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Generation of microseisms in the coastal area

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ABSTRACT

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This paper considers the regularities and properties of microseisms and swell in the coastal area of East Kamchatka. Based on extensive experimental data, dependences were obtained of the value of coastal swell and the amplitude of microseisms on the period. A relationship was revealed between the vertical and horizontal components of microseisms. The dependence of the amplitude of microseisms on the value of swell is non-linear for both the horizontal and vertical components. For the horizontal component, the non-linearity is much more significant.

The mechanism of the generation of horizontal components by bottom friction is considered. Analysis of the hydrodynamic conditions of swell in the coastal area shows the possibility of the generation of vertical and horizontal components by, respectively, bottom pressure and bottom friction. Estimations were made of the dimensions of the microseismic generation areas in the coastal zone.

1. Introduction

The problem related to the mechanism which generates the microseisms has no definite solution. This is probably related to both the complexity of the process of energy transfer from the water layer to the solid medium and the possibility of the existence of several physical transfer mechanisms. Theoretical estimates show the possibility of the existence of several mechanisms (Hasselmann, 1963) but these estimates are made for the ideal cases. The real conditions may complicate one mechanism, simplifying at the same time the situation when excluding other mechanisms.

The present paper considers the regularities and properties of microseisms in the 3-8 s range in Kamchatka. Simultaneously, analysis of similar regularities is made for swell in the coastal zone; the regularities revealed are compared to determine the most likely mechanisms which generate microseisms.

2. Results

Observational data on microseisms and swell in the coastal zone of the main gulfs of the Kamchatka east coast (Avachinsky, Kronotsky and Kamchatsky) were obtained for both different cyclonic and calm meteorological situations. The full range of variations in microseism amplitudes was 40 db. A detailed description of the instrumentation and observation methods has been given earlier (Gordeev, 1979; Gordeev and Chebrov, 1979).

One of the main properties of swell and microseisms is the dependence of signal on period. Figure 1 illustrates the yearly variation in amplitudes and predominant periods of microseisms obtained from observations at one seismic station. The curves were obtained by averaging the data for each month. We see that the increase in microseism amplitude brings about the increase in period. This is apparently related to the hydrody-

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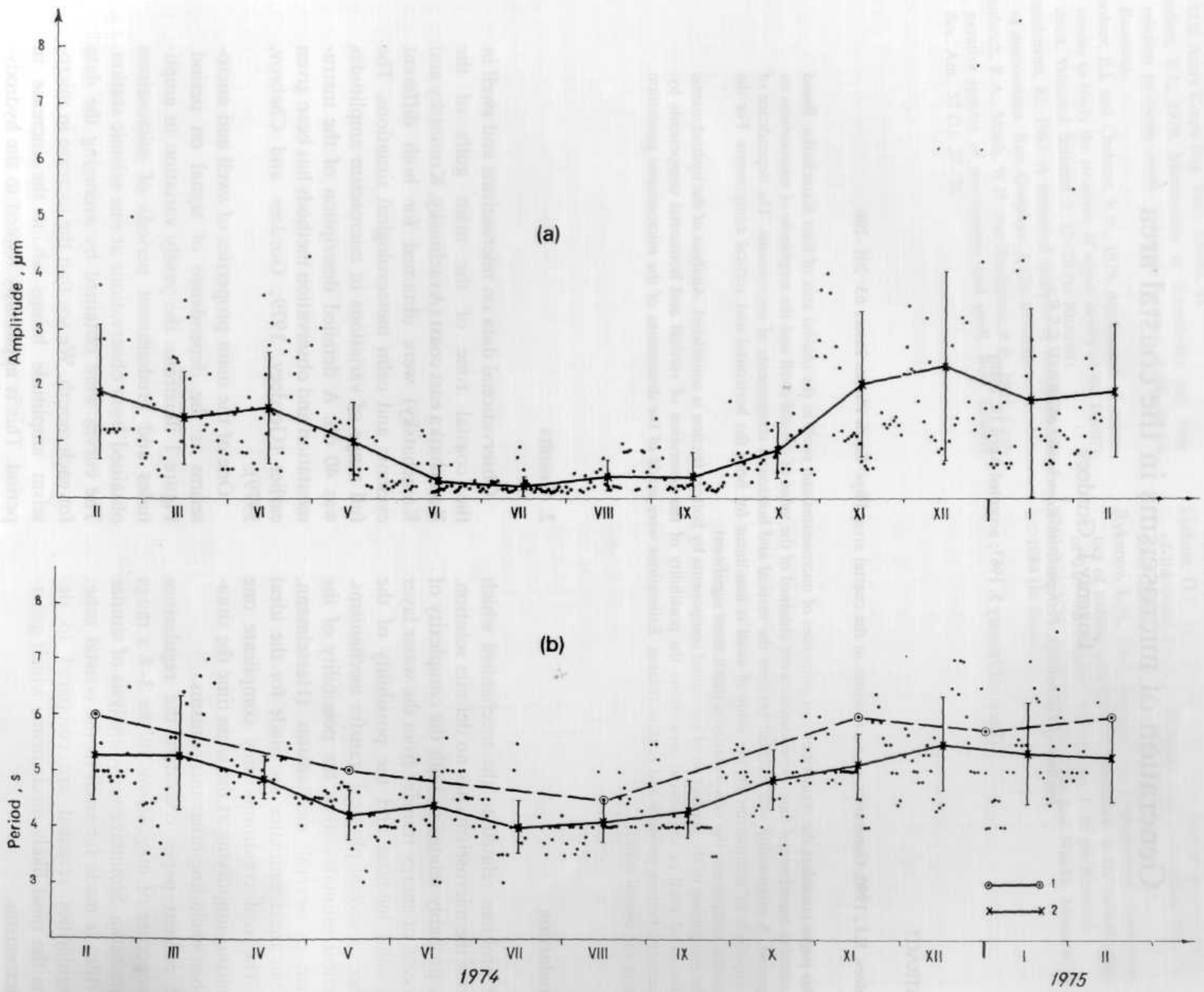

Fig. 1. Seasonal variations in the parameters of microseisms in Kamchatka from observations at the Petropavlovsk seismic station, (a) Variations in microseismic amplitudes by diurnal averagings and the average monthly variations with r.m.s deviation: (b) variations in microseismic periods. The regime of averaging is the same as for amplitudes. Curve 1 indicates the average swell periods near Kamchatka with 50% guarantee (Anon., 1968). Curve 2 indicates the average monthly variations in microseismic periods.

namic peculiarities of ocean swell generation because only the microseisms reflect the characteristics of this swell.

In fact, for wind swell a relationship exists between the mean wave height and the average period which is expressed analytically as (Krylov 1956)

$$\bar{h} = \begin{cases} 0.059\,T^2 & T < T_0 \\ 0.019\,\bar{T}V & T > T_0 \end{cases}$$

where $\bar{T}_0 = 0.32V$, V is the wind velocity (m s^{-1}), $\bar{h}$ is the wave height (m), and T is the period (s).

Consequently, the characteristic features of microseisms are closely correlated with the same features of ocean swell. The predominant periods of microseisms vary during a year within the range 3–7.5 s.

In addition to the general regularities which we wish to reveal for swell and microseisms, it is of

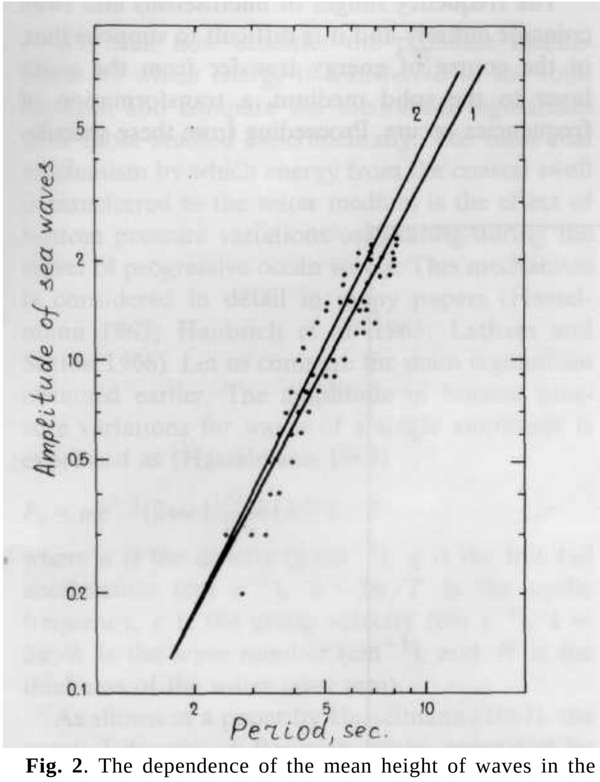

Fig. 2. The dependence of the mean height of waves in the coastal zone on the average period. (1) Obtained by the method of least squares using experimental points; (2) analytical from Krylov (1956).

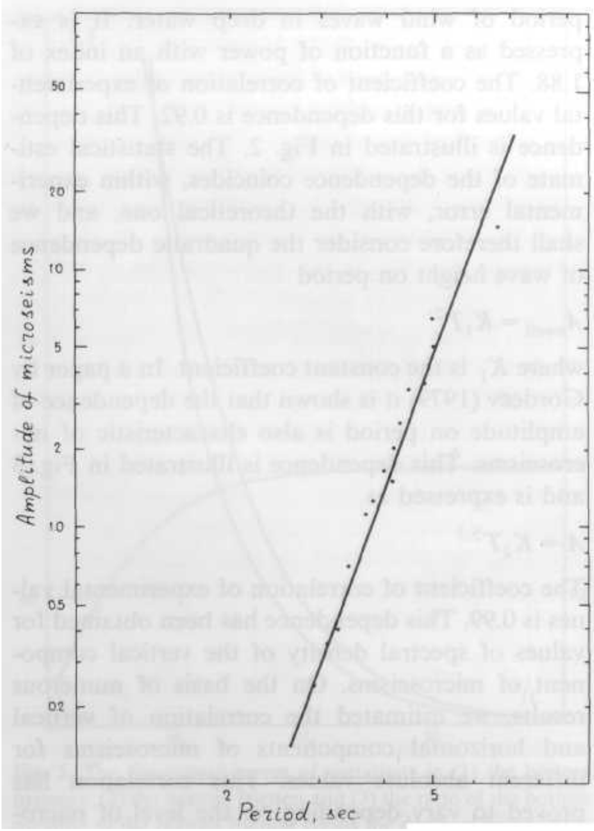

Fig. 3. Experimental dependence of the amplitude of the vertical component of microseisms on the period.

interest to compare the absolute values of their periods as well as the character of their variations during a year. The generalized graph of variations in swell periods with 50% guarantee for the region of the Pacific Ocean near Kamchatka (Anon., 1968) is indicated by the broken line in Fig. l(b). The course of the curve of variations in swell periods practically coincides with the averaged course of microseismic periods. The absolute level differs, on average, by 0.7 s, which is fairly natural because the averaging of the microseismic periods in our case is not the same as the 50% guarantee, although these values should be close to the real values.

2.1. Properties of swell and microseisms

As stated above, there is an analytical relationship between the mean height and the average period of wind waves in deep water. It is expressed as a function of power with an index of 1.88. The coefficient of correlation of experimental values for this dependence is 0.92. This dependence is illustrated in Fig. 2. The statistical estimate of the dependence coincides, within experimental error, with the theoretical one, and we shall therefore consider the quadratic dependence of wave height on period

$$A_{\text{swell}} = K_1 T^2$$

where K_1 is the constant coefficient. In a paper by Gordeev (1979) it is shown that the dependence of amplitude on period is also characteristic of microseisms. This dependence is illustrated in Fig. 3 and is expressed as

$$A = K_2 T^{5.3}$$

The coefficient of correlation of experimental values is 0.99. This dependence has been obtained for values of spectral density of the vertical component of microseisms. On the basis of numerous results, we estimated the correlation of vertical and horizontal components of microseisms for different absolute values. This correlation has proved to vary depending on the level of microseisms. This may point either to a different mechanism of formation of these components or the existence of non-linear conditions in the medium on the ways microseisms are propagated. The second supposition is hardly probable because of medium isotropy and uniformity of propagation conditions for vertical and horizontal components. It is most likely that the differences for the vertical and horizontal components of microseisms originate in the source. The dependence of the vertical component on the horizontal component is illustrated in Fig. 4. Analytically it is expressed as

$$A_{\text{horiz.}} = K_3 A_{\text{vert.}}^{1.4}$$

This dependence differs from the linear one which points to the existence of non-linearity at least for one of the components. To test this supposition, we shall estimate the dependence of the vertical and horizontal components of storm microseisms versus the value of coastal swell.

The frequency ranges of microseisms and swell coincide entirely and it is difficult to suppose that, in the course of energy transfer from the water layer to the solid medium, a transformation of frequencies occurs. Proceeding from these specula-

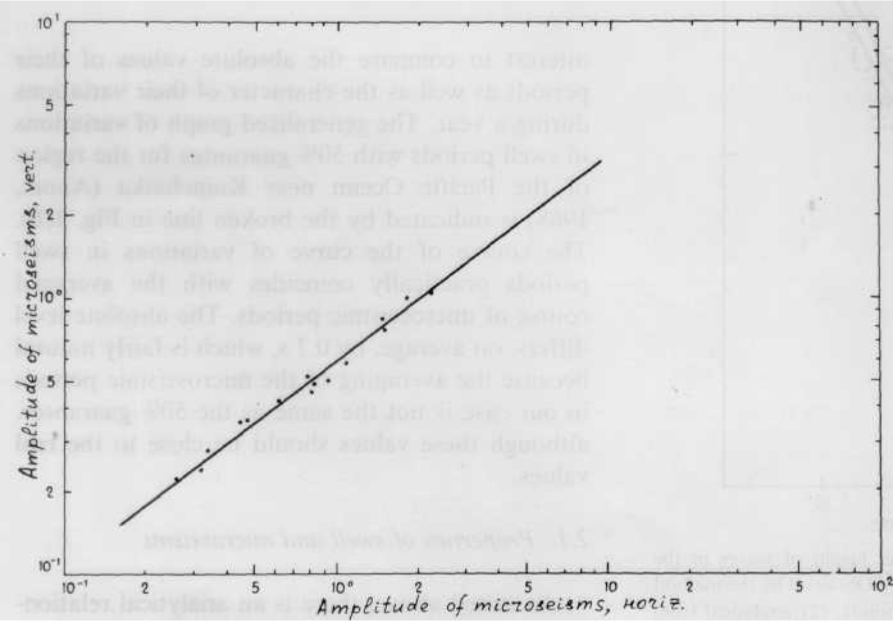

Fig. 4. The relation between the vertical and horizontal components of microseisms.

tions, the dependence of the amplitude of microseisms on the value of swell may be obtained from the synthesis of dependences of swell and microseisms on period (see Figs. 2 and 3). For the vertical component, taking into account the dependence of maximum spectral values and the dependence of amplitudes shown in Figs. 2 and 3, respectively, we obtain the relation

$$A_{\text{vert.}} = K_4 A_{\text{swell}}^{1.3}$$

For the horizontal component, taking into account the dependence in Fig. 4, we obtain

$$A_{\text{horiz.}} = K_5 A_{\text{swell}}^{1.8}$$

The transfer of swell energy for the vertical and horizontal components of microseisms occurs with non-linear effects. For the horizontal component, non-linearity is much more significant.

3. Discussion

We shall now consider the probable mechanisms by which energy is transferred to the solid medium and compare the theoretical regularities with those studied experimentally. The most real mechanism by which energy from the coastal swell is transferred to the water medium is the effect of bottom pressure variations originating during the travel of progressive ocean waves. This mechanism is considered in detail in many papers (Hasselmann 1963; Haubrich et al. 1963; Latham and Sutton 1966). Let us compare the main regularities obtained earlier. The amplitude of bottom pressure variations for waves of a single amplitude is expressed as (Hasselmann 1963)

$$P_0 = \rho g^{3/2} (2\omega c)^{1/2} ch(kH)$$

where ρ is the density (g cm^{-3}), g is the free fall acceleration (cm s^{-2}), $\omega = 2\pi/T$ is the cyclic frequency, c is the group velocity (cm s^{-1}), $k = 2\pi/\lambda$ is the wave number (cm^{-1}), and H is the thickness of the water layer (cm).

As shown in a paper by Hasselmann (1963), the spectral density of Rayleigh waves generated by bottom pressure variations in the coastal zone is inversely proportional to the fifth power of frequency, i.e. it is proportional to the fifth power

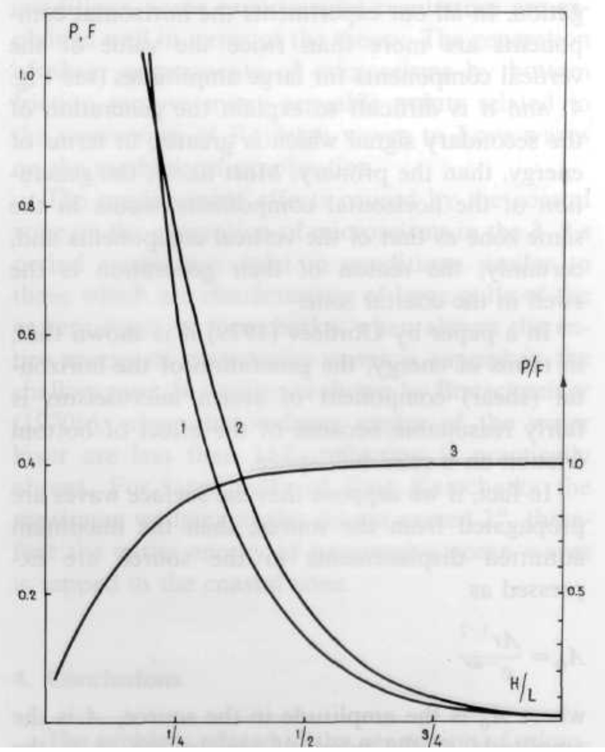

Fig. 5. The theoretical curves of variations in (1) the bottom pressure, (2) the bottom friction and (3) the ratio of the bottom pressure to the bottom friction versus the depth of water layer.

of period that responds simply to the experimental dependence obtained. Thus, we may consider that the generation of the vertical component of storm microseisms occurs largely by variations of bottom pressure due to progressive wave travel. Because the amplitude of pressure variation is inversely proportional to the hyperbolic cosine of the depth to wavelength ratio, then for a constant wavelength the value of the pressure variation decreases rapidly with depth and the microseismic generation zone is confined to the coastal strip to depths of the order of wavelength (Fig. 5). The bottom pressure acts on the bottom by the normal component and it is therefore practically impossible to account for the generation of the horizontal component by bottom pressure. It could probably be assumed that the origin of the horizontal component was due to the transformation of the vertical component because of the effects of heterogeneities of the medium on the method of propagation. In all our experiments the horizontal components are more than twice the value of the vertical components for large amplitudes (see Fig. 4) and it is difficult to explain the generation of the secondary signal which is greater, in terms of energy, than the primary. Most likely, the generation of the horizontal components occurs in the same zone as that of the vertical components and, certainly, the reason of their generation is the swell in the coastal zone.

In a paper by Gordeev (1979) it is shown that, in terms of energy, the generation of the horizontal (shear) component of storm microseisms is fairly reasonable because of the effect of bottom friction on a solid half-space.

In fact, if we suppose that the surface waves are propagated from the source, then the maximum admitted displacements in the source are expressed as

$$A_0 = \frac{Ar^{1/2}}{e^{-\alpha r}}$$

where A_0 is the amplitude in the source, A is the amplitude at the point of registration, r is the distance to the source, and α is the absorption coefficient. Taking the absorption to be of the order of 10^{-3} km^{-1} (Rykunov 1967), the distance to be not more than 10^2 km, and the maximum displacements in the observation point to be 10 μm, we obtain $A_0 = 1$ mm. For the harmonic vibrations, the energy of the unit of volume for one period is

$$E = \rho V^2 = \rho\left(\frac{A}{T}\right)^2$$

where T is the period of vibrations and ρ is the density of the medium. Considering $T = 4$ s and $\rho = 2.5$ g cm^{-3}, we obtain $E = 1.6 \times 10^{-3}$ erg. The power per unit volume is $P_s = E/T = 0.4 \times 10^{-3}$ erg s^{-1}.

It should be noted that the power of the source, if we use this estimate, is overestimated significantly because we supposed the source to be of small dimensions. In fact, there is a volume source of large dimensions and the correct estimate may be made only if we use statistical methods.

The energy losses of the progressive wave caused by bottom friction for unit time per unit area are expressed as (Bretschneider 1970a)

$$P_f = \tfrac{4}{3}\pi^2 \frac{\rho f h^3}{T^3[sh(kH)]^3}$$

Here ρ is the density of water, f is the "calibrating" friction coefficient, h is the wave height, T is the wave period, H is the depth of the water layer, $k = 2\pi/\lambda$ is the wave number, and λ is the wavelength. Taking $h = 2$ m, $T = 4$ s, $\rho = 1$ g cm^{-3}, $\lambda = 40$ m, we obtain $P_f = 10^5 f$ erg s^{-1} for $H = 10$ m.

Using experimental data on swell observations in the Lake Okeechobee (Florida), Bretschneider and Reid (1954) estimated the friction coefficient to be of the order of 10^{-2}. Using this coefficient, we obtain the energy losses caused by bottom friction to be greater than the energy of a seismic source by more than 6 orders of magnitude.

We shall now consider the theoretical dependences and the absolute values of forces for the processes occurring in the border layer near the bottom. The tangential frictional stresses at the boundary of a liquid and a solid body in the case of a fluctuating liquid layer can be expressed as (Batchelor 1970)

$$F_{fr} = \mu\left\{\mathrm{Re}\left(\frac{\partial u}{\partial y}\right)_{y=0}\right\}$$
$$= \mu\left\{\mathrm{Re}\left[(1+i)\frac{U}{\delta}e^{i\omega t}\right]\right\}$$
$$= \frac{\mu U}{\delta}(\cos\omega t - \sin\omega t)$$

where μ is the viscosity, U is the external flow velocity in relation to the border layer, $\delta = (2\mu/\rho\omega)^{1/2}$, and ω is the cyclic frequency of fluctuation of the external flow velocity. The amplitude of the external flow velocity in the bottom layer in the case of progressive wave propagation can be expressed as

$$U = \frac{h\omega}{2sh(kH)}$$

where h is the wave height, $K = 2\pi/\lambda$ is the wave number, and H is the depth of the water layer. The amplitude of frictional stress for waves of a single height can finally be written as

$$F_{fr} = \frac{(\mu\rho)^{1/2}\omega^{3/2}}{2\sqrt{2}\,sh(kH)}$$

The value of the frictional stress is inversely proportional to the hyperbolic sine of the ratio of depth to wavelength. Figure 5 shows the dependence of the amplitude of frictional stresses on the H/λ value. Here we take the value of turbulent viscosity of ocean water in the case of turbulent mixing. As shown by Shuleikin (1968), the turbulent viscosity, providing the mixing is active, may increase by 6-7 orders of magnitude compared with the molecular viscosity. Suspended solid particles, considerable roughness and, in many cases, vegetation are always present in the real bottom conditions as distinct from the ideal liquid-solid body contact. The curve (Fig. 5) showing the change in frictional stress amplitude was drawn for a turbulent viscosity differing by 8 orders of magnitude from the molecular viscosity, which is quite real for the conditions in the coastal shallow zone. In this version the bottom friction and the bottom pressure values are of the same order.

Variations in bottom friction have another power of increase with decreasing value of depth-to-wavelength ratio in comparison with the bottom pressure. The change in ratio of bottom pressure to bottom friction is shown in Fig. 5. Beginning from H/λ = 1/2, decreasing this ratio results in a much more rapid increase in bottom friction until at H/λ = 1/8 it exceeds the bottom pressure twofold subject to the bottom friction and the bottom pressure being equal at depths of the order of one wavelength.

Thus the generation of microseisms by bottom friction may occur in a coastal strip coinciding with the zone of microseism generation by bottom pressure. The width of this strip varies depending on the length of progressive ocean waves in the coastal area and, as follows from the bottom pressure versus bottom friction ratio (see Fig. 5), the increase in pressure and friction occurs non-uniformly with increasing wavelength. This, in particular, may explain the increase in the ratio of the horizontal components of microseism versus the vertical components with increasing microseismic intensity.

Using the general regularities and estimates of absolute values we may consider the generation of seismic signals by bottom friction to be the real phenomenon. Despite the complexity of the real conditions, some experimental regularities are explained well in terms of the theory. The generation of shear components of microseisms by bottom friction remove many arguable points related to the conversion of Rayleigh waves to Love waves on the methods of propagation.

The predominant effects caused by the coastal zone on the generation of microseisms in the 3-8 s period range may exist in conditions similar to those which are characteristic of large gulfs of the eastern coast of Kamchatka, when almost the entire energy of progressive waves is sapped in the shallow zone. In reality, as shown by Bretschneider (1970b), when the wedging angles of the water layer are less than 15°, reflection is practically absent. For large gulfs of East Kamchatka the maximum wedging angles do not exceed 2 °, therefore the entire energy of progressive ocean waves is sapped in the coastal zone.

4. Conclusions

The problem related to the generation of microseisms is of practical significance in assessing the potential sources of seismic noise in terms of the bottom profile in the coastal zone and in terms of the ocean swell regime. The mechanisms for the generation of microseisms presented in this paper should be considered in terms of their reality on the basis of estimates of absolute values and regularities. Knowledge of the conditions in the real bottom layer and a complete set of observations of hydrological parameters are needed to clarify the characteristics of these mechanisms.

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