32-component horizontal antenna with a complex configuration and maximum aperture 480 m, located at a depth of 60 m. The vertical profiles of the temperature, salinity, and velocity of sound in water were determined during these experiments at both transmitting and receiving stations. The exact coordinates of these stations were measured continuously and recorded with the aid of the GPS satellite navigation system.

We shall now consider the main results of the processing of signals received at the 'SIMI' Station at a distance of over 2600 km from the source. Fig. 5 demonstrates the stability of the interference structure of the acoustic field of the tonal signal received by the vertical antenna during transmission lasting 1 h. The distribution of the absolute levels of the signal, received by the vertical antenna hydrophones and averaged over 1 h, is shown in Fig. 6. Fig. 6 includes also the dependence calculated for a signal of the 195 dB level and 19.6 Hz frequency at a distance of 2635 km from the source, corresponding to the positions of the drifting stations at the time of measurements. The calculated curve is in good agreement with the experimental results, both in respect of the vertical distribution of the acoustic pressure and in respect of the absolute level of the signal. This confirms the validity of the proposed model for the propagation of sound in the Arctic waveguide, including the algorithm used to calculate the attenuation of the acoustic modes. The phase of the received tonal signals varied approximately linearly during one transmission run. The rapid variations of the phase were almost completely compensated by correction for the Doppler frequency shift resulting from the relative drift of the transmitting and receiving stations (Fig. 7). The long-term variations of the single parameters such as those associated with the influence of inertial and tidal waves or other mesoscale phenomena in seawater, will be analysed in future. Processing of the phase-modulated signals involved demodulation, correction for the Doppler shift, and calculation of a convolution with the modulating M sequence. This processing yielded the pulsed transfer function of the waveguide in the frequency band $1/T$ where $T = 0.64$ s is the duration of one bit of an M sequence. The signal/noise ratio was improved by phase compensation of the horizontal antenna, which made it possible to ensure that the antenna gain was about 10 dB. Moreover, for a period of 1 h the signal stability was sufficient for effective coherent averaging over time intervals equal to the period of the corresponding M sequences. In this way, the total signal gain after processing reached almost 40 dB, which made it possible to increase the signal/noise ratio to 30 dB or more. This processing of the phase-modulated signals produced by the pulsed response of the waveguide made it possible to identify the peaks corresponding to the arrivals of the individual modes, including the very weak first mode, as well as to determine the model amplitudes and measure the absolute travel times. Fig. 8a shows the shape of the envelope (amplitude) of an M-127 signal received from a source located at a distance of 2640.5 km and processed on absolute scales of time and acoustic pressure. The pulses of modes *1*–*4*, which arrived last and in reverse order, were clearly distinguishable in the signal. Higher-order modes,

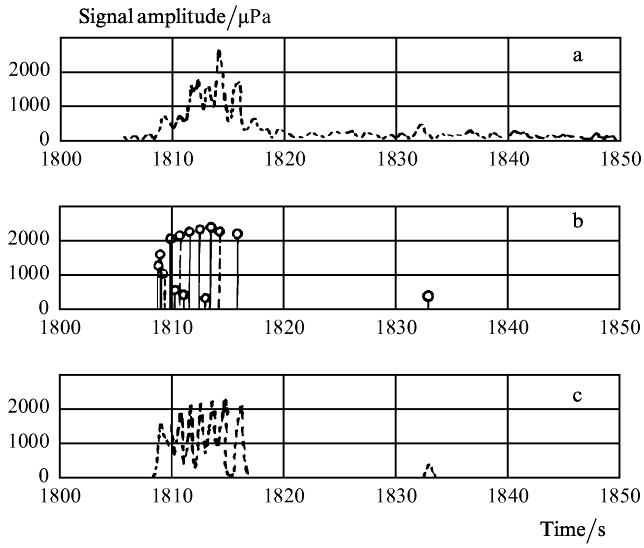

Figure 8. (a) Enveloped of an M-127 signal received by the horizontal antenna and processed; distance 2640.5 km. (b) Calculated distribution of the travel times and amplitudes of the modes of a 19.6 Hz, 195 dB signal at a distance of 2640.5 km from the source, at a depth of 60 m, deduced from the velocity of sound along the path in 1977. (c) Calculated envelope of a pulsed signal of 0.64 s duration.

deep-water modes, and those interacting weakly with the bottom had higher and similar group velocites, and they arrived first in the form of a group spread out in time.

A comparison of the experimental profile of a pulsed signal with the profile calculated numerically on the basis of the model of propagation of sound in the Arctic waveguide was of special interest. The calculations were carried out relying on the hydrological data obtained in the course of the 'Sever' expedition along the Spitsbergen – Barrow Point path, which was almost the same as the 'Turpan – SIMI' path in these experiments, and on the statistical characteristics of the ice cover taken from Ref. [9]. Fig. 8b is a diagram showing the calculated amplitudes and arrival times of the first 15 modes at a depth of 60 m and at 2640.5 km from the source of a pulsed signal of 195 dB level. Fig. 8c gives the corresponding profile of the envelope of a pulsed signal when the duration of the emitted pulses was 0.64 s. A comparison of Figs 8a and 8b showed that, at the distance just given, the agreement between the theoretical and experimental values of the amplitude and travel time of the acoustic modes was very good. The greatest difference between the theoretical and experimental signal profiles was observed in the group of deep-water modes. However, in this case we could not expect a good agreement for a number of reasons. First, in this group the signal modes in the 1.5 Hz band interfered strongly with one another and the pulse profile was very sensitive to the phases of the individual modes, which in their turn depended strongly on the distribution of the sound velocity along the path. Obviously, the real and the model distributions were different. Second, this group included modes which interacted weakly with the bottom of the sea. In view of the absence of sufficiently reliable data on the relief and acoustic properties of the bottom along the path, particularly in the regions of the underwater Lomonosov and Alpha ridges, the results of simulation of this group of modes should be treated with caution. Third, the adopted model ignored the mode transformation effects over sections of the waveguide in the region of relatively steep slopes of the Lomonosov ridge, which could also give rise to the discrepancies which were observed.

The characteristics of the first three – four modes were much more informative from the point of view of acoustic monitoring of the climate in the Arctic Ocean than the parameters of the deep-water modes. It is worth noting a small difference between the travel times of modes *1 – 3*, determined during the experiments carried out in the spring of 1994 and those calculated theoretically on the basis of the oceanographic data for the same season of the year in the seventies. The travel times of these modes, particularly of the first one, calculated on the basis of the data obtained during various years between 1973 and 1979 were considerably longer (by 0.5 – 2 s) than the experimental estimates. The systematic discrepancy could be due to, for example, warming of the upper and Atlantic waters in the Arctic Ocean layers compared with the data obtained in the seventies. The reason for this discrepancy will be investigated in future.

4. Conclusions

Theoretical and experimental investigations showed that trends in the changes in the temperature of the water layers in the Arctic Ocean, significant from the point of view of global climate, can be detected by acoustic monitoring over trans-Arctic paths when measurements are carried out over a period of 10 years. Acoustic monitoring of the ice cover of the Arctic may also help in revealing stable tendencies in climatic changes, but in this case very probably a long monitoring period will be required.

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