Seismological monitoring of the deformation process

Article *in* Geology of the Pacific Ocean · January 2001

V. G. Bykov
Institute of Tectonics and Geophysics, Far Eastern Branch of the Russian Academy of Sciences

Geol. of Pac. Ocean, 2001, Vol. 16, pp. 981-994
Reprints available directly from the Publisher
Photocopying permitted by license only

Seismological Monitoring of the Deformation Process

V.A. BORMOTOV and V.G. BYKOV

Institute of Tectonics and Geophysics, Far East Branch, Russian Academy of Sciences, Khabarovsk

Received May 24, 1999

A technique for determining volume deformation variations is suggested and substantiated theoretically on the basis of the multiple-reflection longitudinal waves specified in the seismograms from distant earthquakes. Results of seismological monitoring carried out in the south of Priamurye testify to the qualitative and quantitative difference of the deformation process at different depth levels. Variations of volume deformation are estimated down to a depth of 50 km.

INTRODUCTION

The crust and upper mantle rocks undergo continuous changes of the stress-strain state related to the geodynamic processes, moon tides and technogenic effects. Significant changes are expected in the volumes of the geological medium with a relatively high porosity characterized by parameters such as density of fractures, fluid saturation, filling with gaseous and liquid phases. It is possible to estimate the stress-strain state changes of the medium at great depths employing the seismic methods, because seismic wave velocities depend on the factors mentioned above, which are associated immediately with the stress changes in the medium.

The stress-strain state of the medium is monitored by continuous geodetic measurements, observations of temporal variations of traveltimes or velocities of seismic waves from the artificial (explosions, pneumoradiators, vibrators) and natural (earthquakes) sources. For the study of the deformation process at depths exceeding 10 km the most economical and и ecologically pure are the methods using the waves from earthquakes, for example, the method of converted waves of distant earthquakes, (MOVZ). Exploration of MOVZ allows observation of temporal changes of the conversion boundaries on the seismic sections constructed for different regions [1-3] and gives a qualitative explanation of this change by transmission of the deformation waves [1-2]. However, quantitative estimates of the volume deformation variations have not been made, which was related, probably, to the significant restrictions in determinations of temporal variations of velocity of the converted waves. Though, in addition to the converted waves, the multiple-reflected waves from earthquakes are also used successfully in MOVZ [6].

This paper suggests and substantiates the method of determining volume deformation variations using multiple-reflected compressional waves. Results of monitoring of the deformation process based on temporal variations of the relative changes of the calculated reflector depth are shown. Variations of volume deformation are evaluated first down to a depth of 50 km using the natural source of waves.

EQUIPMENT, OBSERVATION CONDITIONS AND PROCESSING TECHNIQUE

Seismological monitoring has been carried out in the south of Priamurye employing MOVZ in 1996 through 1997 using multiple-reflected waves from distant earthquakes to determine relative variations of the compressional wave velocities δV_p in the crust and upper mantle.

The observations have been carried out using the apparatus complex device "Cherepakha". The recording station was installed at the observation point with the coordinates of 47.6°N latitude and 134.7°E longitude.

The package program MOVZBOL was used for data digitization, vizualization, processing, and interpretation [4]. The earthquakes with simple pulse form of the transmitting P wave and the presence of the pulse oscillations with lower amplitude in further arrivals were selected for data processing. The identity of the seismic channels was improved by correction of the amplitude-frequency and phase characteristics using the pulse calibration records made in 8 hr time interval while registering. The plot of processing of the corrected seismograms included the frequency-band filtration (1-2,5 Hz) and polargram computation with further linear and nonlinear (acutely directed) projection. Further, the form of the P-pulse was determined, and the amplitudes and times of the pulses whose forms coincide with that for the P wave were computed for the components. The angle of wave emergence in the space was determined from the amplitudes and arrivals suggested from linear polarization. We identified the waves with the angle deviation from direction of the P wave to 10-25° as the muptiple-reflected waves. Construction of the deep section was made with the account of seismic displacement using the one-dimensional velocity law.

RELATIONSHIP OF RELATIVE P-WAVE VELOCITY VARIATIONS AND CALCULATED DEPTH OF BOUNDARIES

Volume deformation variations δe of the medium and relative P-wave velocity variations $\frac{\Delta V_p}{V_p}$ are unified by a relation [13]

$$\delta e = \frac{1}{\chi}\frac{\Delta V_p}{V_p}, \tag{1}$$

where χ is the coefficient of nonlinearity of the geomaterials of the Earth's crust.

To observe deformation variations the accuracy in determination of velocity variations should be, at least, no less than 5%. MOVZ does not allow us to determine the absolute velocity values V_p with high accuracy, but it is quite sensitive to the velocity variations, which is displayed in temporal variations of the reflector depth $\overline{H}$ computed using a priori given constant velocity law. Therefore, it is necessary to modify relation (1) in such a way, that MOVZ data could be employed to calculate variations of deformations δe. For this purpose, let us consider the ray pattern (Fig. 1) of compressional wave propagation and the relationship between velocity variations and the calculated reflector depth that follows from this.

In the initial part of the seismograms (first tens of seconds) from distant earthquakes after arrival of the transmitting P wave whose wavefront is plane, the regular multiple-reflected compressional waves (PPP) from the boundaries of the lithosphere are observed (see Fig. 1). The PPP wave arrives to the

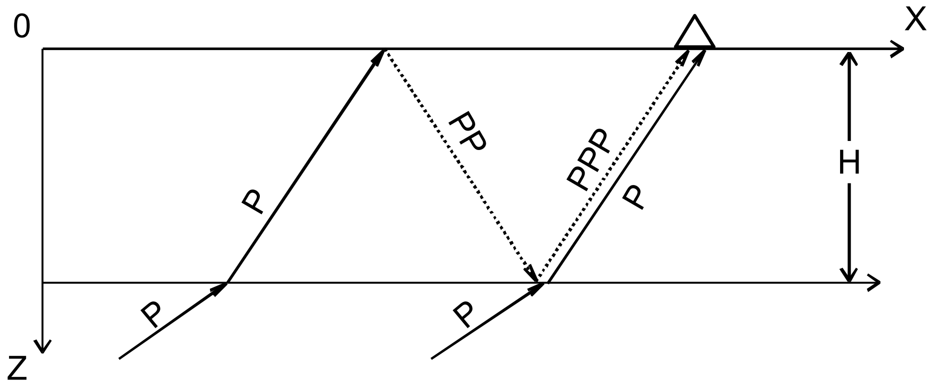

Fig. 1 Ray pattern for compressional wave propagation.

observation point following P wave after two reflections (from the Earth's surface and the boundary) with time delay

$$\Delta t_{ppp} = \frac{2H}{V_p}\sqrt{1-\frac{V_p^2}{V_a^2}}\,, \tag{2}$$

where H is the depth of the horizontal boundary, V_p is the effective velocity of the compressional wave to the boundary (average velocity for the layer), V_a is the apparent velocity of propagation of the compressional wavefronts along the boundaries, whose value is determined by the velocity structure of the medium when P wave travels from the source to the boundary.

The apparent P-wave velocity V_a from distant earthquakes is determined using the Jeffreys-Bullen curves [8]. The accuracy of the calculated apparent velocity $\overline{V}_a$ using the Jeffreys-Bullen curves is equal to 2-3 % for the epicentral distances in the range of 30-60°. For the removal distances of 63-102° the error does not exceed 5%, and for the values lower than 30° the error is significantly higher.

In the case the apparent velocities $\overline{V}_{a1}$ and $\overline{V}_{a2}$ are known for two earthquakes close in time of occurrence (V_p and H remain constant), then at $\overline{V}_{a2} > \overline{V}_{a1}$ from (2) follows

$$V_p = \left[\frac{\Delta t^2_{ppp2} - \Delta t^2_{ppp1}}{\Delta t^2_{ppp2}/\overline{V}^2_{a1} - \Delta t^2_{ppp1}/\overline{V}^2_{a2}}\right]^{1/2}. \tag{3}$$

The accuracy of calculation V_p depends on the accuracy of determination $\overline{V}_a$ of time delay Δt_{ppp} and for the relatively low-frequency P waves (f~1Hz) at depths lower than 40 km (Table), it is not higher than 10%, which is not sufficient enough to reveal V_p variations in the Earth's crust. But in the case the data are available on a set of earthquakes with different epicentral distances, it is possible to more accurately determine the velocities due to statistical effect using equation (2) in quadratic coordinates:

$$\Delta t^2_{ppp} = \frac{4H^2}{V_p^2} - \frac{4H^2}{\overline{V}_a^2}. \tag{4}$$

In these coordinates the results of seismic sounding of the lithosphere are represented as a series of sections of straight lines, whose angle is equal to

$-4H^2$, and the coordinate of the point of intersection with the axis $\frac{1}{\overline{V}_a^2}$ is equal to $\frac{1}{\overline{V}_p^2}$. If to determine the increment Δt^2_{ppp} as

$$\Delta\left(\Delta t^2_{ppp}\right) = \frac{8H^2}{\overline{V}_a^2}\left(\frac{\Delta V_a}{\overline{V}_a}\right) - \frac{8H^2}{V_p^2}\left(\frac{\Delta V_p}{V_p}\right), \tag{5}$$

it can be seen that the influence of velocity variations of V_p and V_a is different. Because the value V_a exceeds that of V_p by a factor of 2 or 3 (for a range of epicentral distances of 30-60°), the variations and errors of V_a influence the increments Δt^2_{ppp} 4-9 times weaker than those for $V_{p.}$ This allows us to obtain the required accuracy in determination of relative velocity variations of V_p for any set of earthquakes using the method of partial compensation of the influence of V_a.

Let us demonstrate that variations $\overline{\Delta t}_{ppp}$ for two seismic events 1 and 2 with a large time interval of occurrence are associated with variation of V_p. It is possible to eliminate the dependence of time delay on the epicentral distance dividing it by kinematic correction calculated for a priori value of $\overline{V}_p$ and for the values of $\overline{V}_{a1}, \overline{V}_{a2}$ using the Jeffreys-Bullen curves. Relation for these two time delays is as follows:

$$\frac{\overline{\Delta t}_{ppp2}}{\overline{\Delta t}_{ppp1}} = \frac{V_{p1}}{V_{p2}}\frac{H_2}{H_1}\frac{\left[1-\left(\frac{V_{p2}}{V_{a2}}\right)^2\right]^{1/2}}{\left[1-\left(\frac{V_{p1}}{V_{a1}}\right)^2\right]^{1/2}}\cdot\frac{\left[1-\left(\frac{\overline{V}_p}{\overline{V}_{a1}}\right)^2\right]^{1/2}}{\left[1-\left(\frac{\overline{V}_p}{\overline{V}_{a2}}\right)^2\right]^{1/2}}, \tag{6}$$

where indexes 1 и 2 are related to the medium parameters for different seismic events. Taking into consideration the fact that depth change is lower than the velocity variation by a factor of 10^4-10^3 [12] and $\frac{\overline{V}_p}{\overline{V}_a} \approx \frac{V_p}{V_a}$, from (6) we obtain:

$$\frac{\Delta t_{ppp2}}{\Delta t_{ppp1}} \cong \frac{V_{p1}}{V_{p2}} \quad \text{or} \quad \frac{\Delta t_{ppp2}}{\Delta t_{ppp1}} - 1 \cong \frac{V_{p1}}{V_{p2}} - 1\,. \tag{7}$$

Thus, (7) establishes a relationship between the relative velocity variations V_p and temporal traveltime variations

$$\frac{\Delta V_p}{V_p} \cong -\frac{\Delta(\Delta t_{ppp})}{\Delta t_{ppp}} . \tag{8}$$

For similar $\overline{V}_p$ and different $\overline{V}_a$ relations of the two values of calculated depths look like

$$\frac{\overline{H}_2}{\overline{H}_1} = \frac{\Delta t_{ppp2}}{\Delta t_{ppp1}} \cdot \frac{\left[1-\left(\dfrac{\overline{V}_p}{\overline{V}_{a1}}\right)^2\right]^{1/2}}{\left[1-\left(\dfrac{\overline{V}_p}{\overline{V}_{a2}}\right)^2\right]^{1/2}} . \tag{9}$$

Substituting (2) in (9), we obtain the approximate expression

$$\frac{\Delta V_p}{V_p} \cong -\frac{\Delta(\Delta t_{ppp})}{\Delta t_{ppp}} \cong -\frac{\Delta\overline{H}}{\overline{H}} = \frac{H_2}{H_1} - 1 . \tag{10}$$

From (10) it follows that relative P-wave velocity variations coincide approximately with relative changes of the calculated depth.

To estimate the corrections the model calculations have been made on the basis of rounded differences in the traveltimes Δt_{ppp} to 0.1 s, with the added corrections ±0.05 s for their different combinations (see the Table). The calculations show that the required accuracy of determination (about 5%) of the relative variations for effective velocities can be obtained for depths greater than 10 km. This refers to determination of the layer velocity variations $V_{lay.}$ in the layers with the thickness h:

$$\Delta t_{ppp} \cong \sum_{i=1}^{n} \frac{2h_i}{V_{p_{lay}i}} \left[1-\left(\frac{V_{p_{пл}i}}{V_a}\right)^2\right]^{1/2} = \sum_{i=1}^{n} \Delta t_{ppp_{lay}i} . \tag{11}$$

TEMPORAL VARIATIONS OF VOLUME DEFORMATION FROM RESULTS OF CONTINUOUS MOVZ OBSERVATIONS

Continuous observations of the reflectors on deep MOVZ sections constructed in the south of Priamurye on the basis of the same velocity law show that the calculated depth of the reflectors changes and in some rare cases the areas are

Table Estimates of determination errors for velocity variations.

Accurate values				Calculated values		Determination error	
Depth, H (km)	Velocity, V_{p1} (km/s)	Velocity, V_{p2} (km/s)	Relation V_2/V_1	$\overline{V}_p$	$\overline{H}_1/\overline{H}_2$	V_{p1} % from equation (3)	$\Delta V_p/V_p$ %
5	5.9	5.4	0.9	5.5	0.97	100	6.5
5	5.1	5.6	1.1	5.5	1.16	-13	5.1
10	6.2	5.6	0.9	5.7	0.94	24	3.2
10	5.3	5.8	1.1	5.7	1.17	-11	5.9
20	6.5	5.9	0.9	6.0	0.92	26	1.4
20	5.6	6.1	1.1	6.0	1.14	-9	3.9
40	6.9	6.3	0.9	6.4	0.91	8	0.1
40	5.9	6.5	1.1	6.4	1.14	-5	3.5
80	7.3	6.7	0.9	6.8	0.89	-1	-2.3
80	6.3	6.9	1.1	6.8	1.15	4	4.4

not observed at all (Fig. 2). This is probably related to the boundaries that are not distinguished substantially in the medium and depend on the temporal changes of stress fields [9].

Reliable determination of the reflectors by multiple-reflected waves is affirmed by their construction using the converted waves (Fig. 2a). Behaviour of the reflectors in time is shown in Fig. 2b. The boundaries are shown as isolines of reflection and conversion point density constructed on the basis of the normal distribution law and with regard for the technique resolution.

Variations of the reflector depths allow us to estimate temporal changes of the relative variations of the average P-wave velocities $\frac{\Delta V_p}{V_p}$, determined from the seismograms for distant earthquakes and the associated volume deformation variations δe by the relation following from (1) and (10):

$$\delta e = -\frac{1}{\chi}\frac{\Delta V_p}{V_p} \approx \frac{1}{\chi}\left(\frac{H_2}{H_1} - 1\right). \tag{12}$$

The value δe characterizes approximately temporal variations of the volume deformation of the medium in the observation area; H_1, H_2 denote the same reflector depth on the depth section in different time intervals (Fig. 3). Positive values of δe show unconsolidation of the volume distinguished or increase of

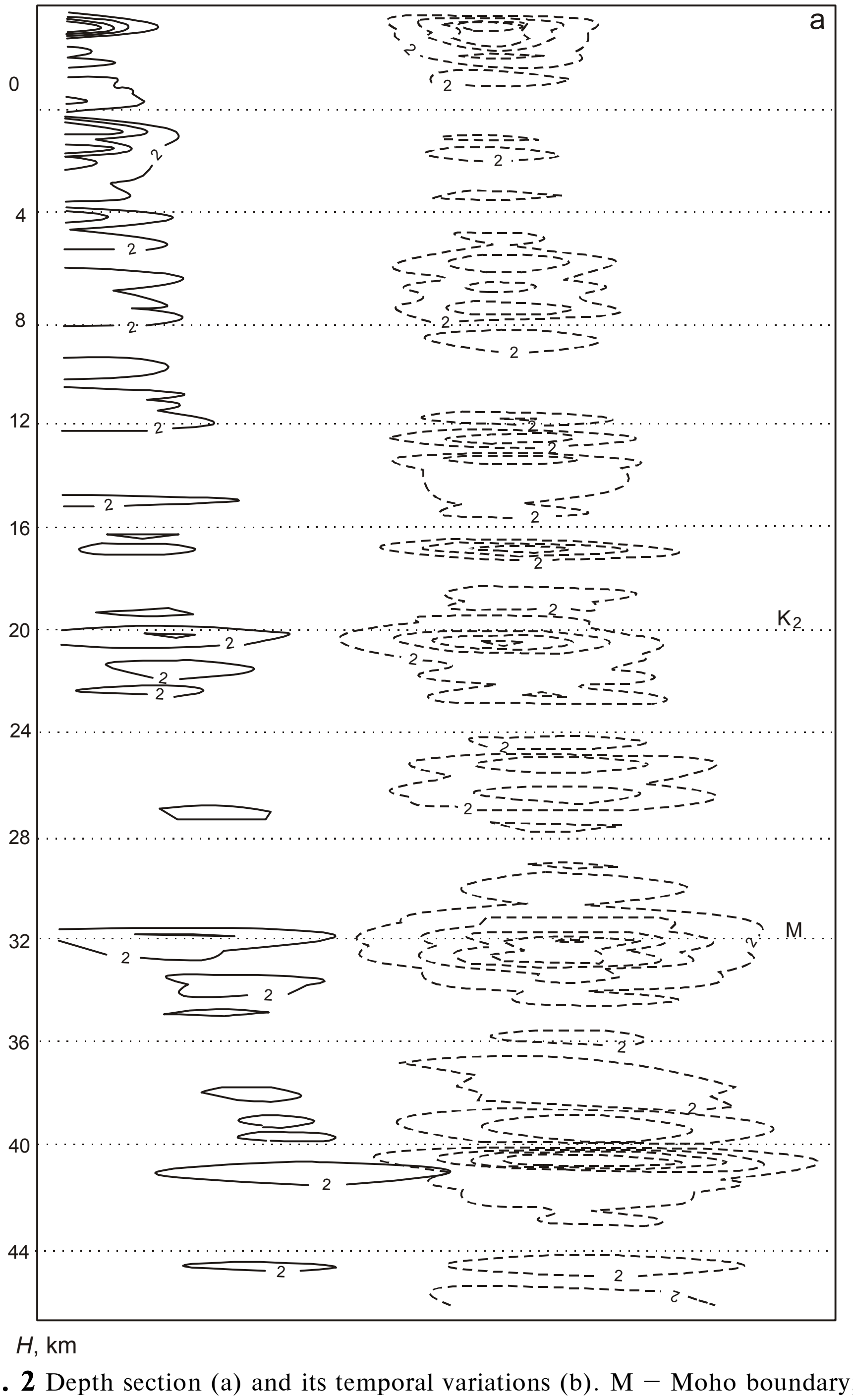

Fig. 2 Depth section (a) and its temporal variations (b). M − Moho boundary; K_2 − top of the high-velocity layer in the lower crust; 2 − number of the isoline chosen.

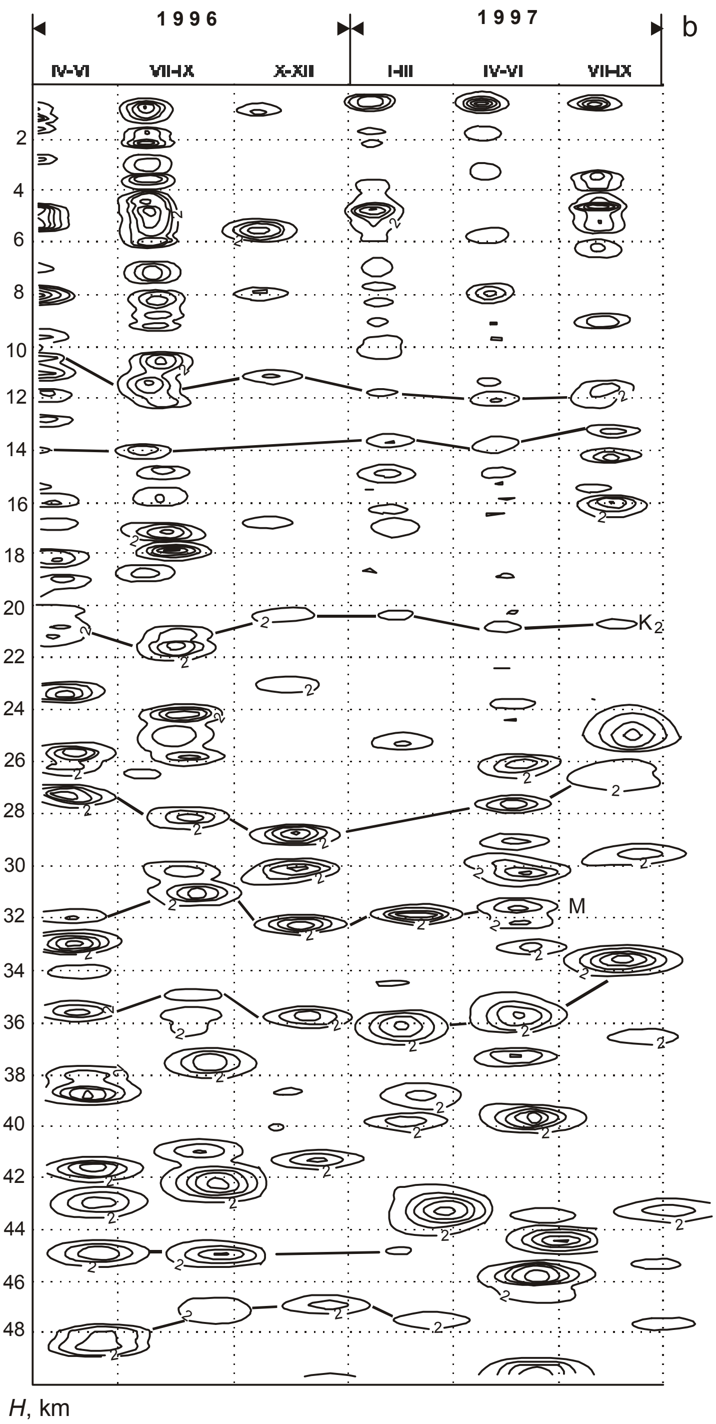

Fig. 2 (continued).

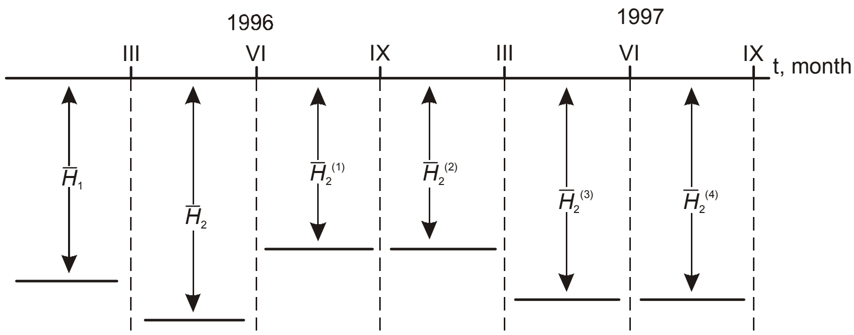

Fig. 3 Scheme of reflector migration at H = 20 km.

the fractured volume. This should lead to a decrease in the average values for V_p, which is displayed in the "downwarping" of the boundary to the depth $\overline{H}_2$. In the case of $\delta e<0$, this means the volume compression and leads to an increase in V_p, and, consequently, to the "upwarping" of the reflector from the level $\overline{H}_1$ to $\overline{H}_2$ $(\overline{H}_1 > \overline{H}_2)$. It is possible to estimate the stress field changes at known δe and the average value of the modulus of volume compression $\overline{M}$ which attains the value of the order of $5{\cdot}10^{10}$ Pa for the crustal rocks $\delta\sigma \cong \overline{M}\delta e$.

For the upper crust (10-15 km) the parameter $\chi \approx 10^3$ [13], and for the crust as a whole, the anomalously high values of the coefficient of nonlinearity are obtained $\chi \approx 3{\cdot}10^3 - 2{\cdot}10^5$ [5]. Good agreement of the experimental and calculated variations of compressional wave velocities is obtained at $\chi = 10^4$ for Kamchatka [7].

The difference in the reflector depths in the study area attains the value of the order of 1000-1300 m depending on the average reflector depth. Calculations by (12) show that this corresponds to the relative velocity variation $(2{,}0\text{-}11{,}0){\cdot}10^{-2}$ or the relative deformations $\delta e \approx (2{,}0\text{-}11{,}0){\cdot}10^{-5}$ at a value $\chi = 10^3$ (Fig. 4) and is agrees well with the results obtained by the other researchers.

Thus, estimates of temporal variations in volume deformation show $\delta e \approx 2{,}5{\cdot}10^{-5}$ for the area of the Caucasus Mineral Springs, $\delta e \approx 1{,}2{\cdot}10^{-5}$ for the Kopet-Dag area, and $\delta e \approx 10^{-6}$ for Central and Southern California [10]. Relative variations in the average P-wave velocities from local and distant earthquakes of the order of 10^{-2}-10^{-3} and those from the artificial sources $\sim 10^{-3}$ correspond to $\delta e \approx 10^{-5}$-10^{-6} The main reasons for this difference in variations $\frac{\Delta V_p}{V_p}$ are the earthquake epicenter migration and their magnitude variation [11]. Thus, using the approximate relation (12) is quite permittable for determination of the volume deformation variations.

Temporal changes of relative variations of the deformations observed for 8 depth levels (see Fig. 4) testify to the qualitative and quantitative differences in the deformation process occurring at different depths. The Figure demonstrates the results of the calculated values for the reflector boundaries that have been averaged every quarter. Of particular interest are the sectors of the counter-phase variation of the relative deformation of the medium, and also, the temporal changes of δe at a depth $H \approx$ 11-12 km, when gradual transition from a brittle shear to the pseudoplastic one followed by plastic deformation [14] takes place. The amplitude of the relative deformation δe increases with depth.

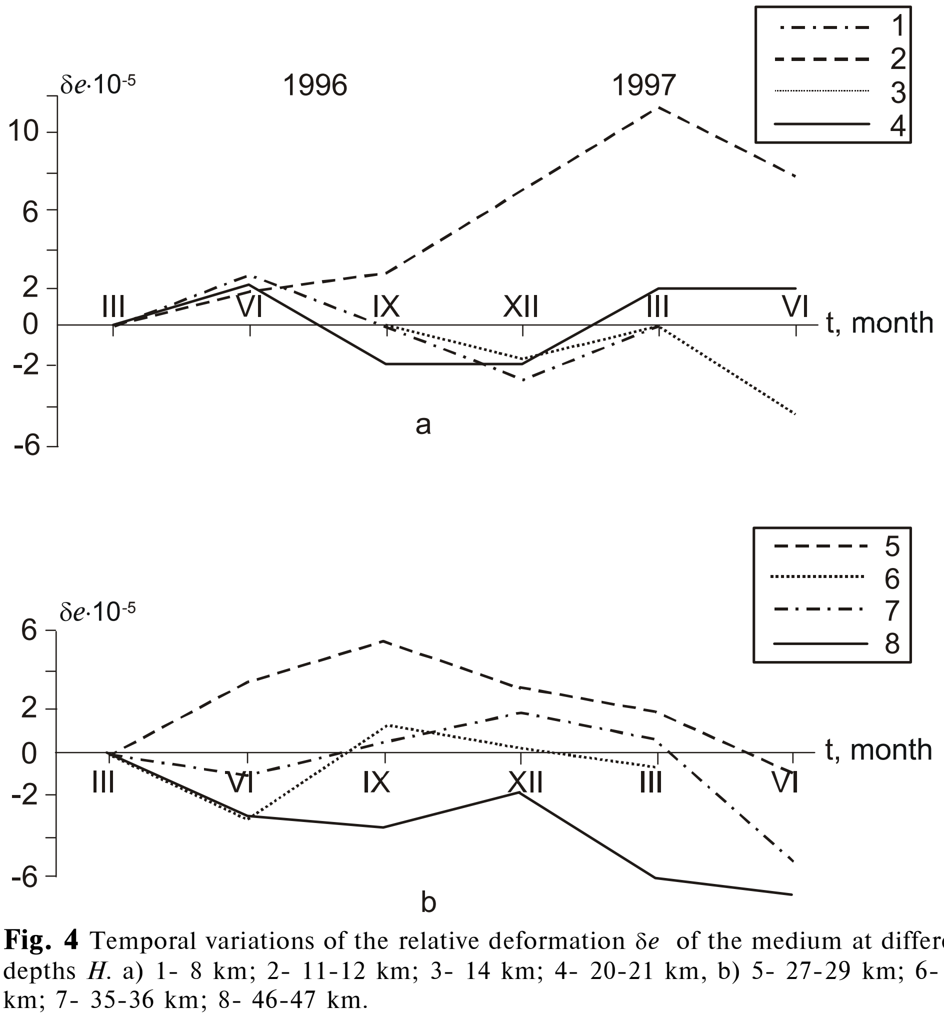

Fig. 4 Temporal variations of the relative deformation δe of the medium at different depths H. a) 1- 8 km; 2- 11-12 km; 3- 14 km; 4- 20-21 km, b) 5- 27-29 km; 6- 32 km; 7- 35-36 km; 8- 46-47 km.

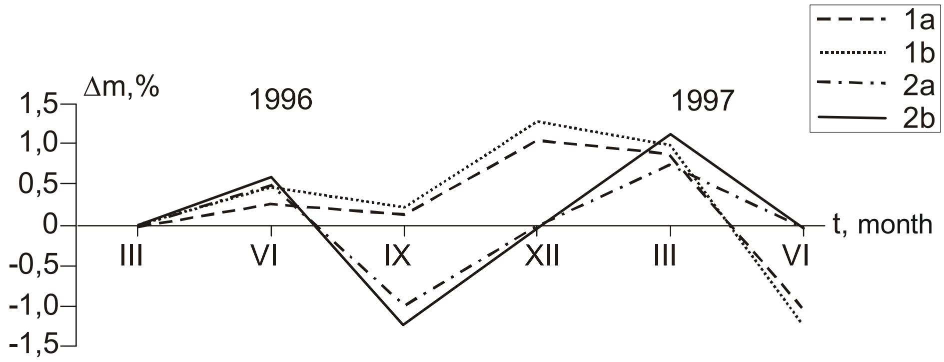

Fig. 5 Relative temporal changes of porosity Δm at different depths: 1- H=11 km; 2- H=21 km. a – dry regime; b – fluid-saturated regime.

It should be noted that the volume deformation variations have been previously measured using the seismic methods only from artificial sources and down to a depth of 10 km [11].

A change in the stress-strain state of the rock massives due to geodynamic processes occurring should result in the changes of the physical-mechanical parameters of the above massives, in particular, fracturing, elastic moduli and density. Calculations based on the Gassman-Domenico model [15] permitted determination (Fig. 5) of a change of porosity Δm of the geological medium from relative velocity variations, which, in turn, may be estimated from temporal variations of the reflector depths. The calculated values Δm in the fluid-saturated and dry regimes correlate with temporal relative volume deformation variations (see Figs 4 and 5). Thus, positive relative deformation corresponds to an increase in porosity ($\Delta m>0$) and, on the contrary, compression of the medium leads to the relative closure of fractures (pores) ($\Delta m<0$), that is, to decrease of the fractured volume that means “upwarping” of the reflector.

The deformation process occurring in the study area at different depths and in different time spans of observation is of different character (Fig. 6). A more stable situation, at least without a change in the phase of oscillation, is observed at a depth exceeding 28 km.

CONCLUSION

The results of the continuous MOVZ observations of the seismic reflector migration, estimates of temporal variations of the volume deformation and comparison with the data obtained employing other methods in different areas show that it is possible to use the technique suggested for monitoring of the

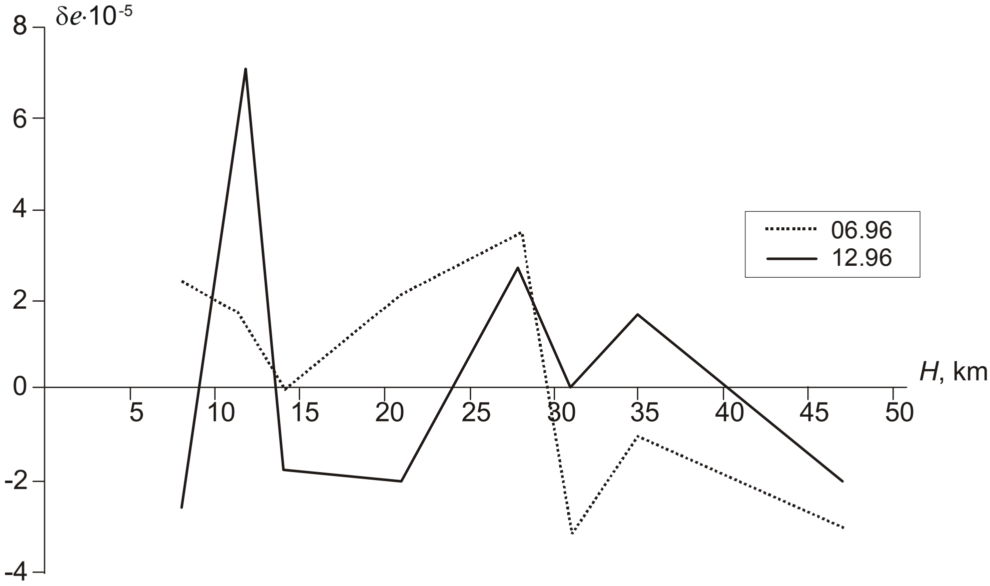

Fig. 6 Relative deformation variations δ*e* of the medium with depth *H* for different time spans of measurements: 1 — June 1996 yr.; 2 —Dec. 1996 yr.

deformation processes. It appears possible to obtain the most reliable data only for the effective velocity variations, starting from a depth of the order of 10 km.

Temporal variations of the deformation process are displayed most clearly at 10-20 km depth. Maybe, it is just for this reason that depths of the earthquake foci lie in the range of 5-15 km. Probably, the deeper horizons undergo changes of the regional stress fields, which are related as to the deformation wave transmission [1-2], so to the changes of the stress-strain state in period of strong earthquake preparation at large distances from the observation point [5]. Disappearance of the reflectors at some levels of the depth sections gives more evidence for relationship between the seismic reflector migration and changes in the stress-strain state of the medium.

It is possible to increase reliability of the data obtained on deformation variations employing comprehensive techniques for monitoring relative velocity variations of the compressional waves and determining the absolute velocity values using quadratic coordinates.

REFERENCES

1. Bazavluk, T. A. and Yudakhin, F. N., *Doklady Akademii Nauk* **329**, N 5: 565-570 (1993).

2. Bazavluk, T .A. and Yudakhin, F. N., *Doklady Akademii Nauk* **362**, N 1: 111-113 (1998).
3. Bezgodkov, V. A. and Chavushyan, R. E., in: *Prognoz zemletryaseniy* (Earthquake forecast) (Dushanbe: Donish, 1988), N 10: 193-205)
4. Bormotov, V. A. and Luneva, M. N., *Tikhookeanskaya geologiya*, N 2: 58-66 (1989).
5. Gamburtsev, A. G., *Seismicheskiy monitoring litosfery* (Seismic monitoring of the lithosphere) (Moscow: Nauka, 1992).
6. Dimakov, A. I., in *Prikladnaya geofizika, Vyp.* 65 (Applied Geophysics, issue 65) (Moscow, 1972). 43-52.
7. Dobrovolsky, I. P., *Teoriya podgotovki tektonicheskogo zemletryaseniya* (Theory of tectonic earthquake preparation) (Moscow: IFZ AN SSSR, 1991).
8. Kogan, S. D., *Izvestiya AN SSSR, ser. Fizika Zemli* N 6: 3-13 (1980).
9. Mints, M. V., Kolpakov, N. I., Lanev, V. S. and Rusanov, M. S., *Geotektonika* N 5: 62-72 (1987).
10. Nevsky, M. V., in *Dinamicheskiye protsessy v geofizicheskoi srede* (Dynamic processes in geophysical medium) (Moscow, 1994): 40-55.
11. Nevsky, M. V., Morozova, L. A., Starchenko, A. I and Bezgodkov, V A., in: *Problemy nelineinoi seismiki* (Problems of nonlinear seismics) (Moscow, 1987): 203-214.
12. Nevsky, M. V., Chulkov, A. B., Volosov, S. G. and Morozova, L. A., in: *Dinamicheskiye protsessy v geofizicheskoi srede* (Dynamic processes in geophysical medium) (Moscow, 1994): 75-94.
13. Nikolaev, A. V., in: *Dinamicheskiye protsessy v geofizicheskoi srede* (Dynamic processes in geophysical medium) (Moscow, 1994): 5-11.
14. Nikolaievskiy, V. N. and Sharov, V. I., *Fizika Zemli* N 1: 16-28 (1985).
15. Domenico, S.N,. *Geophysics* **41**, N 7: 882-894 (1976).

Translated by N.N.Kovriga