Strain data and seismicity

Varga Péter,Mentes Gyula

Geodetic and Geophysical Research Institute
H-9401, Sopron POB 5, Hungary
(E-mails: varga@seismology.hu, mentes@seismology.hu)

1.Introduction

The study of the displacement fields of the Earth surface can be useful in the study of the properties of the past and future potential earthquakes. With the use of Kostrov equation (*1974*) the possible sum of seismic moments can be estimated for a given source zone, if its area is described with the use of geodetic measurements. The possible displacement fields connected with seismo-tectonic activity should be estimated with the use of model calculations. The necessity of such a theoretical work is connected with the very long return period of destroying events (M≥ 6.5).

2. Some problems of the seismicity of the Earth

Since the birth of modern instrumental seismology at the end of XIXth century elapsed a little more than a century. In the same time the return period of characteristic seismic events even in case of most active seismic sources is well above hundred year. This circumstance makes extremely complicated to make predictions concerning future seismic events. To illustrate the deficiency of our knowledge serves as an example a comparison of seismicity of two seismic zones. (Table 1). It can be seen that the seismicity data concerning the Tangshan region just before the most destroying earthquake of XXth century (28 July !976, M=7.8) were very similar to those in Komarom with medium seismicity region of the Pannonian basin.

Table 1. Earthquakes in Tangshan and Komarom prior 1976.

Tangshan				Komárom			
Year	Epicentral intensity	M	Energy (Joule)	Year	Epicentral intensity	M	Energy (Joule)
1527	VII	5.5	$1.10 \cdot 10^{13}$	1599	VIII	5.6	$1.58 \cdot 10^{13}$
1567	VI	4.8	$8.41 \cdot 10^{11}$	1754	V	3.8	$3.16 \cdot 10^{10}$
1624	VII	6.3	$1.50 \cdot 10^{14}$	1759	V	3.8	$3.16 \cdot 10^{10}$
1795	VI-VII	5.3	$4.73 \cdot 10^{12}$	1763	IX	6.2	$1.26 \cdot 10^{14}$
1805	VII	5.5	$1.12 \cdot 10^{13}$	1783	VIII	5.3	$5.62 \cdot 10^{12}$
1880	VI	5.0	$2.00 \cdot 10^{12}$	1806	VII	5.0	$2.00 \cdot 10^{12}$
1934	VI	5.0	$2.00 \cdot 10^{12}$	1822	VI-VII	4.7	$7.08 \cdot 10^{11}$
1935	VI	5.0	$2.00 \cdot 10^{12}$	1822	VI	4.4	$2.51 \cdot 10^{11}$
1945	VIII	6.3	$1.50 \cdot 10^{14}$	1851	VII	5.0	$2.00 \cdot 10^{12}$
1974	V-VI	4.8	$1.00 \cdot 10^{12}$	1857	V	3.8	$3.16 \cdot 10^{10}$
1974	V-VI	4.8	$1.00 \cdot 10^{12}$	1923	V	3.8	$3.16 \cdot 10^{10}$

It can be concluded that between the first half of XVIth century and 1976 the number of the earthquakes with epicentral intensity ≥V was the same in case of both seismic sources and the energy released during this time interval was also similar ($3.35 \cdot 10^{14}$ and $1.52 \cdot 10^{14}$ Joule respectively).

From the five largest known earthquakes (M≥9) ever observed four was connected to the Pacific region and only one occurred outside from this area:

Chile (1960,*M*=9.5)
Alaska (1964,*M*=9.2)
Alaska (1957,*M*=9.1)
Kamchatka (1952,*M*=9.0)
Sumatra (2004,M=9.0)

It is remarkable that the four greatest seismic events of the Pacific region occurred in twelve years

The most of the seismic energy is released by the biggest earthquakes:92% of the seismic energy is connected to seismic events M≥7 (Table 2).

Table 2. Annual number and energy released by earthquakes of different magnitude classes.

Magnitudes	Annual average number	Contribution to the annual seismic energy
M≥8	1	49%
7.9≤*M*≥7	10	43%
6.9≤*M*≥6	10^2	4%
5.9≤*M*≥5	10^3	3%
4.9≤*M*≥4	10^4	1%

When the annual seismicity is discussed one should keep in mind that our knowledge globally is still not complete for the seismic events M≤ 5.0 (Fig.1). If the seismicity of the whole XXth century is considered it can be concluded that our earthquake catalogue is complete only for the seismic events above M=7, what means that we are able to describe the temporal distribution of 92 % of seismic energy only.

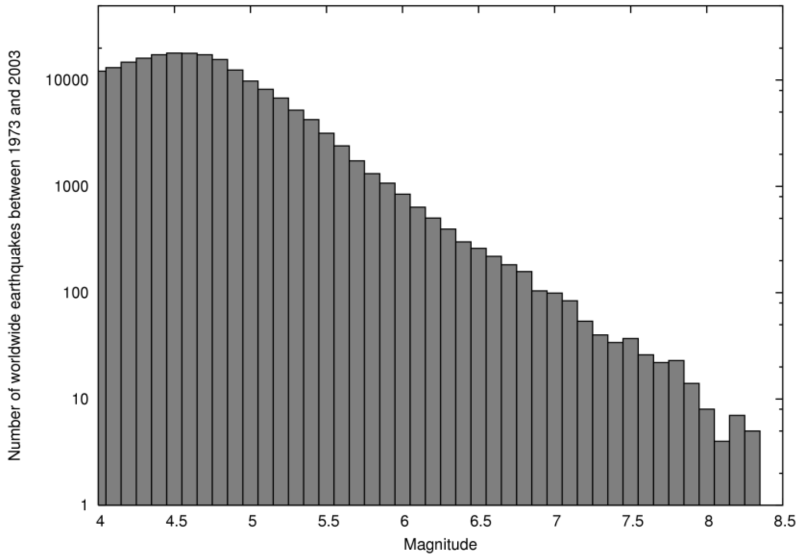

Fig.1 Histogram of earthquakes of the world for 1973-2004

In Section 4 we shall try to determine the volume of the earthquake source. For this purpose the determination of the aftershock area *A* is needed, which can be obtained with the use of the equation (*Kasahara, 1981*)

$$Lg\ A = 6.0 + 1.02 \cdot M \tag{1}$$

(*A* is expressed in cm^2)

A short study of Eq.1 shows that in case of

M= 6.0 we have A=1.3 $\cdot 10^2$ km^2
M= 7.0 A=1.4 $\cdot$ 10^3 km^2
M= 8.0 A=1.4$\cdot$ 10^4 km^2
M= 9.0 A= 1.5$\cdot$ 10^5 km^2
M= 9.5 A= 4.9$\cdot$ 10^5 km^2
M=10.0 A= 1.6$\cdot$ 10^6 km^2
M=10.5 A= 5.1$\cdot$ 10^6 km^2

From what follows that probably there is no possibility for generation earthquakes significantly bigger than M=10, because there is not seismic source zone on the Earth with area is big enough to generate such a giant seismic event.

3. Strain rates and their determination with the use of seismological data.

In geodesy the strain is determined as the ratio of the length variation to the total length $\varepsilon = \Delta L / L$ and the strain rate is $\frac{d\varepsilon}{dt} \approx \frac{\varepsilon}{\Delta}$. Of course the strain has no units while the strain rate is usually expressed in $year^{-1}$.

It is worth to mention here that the seismic strain is $\Delta = \frac{D}{L^F}$ (*D* - dislocation, L^F - fault length). Approximately $\Delta \approx DA^{-1/2}$. Here *A* is the contact area expressed in the same units like *D*. Kasahara (*1981*) has found that the maximal share rate is

$$\frac{d\varepsilon}{dt} = \frac{p}{\mu \cdot \Delta} \tag{2}$$

In above expression p is the tensional strength ($p \le 10^{7}$ Nm^{-2}), μ serves for the shear modulus ($\approx 3 \cdot 10^{10}$ Nm^{-2}) and Δt is the time-interval between the characteristic earthquakes.

The individual fault or system of faults have a preferred magnitude (*Huang et al., 1998*), an earthquake at which exhausts the seismic storage potential of the system, which must then recovered over a period of some time. An event of this magnitude is called characteristic earthquake. With the use of equation (2) if

$$\frac{d\varepsilon}{dt} \approx 10^{-6} \text{ then} \quad \Delta t \approx 10^{2} \text{ year}$$
$$\frac{d\varepsilon}{dt} \approx 10^{-7} \quad \Delta t \approx 10^{3} \text{ year}$$
$$\frac{d\varepsilon}{dt} \approx 10^{-8} \quad \Delta t \approx 10^{4} \text{ year}$$

The statistical mean of the strain rates determined with strainmeters worldwide is (*Varga, 1984*):

$$\frac{d\varepsilon}{dt} = 2.1 \cdot 10^{-} \text{ year}^{-}$$

The seismic strain rate according to Kostrov (*1974*) can be related to the sum of the seismic moment tensors of the earthquakes occurring in a volume ΔV during a time-interval Δt :

$$\frac{d\varepsilon}{dt} = \frac{\sum_{n=}^{N} M_{0ni}}{2\mu\Delta V} \tag{3}$$

If the orientation of the seismic sources is replaced by their average the seismic moment M_0 can be introduced in equation (3) instead of M_{0ni}

$$\frac{d\varepsilon}{dt} \approx \frac{\sum_{n=}^{N} M_{0n}}{2\mu\Delta V} \tag{4}$$

The value of the seismic moment M_0 can be derived from the value of momentum magnitude M_W (*Hanks and Kanamori 1979*)

$$\mathrm{Lg}M_0 = 1.50 M_W + 10.70$$

4. Strain rates derived from earthquake data and their connection to the return period of characteristic earthquakes.

Let us consider that in equation (4) Δt=250 year. The surface of the source volume can be estimated with the use of equation (1) and its thickness one can obtain if it is supposed that the focal depth is in the middle of the layer in which the stress accumulation takes place. In Table 3 the strain rates obtained with the use of equation(4) are listed for the case of the five greatest earthquake observed until now.

Table 3. Seismic strain rate values obtained for the earthquakes $M_W \geq 9.0$.

Place	Year	Mw	M_0 10^{22} Nm	Aftershock area (calc.) 10^5 km^2	Aftershock area (obs.) 10^5 km^2	Volume 10^6 km^3	dε/dt
Kamchatka	1952	9.0	3.9	1.5		3.0	$8.6 \cdot 10^{-7}$
Alaska	1957	9.1	5.6	1.9		3.8	$9.8 \cdot 10^{-7}$
Chile	1960	9.5	22.1	4.9		9.8	$1.5 \cdot 10^{-6}$
Alaska	1964	9.2	7.9	2.4	2.0	4.8	$1.1 \cdot 10^{-6}$
North Sumatra	2004	9.0	3.5	1.5	1.8	3.0	$8.6 \cdot 10^{-7}$

In Table 3 the observed aftershock areas Alaska (1964) and the North Sumatra (2004) earthquakes are compared with the results obtained with the use of Eq. 1. It can be concluded that the calculated and the observed aftershock areas show good agreement.

Table 4 shows seismic strain rates of some arbitrary selected significant and medium earthquakes. The two last examples are the most significant seismic events of the Pannonian basin (*Zsíros, 2000*). The case of the Dunaharaszti earthquake demonstrates that the estimation with equation (1) is in a good agreement with observations of the aftershock area is in case of medium size earthquakes too.

Both in the case of giant ($M_W \geq 9.0$) and smaller earthquakes listed in Table 4 it was supposed that the sum in the r.h.s. of equation (4) is determined by the most significant seismic event of the source area,. what is - as it was shown in *Varga et al., 2004* - a rather good estimation of the reality.

Table 4. Seismic strain rate values obtained for great and medium earthquakes.

Place	Year	Mw	M_0 1019 Nm	Aftershock area (calc.) 10^2 km^2	Aftershock area (obs.) 10^5 km^2	Volume $10^3 km^3$	dε/dt
Gujarat	2001	7.6	32.2	95.52		95.84	$5.0 \cdot 10^{-4}$
Kobe	1995	6.9	2.8	11.05		22.10	$3.3 \cdot 10^{-5}$
Boumerdes	2003	6.8	2.0	8.63		17.26	$2.8 \cdot 10^{-5}$
Komárom	1763	6.3	0.4	4.27		5.97	$1.2 \cdot 10^{-5}$
Dunaharaszti	2004	5.6	0.04	7.58	7.96	0.15	$2.3 \cdot 10^{-6}$

The seismic rates $\frac{d\varepsilon}{dt}$ obtained in case of great and medium seismic events are relatively big. It is probable that in these cases *Δt=250* year is unrealistic and a much bigger time-interval should be considered.

It can be concluded from calculation results of Tables 3 and 4 that the timei-nterval between the characteristic earthquakes of a seismic zone can be estimated on the basis of accurate geodetic observations. For this purpose however in case of GPS measurements 10^{-8} relative accuracy is needed on the base of 30-40 km what is not possible at present. In case of strainmeter observations this accuracy is waranted on the base less then 100 m, but the strain rate values of these records are suffering because of local tectonical, meteorological and hydrological influences. Nevertheless the strainmeter observation provide realistic estimates of $\frac{d\varepsilon}{dt}$ (*Latynina and Karamaleeva1978*, *Varga,1984, Mentes1994, 1995*).

5. Conclusions.

The study of strain rates of great and medium seismic events shows, because the unrealistically high $\frac{d\varepsilon}{dt}$ values, that probably in case of their source zones the typical return period of characteristic earthquakes should be well above *Δt=250year* (around *Δt=1000 year)*. For "giant" earthquakes ($M_W \geq 9.0$) (Table 3) the *Δt=250year* return period seems to be satisfactory.

Presently the prediction of return periods is based on statistical analysis of regional earthquake catalogues. In case of improvement of strain measurement techniques the return periods can be obtained with the use of these geodetic techniques too. For this purpose the accuracy of GPS measurements should be increased or/and the local influences on strainmeter observations should be excluded more accurately.

Acknowledgements.

This paper was completed in the frame of the project supported by the Hungarian Scientific Research Found OTKA (Project: K60394).

References

Hanks T.C., Kanamori H., 1979,A moment magnitude scale , J.Geophys.Res., 84, B5, 2348-2350.

Huang Y.,Saleur H., Sammis C., Somette D.,1998,Precursors,aftershocks,critically and self organised critically. Europhysics Letters,41,43-48.

Kasahara K.,1981,Earthquake mechanics, Cambridge University Press

Kostrov B.V., 1974,Seismic moment and energy of earthquakes and seismic flow of rock. Izv. Acad. Sci USSR, Phys. Solid Earth,1,23-40.

Latynina L.A., Karamaleeva R.M., 1978, Deformographic measurements. Nauka, Moscow

Mentes Gy.,1994, Instruments for precise determination of horizontal deformations in the Pannonian basin, Acta Geod. Geoph. Hung.,29, 1-2, 161-177.

Mentes Gy., 1995, In-situ calibration of quartz tube extensometers, Marees Terrestres Bulletin d'Informations,121, 9070-9075.

Varga, P., 1984, Long-term variations recorded by extensometers. Journal of Geophysics, 55, 68-70.

Zsíros T., 2000,Seismicity and seismic risk of the Hungarian Basin: The Hungarian earthquake catalogue. Budapest, GGKI