SEISMIC EFFECT OF TROPICAL CYCLONES

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Abstract

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It has recently been shown ([1,2]) that, in the microseisms of quiet intracontinental regions, a significant share of the energy is accounted for by longitudinal waves with a very high apparent velocity. Although the earliest data already made it possible to suppose ([1]) that the main source of microseismic longitudinal waves is the disturbance of the sea surface, up to now nothing more definite has been known about the origin of these waves. In the present note we present some results of a study of the sources of microseismic longitudinal waves based on materials from an analysis of the structure of microseisms already published by us earlier ([3]). The method used for analyzing the structure of microseisms is, in principle, a decomposition of a random field by wave numbers. If the statistical spectrum of the wave numbers ω_1 and ω_2 has extrema, then these are associated with the presence of relatively intense plane waves in the field. In this case, for a wave with circular frequency ν, the apparent velocity v and propagation azimuth α are measured:

$$v = \nu/\sqrt{\omega_1^2 + \omega_2^2}, \qquad \alpha = \operatorname{arc\,tg}(\omega_2/\omega_1).$$

From the apparent velocity of a longitudinal wave one can judge the epicentral distance. The azimuth and epicentral distance determine the position of the source.

We found the coordinates of the sources of longitudinal waves for 5 records of 4–6-second microseisms made on different days of October 1961 using a group of stations in the region of Ust-Kamenogorsk. The 70% confidence interval of the estimate of the epicenter coordinates, at a distance of the order of 10^4 km, is approximately determined as a circle of radius 1500 km. It should be noted that, by quite obvious improvements, the accuracy of determining the epicenter by this method can be brought to 300–500 km.

As a result of comparing the epicenters of microseismic longitudinal waves with synoptic data, it was established that one of the main sources of these waves is tropical cyclones (typhoons) in the Pacific Ocean. The simplest case is given as an example. Figure 1 shows a world synoptic map corresponding to 24 h Greenwich time on 5.X.1961. From the microseism records made on that day,

Fig. 1. Legend: 1—pressure in mb; 2—epicentral region; 3—group of seismic stations; 4—seismic shadow zone

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a single longitudinal wave is confidently distinguished, with an epicenter in the equatorial region of the Pacific Ocean (epicentral distance $\sim$ 9000 km). The synoptic situation is simple: in the large region of the ocean surrounding the epicentral area, the sole source of disturbance is the tropical cyclone Violetta, with its center at 18° N, 143° E. The small discrepancy in the position of the typhoon and the epicentral area is probably explained by the error in estimating the epicentral distance.

From the amplitude of the longitudinal wave one can estimate the source parameters. For this we shall use the expression for the mean intensity of a longitudinal wave radiated by a vertical force $P_0 e^{i\nu t}$, applied to a circular disk of small diameter on the surface of a solid half-space ([4]):

$$\delta W = \pi f^2 m^4 P_0^2 |\Phi(e)|^2 \cos e \, de / \rho a^3. \tag{1}$$

Here δW is the mean radiation intensity into a ray tube of small aperture de; the angle e is measured from the horizontal axis and is equal to $\pi/2$ for the downward-directed axis; $[\Phi(e)]^2$ is the radiation pattern, which in the case of interest to us is approximately equal to 0.05; ρ is density; a is the velocity of propagation of the longitudinal wave; $f = \nu/2\pi$; m is the ratio of the velocities of longitudinal and transverse waves. Although in our case the pressure is applied to the surface of the water layer, there is reason to suppose that the formula used gives a sufficiently good approximation. On the other hand, on the basis of the formula of paper ([5]), we may write

$$\delta W = 2\pi R^2 \sin\Delta \sin e \, \rho a \overline{V^2} \exp\left(\nu \int \frac{dt}{Q}\right) |d\Delta|. \tag{2}$$

Here R is the radius of the Earth; Δ is the epicentral distance; $\overline{V^2}$ is the mean square of the displacement velocity in the incident longitudinal wave at the place of observation; t is the travel time along the ray; Q is the quality factor of the medium. In addition, we have ([6])

Fig. 1. Legend: 1—pressure in mb; 2—epicentral region; 3—group of seismic stations; 4—seismic shadow zone

$$\frac{\sin e}{\cos e} \left| \frac{d\Delta}{de} \right| = \frac{R(\mathrm{tg}^2 e - \sin^2 e)^{1/2}}{a \, |d^2 T / d\Delta^2|_i}, \tag{3}$$

where $d^2T/d\Delta^2$ is the second derivative of the hodograph of the longitudinal wave. Using (1), (2), and (3), we obtain

$$P_0^2 = \frac{2R^3\rho^2a^3\overline{V^2}\sin\Delta\exp\left(\nu\int\frac{dt}{Q}\right)(\operatorname{tg}^2 e - \sin^2 e)^{1/2}}{[\Phi(e)]^2 f^2 m^4 |d^2T/d\Delta^2|}.$$

Substituting into this formula the values of ρ, a, m for the mean parameters of the Earth' s crust; the values of $\sin\Delta$, $\operatorname{tg} e$, $\sin e$, $|d^2T/d\Delta^2|$, $\int dt/Q$ for $\Delta = 80°$; $f = 0.2$ cps, $\overline{V^2} = 10^{-11}$ cm^2 sec^{-2} according to the results of our observations, we obtain $P_0 \approx 6.5\cdot10^{16}$ dyn.

Let us make use of the well-known theory of the excitation of microseisms by opposing gravitational waves on the sea surface [7]. The storm region may be divided into a large number of independent small radiators. The resultant force depends on the total area of the storm S, the mean wave height in the opposing groups h, and the area Ω occupied by each group in the plane of wave numbers. In this case

$$P_0 \approx 4\pi\rho_0 h^2\nu_0^2(S/\Omega)^{1/2}. \tag{4}$$

Here $\rho_0 = 1$ g/cm^3 is the density of water; $\nu_0 = 2\pi/T_0$, T_0 is the mean period of sea waves. In the central part of the cyclone the wave energy should be distributed approximately uniformly among all directions; let us assume that the range of wave periods lies between 8 and 12 sec. Then $\Omega = 0.5\cdot10^{-6}$ cm^{-2}, $\nu_0 \approx 0.63$ sec^{-1}. Substituting $P_0 = 6.5\cdot10^{16}$ dyn into (4), we find $h^2S^{1/2} \approx 5.8\cdot10^{12}$ cm^3. Assuming, in particular, $h = 3\cdot10^2$ cm, we find $S \approx 4\cdot10^{15}$ cm^2 $= 4\cdot10^5$ km^2. These estimates are in good agreement with the real scales of the phenomenon.

It is known that a vertical force applied to the surface of a solid half-space gives 68% of the seismic energy to a Rayleigh wave and only 7% to a longitudinal wave [4]. The presence of a water layer increases the transfer of energy into the surface wave by roughly an order of magnitude. The geometrical spreading of the front and the associated decrease of amplitude with distance are substantially smaller for a surface wave than for a longitudinal one. Nevertheless, the long travel time, low quality factor, and variable structure of the Earth' s crust lead to complete attenuation of the surface wave if the source is located far out in the ocean. This is precisely what explains the failure of numerous attempts at early detection of cyclones over the ocean by means of surface waves. Moreover, observations in most cases were carried out near the coast; the high level of noise of coastal origin hindered the isolation of weaker waves from distant sources.

The difference in our results is due to the fact that the observations were carried out in the center of the continent, where the influence of disturbances—surface waves of coastal origin—was reduced to a minimum. In addition, a more effective method of analyzing the record was used.

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Note: Figure translations are in progress. See original paper for figures.