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ELECTROMAGNETIC VARIATION ANOMALIES IN THE ARCTIC OCEAN (ANOMALII
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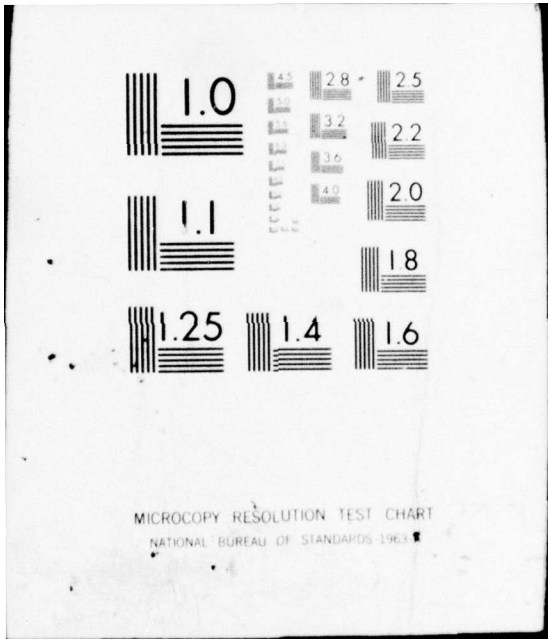
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ELECTROMAGNETIC VARIATION ANOMALIES IN THE ARCTIC OCEAN

[Shneyer V. S.; Arkticheskiy i antarkticheskiy nauchno-issledovatel'skiy institut; Trudy, vol. 310, 1972, pp. 100-113; Russian]

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It is known that, unlike the permanent terrestrial magnetic field, the variable field is distributed rather uniformly in space. This is brought about by the external distant location of the variable field sources and their relatively uniform distribution above the Earth. More or less significant gradients of the variation field should be observed only near the polar and equatorial electro-jets. This is what was thought until quite recently when the net of magnetic observatories was still quite thin. However a number of cases have become known lately where the variations differ substantially at closely located points.

These differences embrace a wide frequency spectrum ranging from micropulsations and "bays" to $S_{\text{sub } q}$ and $D_{\text{sub } st}$ variations. Such anomalies have been surveyed by T. Rikitaki /9/, Schmukcer /22, 23/, Witham /24/, and Wiese /25/, for Japan, Canada, Western Europe, and the western United States, as well as for East Europe and North Canada. In the Soviet Union a number of anomalies in Siberia and the Near East ^{have been} discovered and interpreted by L. L. Van'yan and B. Ye. Marderfel'd /4/. ~~We studied the~~ studies of anomalies in the Arctic Ocean, /16/. ^{This report covers}

↑
The magnetic disturbance field can be presented in the form of the sum:

$$\delta H_{x,y} = \delta H_{en} + \delta H_{in} + \delta H_e + \delta H_i, \quad (1)$$

where δH_{en} is the normal external part constant in the region of investigation;

δH_{in} is the normal internal part (induced in the spherical layered horizontally homogeneous Earth);

δH_e is the external anomalous part;

δH_i is the internal anomalous part.

The object of our investigation is field δH_1 which contributes the most substantial local variation anomalies. In addition, the change in variations in the sea along the vertical (in depth) $\delta H(z)$, which also creates "anomalous" variations at depths as compared with "normal" variations on the Earth's surface, is examined. The law of electromagnetic induction can be written in the form of a generalized second Maxwell equation:

$$\text{rot } \vec{E} = \frac{\partial \vec{B}}{\partial t} + \text{rot } [\vec{V} \cdot \vec{B}]. \quad (2)$$

It follows from this equation that the induced field in the general case is caused by two kinds of induction -- by a change in magnetic field \vec{H} in the time and by the "Lorentz force" that generates an electrical field because of charge division when the medium moves at velocity \vec{V} at a certain angle to vector \vec{B} . Therefore we shall examine not only the usual "induced anomalies", but also anomalies produced by marine hydrodynamic sources. The investigations were made for variations with periods from minutes to days.

To investigate the magnetic variations on the surface of the Earth, both data from polar magnetic observatories equipped with stationary Eschenhagen-Tepfer magnetographs as well as data obtained in field magnetic variation stations of the Arctic and Antarctic Institute (MVS AANII) were used.

Field magnetic variation observations were made by expeditions of the AANII-IZMIRAN and the Institute of Arctic Geology (NIIGA) from 1956 through 1967 (with interruptions) on Franz Josef Land.

In 1964, pursuant to a special AANII program, observations were made of variations from the ice floe station "Severnny polyus 13." To record magnetic variations under water the AANII designed several original instruments /13/: an underwater Z-variometer, an underwater proton magnetometer, and a bottom magnetic variation station. The first two instruments were used in operations from an ice flow in the Central Arctic Basin, the bottom MVS was used from landfast ice in bay waters of the Franz Josef Land archipelago. Since there was not enough equipment and transport facilities to carry out synchronous observations during operations on the Franz Josef Land islands, relative variation values were derived.

To do this, the variations in each field point were standardized with respect to the base point -- the magnetic station on Hayes Island. The duration of variation recordings ranged from several days to two months. As a rule, variations of the D, H, and Z components were registered; the time scan rate of the operating magnetograms was 20 mm/hr,

while scale division was of the order of 5-10 ν/mm . In the underwater investigations the sensitivity of the variometers was brought to a scale division of 1-5 ν/mm . When the undersea instruments were submerged they were tracked so that they were in about a vertical with the MVS at the surface.

Besides magnetic field variations, variations of marine telluric currents and marine currents were registered. The telluric currents were recorded with the IZMIRAN 3-component unit from the ice of Austrian Channel (Franz Josef Land), while the water currents were recorded with the electric current recorder near where the electromagnetic variations were being registered. In several points magnetic variation observations were made. The longest series of observations were made in the Tikhaya Bay and Hayes Island magnetic observatories. When simultaneous magnetograms from these observatories were compared, a large amplitude difference was revealed, especially in the Z variations. Fig. 1 shows part of the simultaneous record on Hayes Island and Tikhaya Bay. The variations on Hayes Island have amplitudes several times less than in Tikhaya Bay.

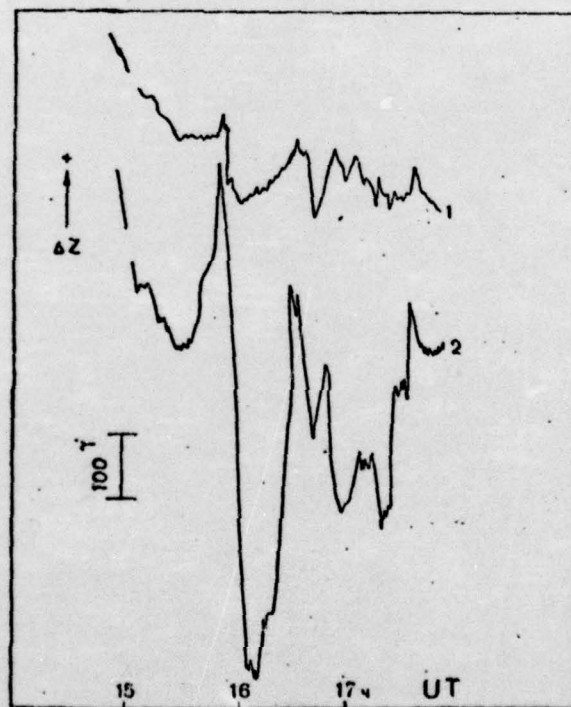


Fig. 1. Nature of the magnetic variations on Hayes Island (1) and in Tikhaya Bay (2).

Observations made in seven field stations made it possible to obtain a broad profile of normed variation amplitudes running through a large part of the archipelago (Fig. 2). Plotted along the axis of the ordinates are the relative amplitudes $\frac{\delta D_0}{\delta D_0}, \frac{\delta H}{\delta H_0}, \frac{\delta Z}{\delta Z_0}$, where index "0" designates the amplitude on Hayes Island. The average values of these relationships are derived from 20-40 elementary variations with periods from 5 to 100 minutes.

As may be seen from the diagram, there is a sharp inhomogeneity in the Z-variation distribution. In the sector Graham Bell-Island-Hayes Island-Newcomb Island, δZ remains almost constant, in sector Newcomb Island-Tikhaya Bay within a distance of 60 km the δZ amplitudes increase by almost five times. Further between Tikhaya Bay and Nagurskaya δZ remains almost unchanged. Variations of the horizontal components δD and δH change considerably less, chiefly in the sector Tikhaya Bay-Nagurskaya where they are almost halved.

These peculiarities could be caused both by external as well as internal factors. Although in this case external causes could have little effect because of the following circumstances: the field of the polar electro-jet in the region of the profile is quite homogeneous due to the considerable distance (800 km) from it and the latitudinal direction of the profile; the amplitude ratio does not change for a 24-hour period.

Having assumed that internal causes are the main element in the observed distribution, namely, the presence of horizontal inhomogeneities of electrical conductivity, we attempted to determine their position. Assuming that the Z-variations attenuate in proportion to the cube of the distance from the linear source, it is possible to reckon the primary δZ to be very small and the main source of Z-variations to be currents in the Earth.

In this case the inhomogeneities are easily explained, since the electrical conductivity of the underlying rocks is not uniformly distributed in the archipelago. It is known from the theory of plane electromagnetic waves that above the horizontal homogeneous conducting semispace, the vertical component of the magnetic field is equal to zero. This conclusion stems from the second Maxwellian equation:

$$\delta Z = \frac{1}{i\omega\mu_0} \left(\frac{\partial E_x}{\partial y} - \frac{\partial E_y}{\partial x} \right). \quad (3)$$

Consequently, $\delta Z \neq 0$ only in the case of the inequality of derivatives of the electrical field E with respect to coordinates x and y. The excess current flowing along the linear formation with high conductivity will also create a certain anomaly in the variations of the horizontal component, perpendicular to the current direction. With the Biot-Savart law it is possible to determine the excess current field:

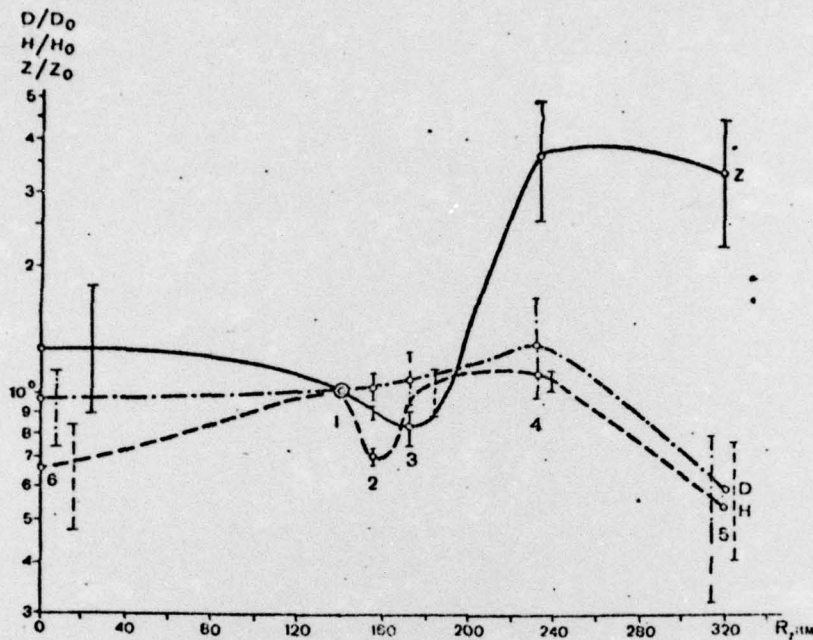


Fig. 2. Distribution of amplitudes of magnetic variations of the Z, D, and H-components along the profile. 1 - Hayes Island; 2 - Bliznetsa Cape; 3 - Newcomb Island; 4 - Tikhaya Bay; 5 - Nagurskaya Station; 6 - Graham Bell Island. Vertical sections are mean square errors.

$$\delta H_x = \frac{2\pi I}{\Delta y} \quad (4)$$

where I is the current intensity between the inhomogeneity boundaries;
 Δy is the inhomogeneity width.

If the ratio of anomalous variations $\frac{\delta Z}{\delta H}$ in the far wave zone is known, where the boundary conditions of plane waves should already be satisfied, then it is possible to find the location of the geoelectrical inhomogeneities by constructing vectors of the internal field. Such vector constructions have been made by several investigators who have proposed some modifications in the method of construction. Methods are known that were developed by V. L. Parkinson /20/, U. Schmucker /22/, M. N. Berdichevskiy /1/, and H. Wiese /25/. We have used the last named method which consists in finding component δH_x in expression (4) as a function of δZ . Qualitatively, δH_x variations will most resemble δZ variations from all the other horizontal components.

After having designated as θ the angle between the magnetic parallel and δH_x , we write the equation:

$$\delta Z = C(\delta H \cdot \sin \theta + \delta D \cdot \cos \theta), \quad (5)$$

where C is the relative value of the modulus of the vector $\left| \frac{\delta Z}{\delta H_x} \right|$ of the internal field.

Thus, having determined the value and direction of the internal field we calculate the relative current intensity and direction along the inhomogeneity. Equation (5) is either solved graphically or using the method of least squares. We calculated and constructed Wiese vectors for bay-shaped disturbances having periods of $100 > T > 30$ minutes and for elementary disturbances having periods of $30 > T > 5$ minutes. Reference /17/ shows Wiese vectors for bay-shaped and elementary disturbances. The vector arrows are directed normal to the current, in the direction of current increase. It is certain that the current is mostly localized to the east of the British Channel along the center of the archipelago and is directed at an angle of about 40° to the meridian. On the whole, this current position agrees with the structure of the Earth's crust in this region.

There is a broad syncline here that is filled with thick sedimentary deposits having low specific resistance ($\rho'_k \approx 5 \text{ Ohm}\cdot\text{m}$), when there is a high intruding medium resistance $\rho'_k \approx 100 \text{ Ohm}\cdot\text{m}$ /5/. The syncline is the geoelectric inhomogeneity that causes the variation anomalies observed.

The variation anomaly in the Nagurskaya Station is associated with the shore effect on the continental slope and is apparently determined by the excess current in the good conducting mantle beneath the oceanic crust. It must be noted that the anomaly in Z variations manifests itself not only in an increase, but also in an amplitude decrease as compared with the horizontal homogeneous geoelectric conditions. This phenomenon is observed particularly on Hayes Island, where δZ amplitudes are significantly less than at other points on the same geomagnetic latitudes. This anomaly, on the one hand, is explained by the high electrical conductivity of the Earth's crust beneath the island and, on the other hand, by the location of the island in the center of the syncline, distant from the boundary effects of the inhomogeneity.

Thus, when primary Z variations occur, they induce in the Earth a horizontal current system, the secondary field of which compensates the primary, according to Lenz law. Analogous anomalies are observed at the surface of deep-water ocean regions.

Such anomalies were first discovered by L. N. Zhigalov from ice floe station data /6/. He established the close inverse correlation of the ratios of hourly amplitudes r_z to r_H with ocean depth. When the ocean

depth exceeds 2 km, a large part of the Z-variations at the water surface practically disappears. Similarly, close analogies are found between the behavior of variations at the boundaries of any geoelectric inhomogeneities on dryland, in the sea, and along coasts that expresses itself in an increase in the variation amplitudes vertical and normal to the boundary of the horizontal component /3/. It is interesting to note that the most widespread induced anomaly -- the shore effect -- is almost absent in the case of Hayes Island. This can apparently be attributed to the small difference in the conductivity of the seawater and the coastal formations, low-resistance deposits.

It is of interest to examine some special features of the daily variations of the magnetic field on Franz Josef Land. We processed magnetic variation data on Hayes Island for several magnetically quiet years using the familiar Chapman-Bartels method. The method was proposed to separate the lunar daily variations and is based on difference in the lengths of solar and lunar days. As a result of the processing, a semi-daily wave was identified in all three components and a hodograph constructed for the total horizontal vector.

After a qualitative comparison of the elements of the variation vector hodograph with that of the velocity of the water current in the Austrian Channel near Hayes Island it was suggested that the source of this variation is the current induced by the tidal current in the sea /14/.

The order of the magnetic field value of the current can be determined from magnetic hydrodynamic relationships /7/. The degree of magnetic hydrodynamic interaction of the conducting liquid with the magnetic field is characterized by the Reynolds magnetic number (Re_m):

$$Re_m = \mu\sigma Vd, \quad (6)$$

where μ is the magnetic permeability;

σ is the specific electrical conductivity;

V is the flow velocity;

d is the characteristic flow size.

Under our conditions $\mu = \mu_0 = 4\pi \cdot 10^{-7}$, $\sigma = 4 \text{ mho}\cdot\text{m}^{-1}$, $V = 0.24 \text{ m/s}$, $d = 2 \cdot 10^2 \text{ m}$ is the flow depth:

$$Re_m = 2.4 \cdot 10^{-4}.$$

When the values are so low, the mechanical action can be disregarded. However in such a case Re_m yields an order of value of the ratio of the induced field to the primary /12/. Since in the vicinity of Hayes Island

$H_z = 6 \cdot 10^4 \gamma$, the induced horizontal field will be $h = 14\gamma$.

After the electrical fields of the water currents had been measured in the channel off Hayes Island, it became possible to calculate the magnetic field of the currents more rigidly. Let us assume a current system induced by a horizontal flow in the vertical magnetic field H_z , in the channel with conducting shores and bottom. Obviously it will have the form of a solenoid concentric with the channel axis. We shall calculate the magnetic field of this solenoid on its axis, in the center of the system. We shall use the coordinate system shown in Fig. 3. We shall assume that the current penetrates to depth $d' = \frac{D}{2}$. We shall designate the axial component of the magnetic field in the center of the flow h_y . For the calculation we use the expression for the field of a rectangular solenoid /10/:

$$h_y = \frac{WI}{\pi L} A_0 \left[1 + \frac{A_{zz} + A_{xx}}{3A_0} \frac{(d')^2}{L^2} + \frac{A_{zz}}{A_0} \frac{z^2 - y^2}{A_0} + \frac{A_{xx}}{A_0} \frac{x^2 - y^2}{L^2} \right], \quad (7)$$

where W is the number of solenoid "loops";

I is the total current;

L is half the length of the solenoid;

d' is the thickness of the "winding";

A_0, A_{zz}, A_{xx} are functions of the length L , width D , and depth d' of the current system.

Knowing that $L = 4 \cdot 10^4$ m; $D = 10^4$ m; $d' = 5 \cdot 10^3$ m, we find W as the area of a section of the system along axis y , and I as the modulus of current density $|\vec{j}|$ through an element of the section having an area of 1 m^2 . Since on the basis of measurements the amplitude of the lunar semi-daily harmonic M_2 of current density $\vec{j} = 6.8 \cdot 10^{-6} \text{ A/m}^2$, then $h_y = 16.5 \cdot 10^{-3} \text{ A/m} = 20\gamma$.

As has been shown in reference /14/, the value of the semi-daily variation identifiable with tidal induction on Hayes Island reaches 10γ . Although the calculation is quite approximate, experiment confirms it with respect to the order of value. The field could exit on the surface when there is a current density gradient. In our case the current density gradient is governed by the gradient of the water flow velocity component, its depths, and the electrical conductivity of the bed /12, 14/.

It is of interest to determine the vertical distribution of the mag-

netic field in the flow. To do this, we select a point far from the current boundaries where the velocity is independent of the horizontal coordinates and only change with depth. Let the velocity vector have only one component V_y , and the vector of the primary magnetic field component H_z . Then the current density vector will have component j_x directed to the left of the current vector in the northern hemisphere (Fig. 4). We shall find current density using Ohm's law for moving media

$$\vec{j} = \sigma\{[\vec{V} \cdot \vec{H}] + \vec{E}\}. \quad (8)$$

Now we shall write the expression for current density from the second Maxwell equation, disregarding displacement current.

$$\vec{j} = \text{rot } \vec{h}. \quad (9)$$

Describing rot. in rectangular coordinates, we obtain

$$\vec{j} = -\vec{i} \frac{\partial h_y}{\partial Z}, \quad (10)$$

where \vec{i} is the unit vector with respect to axis x.

Substituting expression (8) in formula (10), we write

$$\vec{i} \frac{\partial h_y}{\partial Z} = \sigma\{[\vec{V} \cdot \vec{H}] + \vec{E}\}. \quad (11)$$

Since $V = V(Z)$, then, as shown in reference /19/,

$$\vec{E} = \vec{E}(Z) = \frac{H_z}{a} \int_0^d V(Z) dZ,$$

then

$$\frac{\partial h_y}{\partial Z} = \sigma H_z \left\{ \frac{1}{a} \int_0^d V(Z) dZ - V(Z) \right\}. \quad (12)$$

The first term in equation (12) is the average current velocity \bar{V} throughout the entire depth. After integrating equation (12) from the surface to the depth Z, we obtain

$$h_y = \sigma H_z \mu_0 \left\{ \int_0^z \bar{V} dZ - \int_0^z V(Z) dZ \right\}. \quad (13)$$

Knowing the distribution function $V(Z)$, the distribution $h_y(Z)$ can be calculated by using formula (13).

Fig. 4 shows the distribution of the magnetic field of the current in depth in a sea covered with ice. For the conditions prevailing in the Austrian Channel, the computed amplitude value $h_y = 14\gamma$.

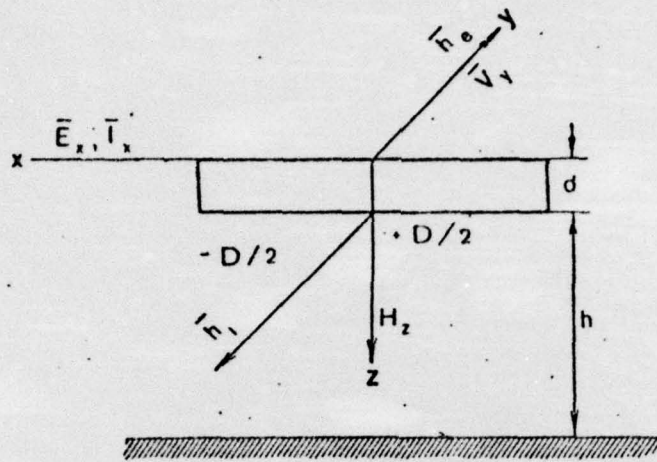


Fig. 3. Coordinate system for a rectangular cross section flow.
 D - flow width; d - flow depth.

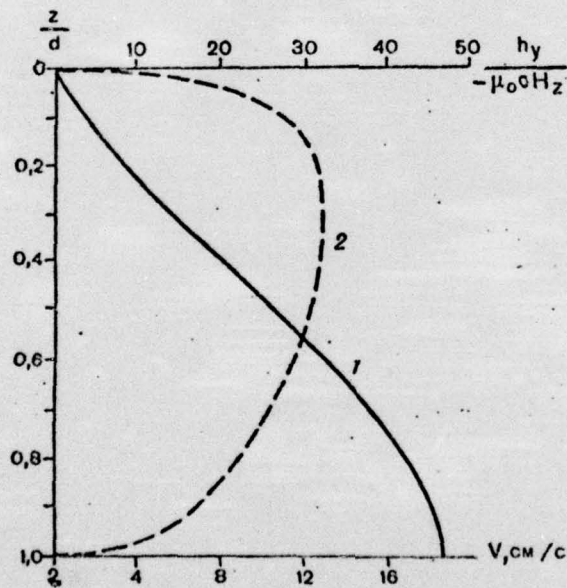


Fig. 4. Distribution of the magnetic field of the current with depth in the center of a wide current.
 1 - magnetic field h_y ; 2 - current velocity V .

Thus, for all models examined the value of the magnetic field of the current and the field delineated in the Hayes Island observatory from observations have one order. It must be noted, however, that for a precise calculation of the magnetic field of the current the three-dimensional distribution of current velocity, the true flow configuration, and the distribution of the electrical conductivity in a fixed "sector of an electrical circuit," which are the sea bottom, shore, and quiescent water layer, have to be taken into account. It is quite a complicated matter to take these factors into account, and presently the problem has only been partially solved, only for the field inside the current /18/.

Obviously, the tidal variation can occur wherever there are sufficiently great tidal current velocities. The maximum fields will be observed in the narrow and deep flows where there is a high electrical conductivity in the bed and strong primary fields. The value of the tidal variation is comparable with quiet daily variations. Consequently, it has to be taken into account in investigations of S_q and L variations.

It was noted above that by using underwater equipment it was possible to investigate magnetic variations in the seawater layer and on the sea floor. Synchronous observations of variations made in 1967 on the ice of the Austrian Channel and on its bottom made it possible to study the attenuation of variations (skin effect) in seawater. Theoretically the question of the skin effect in seawater was solved by A. T. Price and G. A. Fonarev /21, 11/. In reference /11/ it was shown that the attenuation of variations in the sea depends not only on the electrical conductivity of the seawater, but on the electrical conductivity of the bottom formations (σ_k) as well, if the skin layer is less than the depth of the sea. Since in the channels off Hayes Island the depth does not exceed 200-300 m, it follows from the expression for the depth of field penetration, that variations having a period of $T \geq 1$ s will penetrate to the bottom

$$T = d^2 \mu_0 \sigma \pi = 0.7c, \quad (14)$$

where T is the period of variations, and d is the depth of the sea.

Consequently, the attenuation of the entire spectrum of variations resolved by our equipment depends on the electrical conductivity of the bottom. From the magnetograms of surface and bottom magnetic variation stations, 60 synchronous quasisinusoidal elementary variations of δD and 30 variations of δH having periods of from 200 to 500 s were selected. Comparison of the amplitudes of these variations at the surface and on the sea floor showed very little attenuation /4/. Reference /16/ showed theoretical curves of the moduli of the relationships between variation amplitudes on the surface and those on the sea floor for different periods

at a sea depth of $h = 210$ m and specific water conductivity of $\sigma_M = 4$ Mho·m⁻¹ for $n = \sqrt{\delta_M \delta_K} = 1, 10, 100$. The curves were calculated using G. A. Fonarev's formula /11/ based on the relationships of the theory of plane electromagnetic waves for a three-layered geoelectric model

$$\frac{\delta H_0}{\delta I_h} = nsh \kappa h + ch \kappa h, \quad (15)$$

where the subscripts 0 and h refer to the surface and bottom of the sea, κ is the wave number. Each experimental value was obtained as an average from 5-6 elementary variations. As seen from the figure (reference /16/) the points are located only slightly higher than the curve corresponding to $n = 1$. Consequently, in our case σ_K is very great. The precise calculation using formula

$$\sigma_K = \left[39 \left(\left| \frac{\delta H_0}{\delta I_h} \right| - 1 \right)^2 T \right]^{-1} \quad (16)$$

yielded for the ratio $\left| \frac{\delta H_0}{\delta H_h} \right| = 1.19$ and period $T_{\text{aver.}} = 600$ s the value $\sigma_K = 0.12$ Mho·m⁻¹ and for $\left| \frac{\delta D_0}{\delta D_n} \right| = 1.09$ and $T_{\text{aver.}} = 900$ s, $\sigma_K = 0.86$ Mho·m⁻¹.

For periods $T_{\text{aver.}} = 240-400$ s the apparent electrical conductivity of the bottom formations almost coincides with the electrical conductivity of the seawater. Characteristically, the electrical conductivity calculated from both horizontal components δH and δD differ by half an order, while magneto-telluric sounding data, obtained from the same point, reveals a sharp anisotropy, and the corresponding components of the tensor of conductivity differ by almost two orders /17/.

As a result of the investigations conducted it was established that attenuation of magnetic variations in the waters of the Austrian Channel is slight due to the anomalously high conductivity of the bottom formations.

We did not have the opportunity to investigate attenuation of variations of the horizontal components at great oceanic depths because of several technical difficulties. Nonetheless using the equipment on hand we succeeded in investigating the nature of the vertical distribution of vertical component variations.

Inasmuch as δZ is the normal component to the surface of the medium interface, then, according to the boundary conditions for a magnetic field when $\mu = \mu_0$, δZ will be continuous when crossing the air-sea interface /8/. For plane electromagnetic waves this boundary condition will be satisfied exactly. A. T. Price showed that only when the source has finite dimensions will δZ attenuate in the sea the same as the density of the induced current,

i.e., considerably more weakly than tangential components δH and δD /20/.

In 1964, ice floe station Severnyy polyus-13 investigated the attenuation of δZ with the underwater Z-variometer. The field MVS AANII (magnetic variation stations of the Arctic and Antarctic Institute) operated on the ice surface. In all, more than 30 synchronous elementary variations of a quasisinusoidal form having periods from 300 to 3,000 s were taken. The underwater instrument operated in the 200-500-m horizons. The depth of the sea varied from 1,000 to 1,500 m during the observations. Reference /15/ showed the $\frac{\delta Z_0}{\delta Z_h}$ ratios determined on the SP-13 as well as the experimental points obtained in 1966 in the Austrian Channel at a depth of 200 m using the bottom MVS.

All experimental values of $\frac{\delta Z_0}{\delta Z_h} \approx 1$. Since the geoelectric conditions on the ice station and in the channels were totally different, it can be concluded that the attenuation of δZ in the seawater layer was generally negligible and almost independent of the electrical conductivity of the seawater and the bottom. The investigations made of local magnetic induced anomalies in the Arctic Ocean attest to the different causes for the anomaly in each individual case. The delineated anomalous part bears information on the source of the anomaly that can be used in specific applications.

To investigate magnetospheric processes it is proposed to set up a broad net of magnetic variation stations both on land and at sea. To interpret the data reliably, the distortions introduced by the induced anomalies in the observed field must be accurately taken into account. On the other hand, the induced anomalies make it possible to identify the geologic structural anomalies at different levels of the earth's crust and mantle. The interpretation of such anomalies is an important link in geophysical methods of exploring resources and studying the deep structure of the Earth. The newest methods of geoelectric sounding and profiling -- magnetic variation sounding and magnetic variation profiling -- are based on this.

As was shown earlier, the nature of the attenuation of magnetic variations in the depths of the oceans also makes it possible to draw conclusions as to the geoelectric structure of the ocean floor. This is the basis of gradient methods of marine magnetic variation sounding. Moreover, in investigations of magnetospheric processes it is necessary to know the degree of attenuation of variations in the sea to interpret the magnetograms of bottom magnetic variation stations. This same data is required to account for variations when conducting magnetic surveys of the seas and oceans. Oceanographic investigations also require a knowledge of local features of geomagnetic variations.

On the one hand, the variations create interference when using electro-

magnetic methods of measuring water currents. But on the other hand, the electromagnetic fields of the currents themselves bear information about the currents and can be used to measure them. Information on the electrical conductivity of the bottom formations obtained as the result of electromagnetic soundings make it possible to determine the impedance ratio $\frac{\delta E}{\delta H}$ thereby making it possible to calculate the correction δE from measured δH variations and thus to correct the current velocities measured by electromagnetic current meters during magnetic disturbances.

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