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Abstract

Full Text

GEOPHYSICS

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DETERMINATION OF THE ENERGY FLUX OF EARTHQUAKE SOURCES ON THE BASIS OF SEISMIC ACTIVITY

As is known ([1–3]), the numbers N of earthquakes of different energy E in a source obey the statistical "law of earthquake recurrence" $N = N(E)$, which, in the zeroth approximation, may be represented by the equality

$$N = A \cdot 10^{-\gamma(K-K_0)}, \tag{1}$$

for $K \le K_{\max}$, and $N = 0$ for $K > K_{\max}$. Here N is the number of earthquakes of a given class K, reduced to unit time and space, with energy in the source within $E = 10^{K\pm0.5}$ J; $A = N|_{K=K_0}$ is the seismic activity, determined at the selected value $K = K_0$; $\gamma = -d\lg N/dK$ is a conditionally constant parameter determining the slope of the approximately rectilinear recurrence plot of earthquakes in the coordinate system $K, \lg N$; $K_{\max}$ is the energy index $E_{\max} = 10^{K_{\max}}$ of the maximum earthquake occurring in the given region.

To determine the mean specific power, or "flux of seismic energy," w, of all n earthquake sources released over time T in some region of volume V or area (on the earth' s surface) S, one would have to find the sum

$$w = \frac{1}{VT}\sum_{i=1}^{n} E_i \quad \text{or} \quad \frac{1}{ST}\sum_{i=1}^{n} E_i, \tag{2}$$

where E_i is the seismic energy of each earthquake in this region. This method of determining w and mapping this quantity is acceptable, however, only for small-scale maps of a survey character. For a detailed comparison of w with the geological-geophysical setting of regions and for studying the time course of the seismic regime, this approach encounters great difficulties because the quantity w, determined mainly by rare strong earthquakes, fluctuates strongly.

To reduce the fluctuation, it was proposed ([2]) to use the dependence

$$w = \int_{-\infty}^{K_{\max}} NE\,dK, \tag{3}$$

where N is the density of the distribution of earthquakes in time, in space, and in K, a quantity determined taking into account the recurrence law, in particular and usually in the form (1). In this case

$$w = \frac{10^{\gamma K_0}}{(1-\gamma)\ln 10} A \cdot 10^{(1-\gamma)K_{\max}}. \tag{4}$$

Here the difficulty has not yet been overcome, since the quantity $K_{\max}$, considered independently in each separate region, remains strongly fluctuating. But it can be bypassed ([4]) if $K_{\max}$ is related by a general dependence to less fluctuating quantities. In ([5]) a correlation relation was established between $K_{\max}$ and the more stably determined quantity of seismic activity A:

$$A = A_M \cdot 10^{\beta(K-K_M)}. \tag{5}$$

Here the parameters A_M and β are determined from observations; K_M is the chosen value of $K_{\max}$ at which the value of the parameter A_M is established. Eliminating now $K_{\max}$ from the system of equations (4) and (5), we obtain

$$w = \frac{10^{K_M+\gamma(K_0-K_M)}}{A_M^{(1-\gamma)/\beta}(1-\gamma)\ln 10} A^{1+(1-\gamma)/\beta}, \tag{6}$$

which gives the desired solution. Its greater stability in comparison with (2) is achieved by the fact that here, besides local data, use is made of the average regularities (1) and (5), obtained from more extensive statistics.

Let us give some numerical comparisons. If in (6) we take, according to (5), $\gamma = 0.5$, $K_0 = 10$, $K_M = 15$, $\lg A_M = \overline{2}.8$, $\beta = 0.2$, then we obtain $w = 2.74 \cdot 10^{15} A^{3.5}$ J $(1000\ \text{km}^2)^{-1}$ yr^{-1} ($A = A_{10}$ is the activity measured by the number of earthquakes with $K_0 = 10$ over an area of 1000 km^2 per year). Then for regions with moderate ($A = A_1 = 0.01$), strong ($A_2 = 0.1$) and very strong ($A_3 = 1$) seismicity we find, respectively, $w_1 = 8.7 \cdot 10^{-9}$, $w_2 = 2.8 \cdot 10^{-5}$, and $w_3 = 8.7 \cdot 10^{-2}$ J m^{-2} s^{-1}. If we take the value $\gamma = 0.6$, which apparently corresponds better to the seismic energy radiated from the surface of the source ([3]), then, for the same values of the other parameters, from (6) we obtain $w = 2.72 \cdot 10^{14} A^3$ J $(1000\ \text{km}^2)^{-1}$ yr^{-1}, and then $w_1 = 8.6 \cdot 10^{-9}$, $w_2 = 8.6 \cdot 10^{-6}$, and $w_3 = 8.6 \cdot 10^{-3}$ J m^{-2} s^{-1}, which differs from the preceding result by no more than about an order of magnitude.

Let us compare these values with the heat flow of the Earth. The mean heat flow through the Earth' s surface is ([5]) $Q = 1.2 \cdot 10^{-6}$ cal cm^{-2} s^{-1} $= 5.0 \cdot 10^{-2}$ J m^{-2} s^{-1}. In this case the energy flux carried by elastic waves from earthquakes is comparable in order of magnitude with the mean heat flow of the Earth only in areas of very high seismicity. Of course, it would be more interesting to compare the seismic-energy flux with the heat flow in local areas, but for this we still lack detailed thermometric data.

It appears that a dependence of the form (6), relating seismic activity A to the flux w of seismic energy of earthquake sources and based on two observed dependences—the earthquake recurrence law (1) and the relation between maximum earthquakes and activity (5)—will make it possible to map the seismic-energy flux in sufficient detail for comparison with other mapped quantities, primarily with geodetic and geological data on slow deformation movements of the Earth. This should contribute to the development of a general physical theory of tectonic movements, which include the "seismic flow," whose energy has been discussed here. It should be borne in mind that the numerical values of the parameters entering into (1) and (5) must be refined from observations in order to determine both the best mean values and the deviations from them that may occur under local conditions and over limited periods of time.

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