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Seismogeodynamic Migration Processes in the Central Caspian Sea and Adjacent Structures of the Caucasus and Kopet Dagh

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Abstract—The migration of seismogeodynamic activation is traced along the Caucasus–Caspian Sea–Kopet Dagh zone of transition from the Alpine fold system to a young epi-Paleozoic platform (the Scythian-Turan plate). A dynamic relation between earthquakes of the Greater Caucasus, the central Caspian Sea, and Kopet Dagh is established within the Caucasus–Kopet Dagh structure. Properties of aftershock processes developing after the two largest earthquakes in the central Caspian Sea are analyzed.

INTRODUCTION

The territory under study is a zone of transition from the Alpine Crimea–Caucasus–Kopet Dagh fold system to a young epi-Paleozoic platform, namely, the Scythian-Turan plate. This zone belongs to the northern part of the Iran–Caucasus–Anatolia region, one of the seismically most active parts of the extended Mediterranean–Himalayan Alpine belt. Seismicity of this region is caused by an intense geodynamic interaction between the African, Arabian, and Eurasian lithospheric plates.

The high seismic activity of the Crimea–Caucasus–Kopet Dagh structure, including the central Caspian Sea, is largely related to marginal seismogeodynamic interactions between mountain and platform structures. These interactions can give rise to migration processes associated with strain waves traveling along fault structures of various ranks.

The N–S strike of the Caspian basin, across the Caucasus and Kopet Dagh structures, emphasizes a contrasting pattern of regional tectonic movements. A peculiar feature of local seismogeodynamics is the fact that the central Caspian lithosphere experiences a significant subsidence, whereas the adjacent mountainous structures of the Caucasus and Kopet Dagh are intensely rising. Earthquake sources at depths of up to 100 km caused by continuing subduction of the lithosphere at the boundary between Alpine and epi-Paleozoic structures are evidence of these processes.

Studies of the deep structure, neotectonic and recent crustal movements, focal mechanisms of local earthquakes, spatial–temporal structure of seismicity, and migration processes of seismic activation in this region can provide more detailed constraints on the contemporary process of involvement of platform territories into geodynamic movements and this, in turn, opens new ways for seismic hazard assessment and long-term prediction of strong earthquakes in southern European Russia.

Results obtained previously by one of the authors from seismogeodynamic studies of the transition zone between the Tien Shan epi-platform orogen and the epi-Paleozoic Turan plate [Karzhauv and Ulomov, 1966; Ulomov, 1974] demonstrate that a realistic estimate of the seismic potential of the territory examined in the present paper can be obtained. Nearly ten years before the destructive Gazli earthquakes of 1976 (M = 7.0 and 7.3) and 1984 (M = 7.2), these studies predicted, with a high accuracy, a low seismicity area in the central Turan plate where these seismic events actually occurred (see Fig. 1).

The results presented below are a continuation of seismogeodynamic investigations of the lithosphere of the Caspian Sea and the entire Crimea–Caucasus–Kopet Dagh region [Ulomov, 1993b, 2003; Ulomov *et al.*, 1999, 2002].

SEISMICITY AND SEISMIC REGIME OF THE REGION

A general idea of the seismicity of the Caucasus–Kopet Dagh structure can be gained from a map showing epicenters of earthquakes (Fig. 1) from the Specialized Catalog of Northern Eurasia (http://socrates.wdcb.ru/scetac/) complemented by data up to 2002. The seismic potential of the territory studied is very high. The strongest earthquakes (M ≈ 8.0) occurred on the western and eastern coasts of the Caspian Sea (the Shemakha earthquake of 1668 and the Krasnovodsk earthquake of 1895). Fairly large seismic events that occurred in these regions in recent years are the Apsheron (Azerbaijan) M ≈ 6.4 earthquake of 2000 and the Balkhan (Turkmenistan) M ≈ 7.5 earthquake of 2000. (Here and below, the magnitude M stands for the magnitude M_s determined from surface waves.)

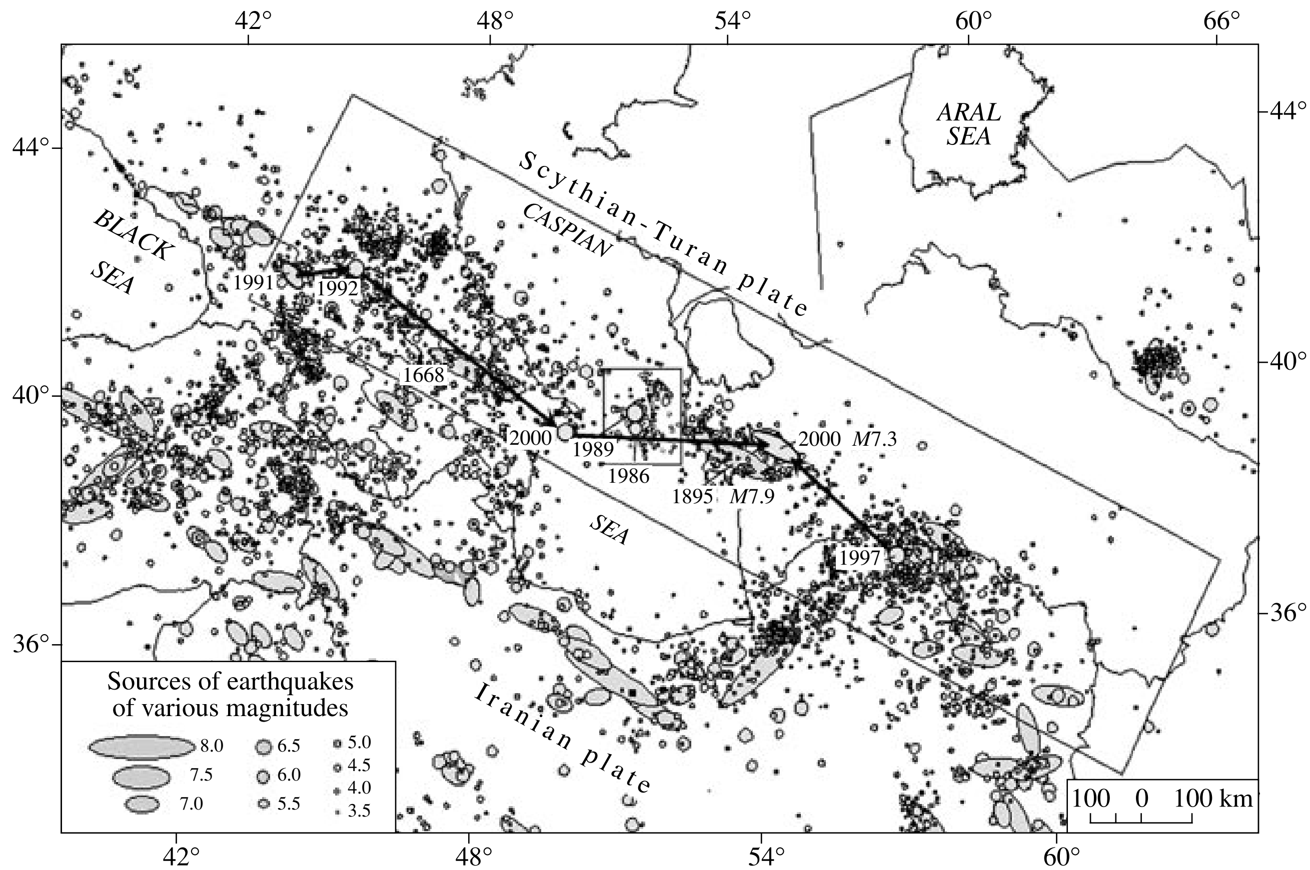

Fig. 1. Seismicity of the Caspian region [Ulomov, 2003]. The territory under study is bounded by a rectangle striking SE. The rectangle in the central part of the Caspian Sea is the area of the earthquakes of 1986 and 1989 and their aftershocks. The thick arrows show the direction of the source migration of $M \geq 6.3$ earthquakes within the study region in the period from 1991 through 2000. The numbers are the occurrence years and magnitudes of earthquakes. The source area of the three Gazli (Turan plate, western Uzbekistan) earthquakes of 1976 ($M = 7.0$ and 7.3) and 1984 ($M = 7.2$) can be seen on the right at 40°N. The sizes and orientations of $M \geq 6.8$ earthquake sources shown in this figure correspond to real values.

A characteristic feature of the Crimea–Caucasus–Kopet Dagh seismicity is the presence of deep earthquake sources related to subduction remnants of lithospheric plates [Ulomov, 2003]. The deepest sources reach a depth of 100 km and are confined to the Northern Caucasus and the most mobile area of the central Caspian seafloor. The majority of earthquake hypocenters concentrate, as in other regions, in the upper crust and the depth interval 10–15 km.

As was shown in [Ulomov, 2003], a very high degree of ordering inherent in the seismicity structure of the Caspian Sea region is due to the convergence of the Iranian and Scythian-Turan lithospheric plates (Fig. 1). The symmetry of arcuate and linear structures is noteworthy. One such linear structure extending along the Greater Caucasus, central Caspian Sea, and Kopet Dagh is the object of our study. It is located within a SE striking rectangle (Fig. 1). The seismically most active offshore structure, approximately striking along the 52°E meridian, is clearly recognizable at the center of the region studied. It is shown as a smaller rectangle and is clearly traceable as a chain of epicenters of local earthquakes and aftershocks of the two recent Caspian earthquakes of March 6, 1986, and September 16, 1989, which had significant magnitudes ($M = 6.1$ and 6.3, respectively). Their sources were confined to the intersection of the meridional (52°E) structure and the regional Caucasus–Kopet Dagh structure striking SE.

On the basis of estimates obtained for the representativeness of data on earthquakes of various magnitudes in the region studied, we constructed recurrence plots of the earthquakes for the entire territory of the study region and separately for the Caspian Sea (Figs. 2a, 2b). The same procedure was performed for the aftershock sequences of the two aforementioned Caspian earthquakes (Figs. 2c, 2d).

The recurrence plot slope of earthquakes in the Caspian Sea is seen to be smaller in absolute value ($b = -0.92$) compared to the entire Caucasus–Kopet Dagh zone ($b = -0.97$). Judging from these plots, the recurrence rate of seismic events in the entire study region averages 450 yr^{-1} for $M = 8.0 \pm 0.2$, 140 yr^{-1} for $M = 7.5 \pm 0.2$, and 50 yr^{-1} for $M = 7.0 \pm 0.2$ (Fig. 2a). Earth-

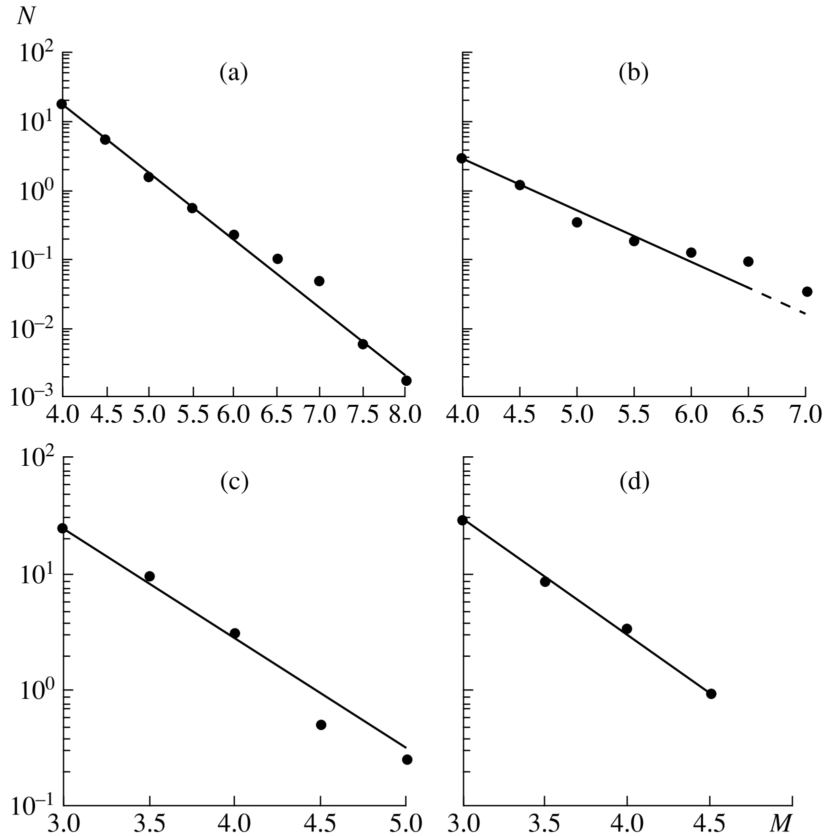

Fig. 2. Year-averaged recurrence plots of earthquakes constructed by the maximum likelihood method with the magnitude intervals $M \pm 0.2$ at a magnitude step of 0.5: (a) Caucasus–Kopet Dagh zone (the slope is $b = -0.97$); (b) Caspian Sea ($b = -0.92$); (c) aftershocks of the Caspian earthquake of 1986 over the period from March 5, 1986, to September 15, 1989 ($b = -0.96$); (d) aftershocks of the Caspian earthquake of 1989 over the period from 1989 through 1992 ($b = -1.03$).

quakes with $M = 7.0 \pm 0.2$ in the Caspian Sea can occur, on average, once every 140 years (Fig. 2b). Earthquakes of such magnitudes are estimated as "threshold" events capable of generating tsunami waves [Dotsenko *et al.*, 2003].

SPATIAL–TEMPORAL DISTRIBUTION AND MIGRATION OF EARTHQUAKE SOURCES ALONG THE CAUCASUS–KOPET DAGH GEOLOGIC STRUCTURE

Spatial–temporal migration of earthquake sources in the study region (see Fig. 1) is evident in Fig. 3. The next cycle of seismic activation along structures of the Greater Caucasus, central Caspian Sea, and Kopet Dagh has started after a nearly 20-year period of relative seismic quiescence that lasted from 1970, i.e., after two last earthquakes of $M = 6.6$ that occurred within the study region on both sides of the Caspian Sea. This activation was initiated by two earthquakes of 1986 ($M = 6.1$) and 1989 ($M = 6.3$), relatively large for the Caspian Sea area. They were almost immediately followed by two other strong events in the western part of the study region: the Racha earthquake of 1991 ($M = 6.9$) (such strong earthquakes had not occurred in this source area, at least, since 1880) and the Barisakho earthquake of 1992 ($M = 6.3$). These earthquakes seem to have initiated the current southeastward migration of seismogeodynamic activity along the entire Caucasus–Kopet Dagh zone. The activation involved the central Caspian Sea, where two fairly large earthquakes ($M = 6.4$ and 6.1) occurred near the Apsheron Peninsula on November 25, 2000, and the offshoots of the Greater Balkhan Ridge on the western coast of the Caspian Sea, where, nearly 100 years after the largest ($M = 7.9$) Krasnovodsk earthquake of 1895, a destructive ($M = 7.3$) earthquake occurred on December 6, 2000, comparable in magnitude with the notorious Ashkhabad catastrophe of 1948. The Balkhan earthquake was felt throughout Turkmenistan, the Caucasus, and Iran and was perceptible even in Moscow.

The source of the Balkhan earthquake was located about 90 km northwest of the epicenter of the 1895 Krasnovodsk earthquake. The Kazanjik earthquake ($M = 7.0$) of 1946 is the only known largest seismic event in this area over the last 120 years, and its magnitude was smaller than that of the Balkhan earthquake ($M = 7.3$).

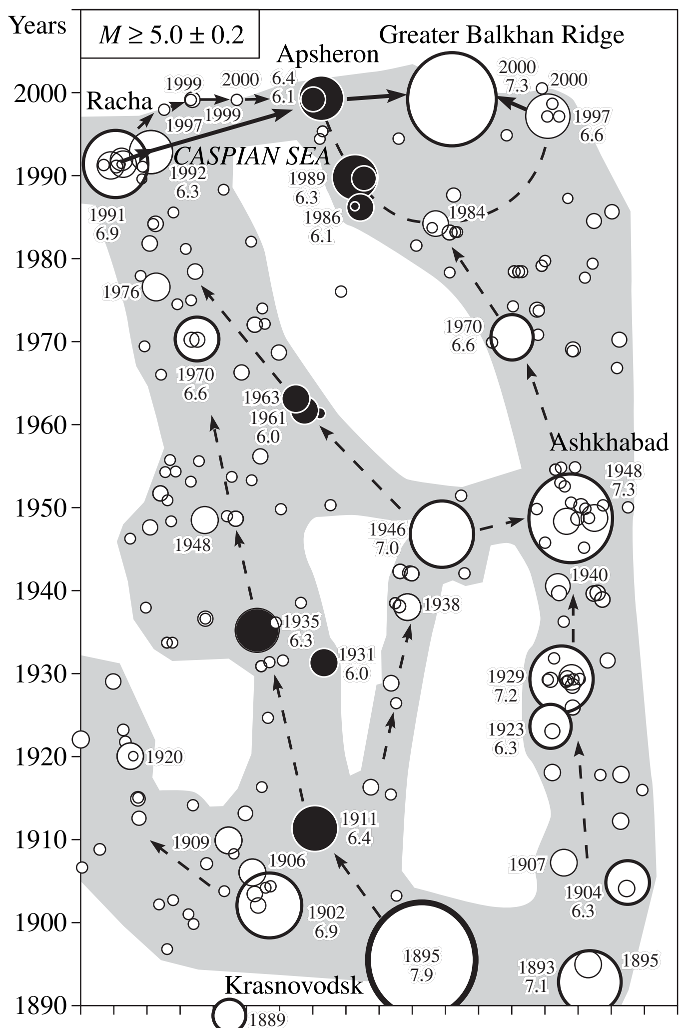

Fig. 3. Spatial–temporal distribution and migration of $M \geq 4.8$ earthquake sources in the study region (the Caucasus–Caspian Sea–Kopet Dagh geologic structure) in the period from 1889 through 2000. Diameters of the circles correspond to magnitudes in the range $4.8 \leq M \leq 7.9$. Sources of Caspian Sea earthquakes are shown as solid circles. The numbers are earthquake occurrence years and magnitudes. The shaded areas (or spatial–temporal channels after [Ulomov, 1993a, 1993b]) show clusterization of earthquake sources, and the arrows indicate the direction of seismogeodynamic migration. Other explanations are given in the text.

The migration of seismogeodynamic activation between the epicenters of the Racha and Apsheron earthquakes is also traceable from the consideration of weaker (M = 5.0–5.5) seismic events (short arrows in Fig. 3; this branch is not shown in Fig. 1). This chain of earthquakes includes events of 1997 (M = 5.0) and 1999 (M = 5.5, 5.1, 5.0, and 5.5). The last earthquake of M = 5.5 occurred some 20 km northwest of Makhachkala and was felt there at an MSK-64 intensity of up to 6.

Also noteworthy are migration features in the preparation area of the 2000 Balkhan earthquake, tentatively delineated by an arcuate broken line and resembling a typical seismic gap. Initially, a series of seismic events of M = 5.7–5.9 occurred in the epicentral area in 1983 and 1984. Afterward, the aforementioned Caspian and Apsheron earthquakes occurred west of the area, while an earthquake of M = 6.6 occurred in the eastern part of the study region, southwest of the epicentral area of the 1948, Ashkhabad earthquake. Another shock with M = 5.6 took place here on August 6, 2000.

A general inspection of the spatial–temporal development pattern of seismogeodynamic processes along the Caucasus–Caspian Sea–Kopet Dagh structure (Fig. 3) reveals other migration paths of seismogeodynamic activation that cross the structure in various directions independently of geologic distinctions between its marine and terrestrial areas.

DYNAMIC PARAMETERS OF CASPIAN EARTHQUAKE SOURCES

As noted above, the March 6, 1986 (M = 6.1), and September 16, 1989 (M = 6.3), Caspian earthquakes occurred at the intersection of the meridional (52°E)

Table 1. Parameters of focal mechanisms of Caspian earthquakes of 1986 and 1989 (according to [Balakina *et al.*, 1993])

Date	M	Axes of principal stresses						Nodal planes							
		T		N		P		$NP1$				$NP2$			
		Az°	Pl°	Az°	Pl°	Az°	Pl°	Str°	Dp°	Slip	Err°	Str°	Dp°	Slip	Err°
Mar. 6, 1986	6.1	40	33	310	0	220	57	310	78	–90	±15	130	12	–90	±15
Sept. 16, 1989	6.3	29	30	123	7	225	59	305	75	–83	±10	100	17	–114	±15
Sept. 17, 1989	6.2	0	15	270	0	180	75	270	60	–90	±15	90	30	–90	±15

Table 2. Dynamic source parameters of the Caspian, March 6, 1986, and September 16, 1989, earthquakes

Date	Time	M_0, 10^{18} N m	$\log E_1$, J	$\eta\sigma \times 10^5$, Pa	R_1, km	D_1, cm	R_2, km	D_2, cm	$\log E_2$, J	E_2/E_1
Mar. 6, 1986	00–05	5.0	15.2	200	29.0	10–15	–	–	–	–
Sept. 16, 1989	02–05	6.3	15.4	114	24.4	10.8	0.61	167	15.1	0.5
Sept. 17, 1989	00–53	3.2	15.1	103	20.3	10.3	0.36	139	14.7	0.4

structure and the regional Caucasus–Kopet Dagh structure striking SE (Fig. 1). Until these events, no earthquakes of $M \geq 4.8$ had been noted in their source area since at least 1880. The hypocentral depths of these Caspian earthquakes were estimated at 20 and 26 km, respectively. The stronger earthquake of 1989 was distinguished by its rather strong ($M = 6.2$) aftershock that occurred 30 km southeast of the main shock about a day later, which allows one to consider this earthquake as a double-shock event.

Their focal mechanisms are very similar in the faulting type (a steeply dipping normal slip, which confirms their subduction origin), the orientation of the principal axes of compressive and tensile stresses, and the orientation of both nodal planes. Focal parameters of these earthquakes and the strongest aftershock of the second event are presented in Table 1, where the following notation is used: P, T, and N, the principal axes of compressive, tensile, and intermediate stresses, respectively; $NP1$ and $NP2$, fault and auxiliary nodal planes, respectively; Az, strike azimuth (in degrees); Pl, dip angle of the axes T, N, and P; Str and Dp, strike and dip of the nodal planes, respectively; and Slip, direction of the slip on the nodal planes.

We should note that, although the focal mechanisms of the main earthquakes are generally similar, the focal mechanism of the repeated strong earthquake of September 17, 1989, also having a normal fault mechanism, has appreciably different parameters: its principal axes trend N–S, whereas its intermediate stress axis trends E–W; the nodal planes are rotated counterclockwise through either 35°–40° (plane 1) or 10°–40° (plane 2); its principal tensile and intermediate axes are horizontal, whereas the principal compressive axis is oriented nearly N–S (75°); the first nodal plane is inclined at a shallower angle, and the second nodal plane is steeply dipping.

It is of interest to compare dynamic parameters of all these earthquakes using their values estimated from analysis of frequency-selective spectra of records obtained at the Vannovskaya station (Δ = 600 km) (Table 2) [Rautian, 1988].

We should note that, judging from the spectral data, rupturing in all of the earthquake sources developed in two stages, but estimates were obtained only for parameters of the main source and the subsource of the September 16, 1989, earthquake and its September 17, 1989, aftershock. The spectra calculated for these events were used to estimate the radii of the entire source (R_1) and the subsource (R_2), the slip values in the source (D_1) and the subsource (D_2), and the subsource energy $\log E_2$. Dynamic parameters of the March 6, 1986, and September 16, 1989, earthquakes are found to be somewhat different. These parameters are the seismic energy E, the seismic moment M_0 (its value for the earlier event is 1.3 times smaller compared to the later one), the apparent stress $\eta\sigma$ (its value for the earlier event is 1.75 times greater compared to the later one), and the size of the main source (its average radius (R_1) for the earlier event is 1.2 times larger than for the later one). Notwithstanding these distinctions, the estimated slip values in the sources D_1 are comparable, and so are the sizes of the intensity-8 zones (30 × 16 km for the March 6, 1986, earthquake and 25 × 15 km for the September 16, 1989, earthquake; i.e., they differ by a factor of 1.3).

DYNAMICS OF AFTERSHOCK ACTIVITY

Figure 4 illustrates the spatial–temporal development of the seismic process in the region (39°–41°N, 50°–53°E) in the period from 1985 through 1994. It is of interest to note that the epicenters of major Caspian earthquakes lie in seismic gaps bounded by the foreshock activity (periods *1* and *5* in Fig. 4). The gaps virtually coincide in size with the source area of earthquakes of such strength. The 1989 earthquake was preceded by the January 1, 1989, foreshock with $M = 3.1$. The aftershock process can be traced up to September 16, 1989, for the earthquake of 1986 and from 1989 through 1995 for the 1989 earthquake.

According to Gardner and Knopoff [1974], spatial–temporal sizes of the area occupied by aftershocks average 54 km in diameter and about 510 days in time for earthquakes with $M = 6.0 \pm 0.2$ and 61 km in diameter and 790 days for $M = 6.5 \pm 0.2$. Judging from Figs. 4 and 5, these parameters for the Caspian earthquakes are different. Thus, aftershocks of the 1989 earthquake were observed until 1995, i.e., almost 2000 days after the main event, and the sizes of the aftershock areas estimated from their main cluster of the few first months after the main events were 80 and 85 km, respectively.

Due to the low determination accuracy of epicenters of weak earthquakes (and the majority of aftershocks are weak events), we examined the development of repeated shocks of Caspian earthquakes. The aftershock concentration areas of the earthquakes of 1986 and 1989 nearly coincide (*3* and *7* in Fig. 4). We should also note that the scatter in the aftershock epicenters of both earthquakes decreases significantly with an increase in aftershock magnitudes. The strongest repeated shock of September 17, 1989, did not affect the distribution of aftershocks of the September 16, 1989, earthquake. Its epicenter was located on the eastern periphery of the area of low magnitude aftershocks ($M = 2.5$–3.3), and stronger aftershocks were displaced to the west of this epicenter.

The spatial coincidence of the aftershock areas of the earthquakes of 1986 and 1989 can be considered as evidence of an inherited nature of the fracture associated with the later earthquake, which is corroborated by

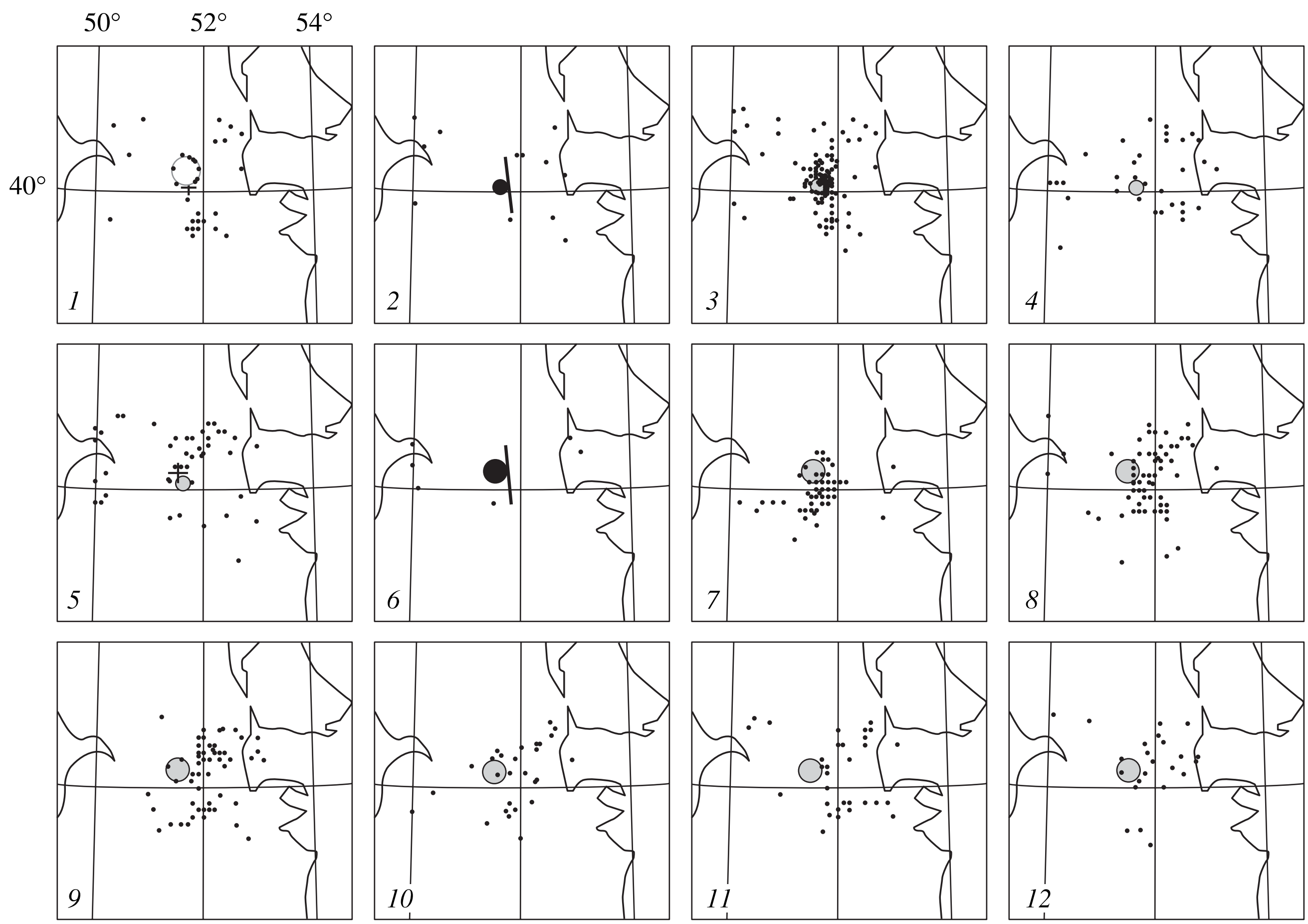

Fig. 4. Caspian earthquakes of 1986 and 1989 (solid circles) and the spatial–temporal distribution of $M \geq 2.3$ earthquakes within the area (39°–41°N, 50°–53°E) over the following periods: (*1*) 1985; (*2*) January 1, 1986, to March 6, 1986 (main event); (*3*) March 6, 1986 (without the main event) to December 31, 1986; (*4*) 1987; (*5*) 1988; (*6*) January 1, 1989, to September 16, 1989 (main event); (*7*) September 16, 1989 (without the main event) to December 31, 1989; (*8*) 1990; (*9*) 1991; (*10*) 1992; (*11*) 1993; (*12*) 1994. The predicted position of the Caspian earthquakes of 1986 and 1989 is shown by the crosses. The area of quiescence before the main seismic events is bounded by a broken line. The shaded circles indicate the positions of main events that occurred in the subsequent years. The Central Caspian fault (close to the 52°E meridian) is tentatively shown by a thick line (according to [Golinskii *et al.*, 1989]).

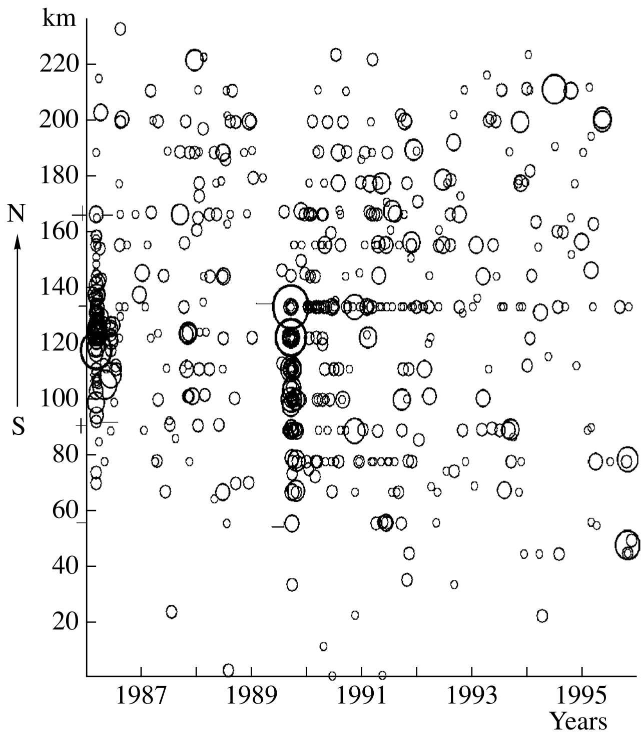

Fig. 5. The spatial–temporal distribution of aftershocks of the earthquakes of 1986 and 1989 projected onto the 52°E meridian. The crosses and dashes on the ordinate axis mark the sizes of areas estimated from aftershocks of the first months after the main events of 1986 and 1989, respectively.

significantly smaller values of apparent elastic stresses in its source (see Table 2). Probably, this accounts for a significant excess in the number of aftershocks in the first days after the earthquake of 1989 (according to [Golinskii *et al.*, 1989, 1993]).

Moreover, the total number of M = 2.2–2.4 aftershocks that occurred over 15 days (September 16–30, 1989; 1208 shocks) is 6.9 times greater than the number of 1986 earthquake aftershocks that occurred over 26 days (March 6–30, 1986; 175 shocks). The recurrence plots of the aftershocks (Figs. 2c, 2d) also yield evidence for these distinctions.

For the M = 2.2–2.4 aftershocks of the 1986 earthquake, $b = -0.86 \pm 0.05$ in the first few days and $b = -0.78 \pm 0.04$ in the subsequent 25 days. As regards the M = 2.2–2.4 aftershocks of the 1989 earthquake, $b = -1.01 \pm 0.04$ in the first few days and $b = -1.19 \pm 0.07$ in the subsequent 15 days. The significant distinctions in the recurrence plot slopes mean that, after the 1989 earthquake, stresses were mainly released via weak aftershocks.

Following [Ulomov and Ulomova, 1971; Ulomova *et al.*, 1979], we analyzed the time distribution of repeated shocks $M \geq 2.5$ of the Caspian earthquakes of 1986 and 1989 over the period from 1986 through 1995. Figure 6 illustrates regular trends in the decay of Caspian aftershock sequences. It is noteworthy that stagelike development of the aftershock process was also characteristic of the Tashkent, 1966, earthquake (M = 5.2) and the largest Gazli, 1976, earthquakes (M = 7.0 and 7.3) in western Uzbekistan.

Following the aforementioned works, we determined daily numbers of Caspian shocks during the first 100 days and afterward every 100 and 1000 days. In the latter two cases, the daily average was assigned to the center of the 100- and 1000-day intervals. The decay stages of the focal seismic process are fairly distinctly identified in the plots. The lines approximating the time dependences of the number of shocks $N = N(t)$ are also presented in Fig. 6.

A rather abrupt decrease in the number of aftershocks took place one day after the main shock of March 6, 1986; afterward, during the first ten days, the daily number of shocks decreased rather monotonically. The first stage is described by the equation $N(t) = 9.73t^{-0.620}$, $R^2 = 0.80$. A new burst of activity took place after about a month of quiescence and was distinguished by two nearly simultaneous strong (M = 4.3 and 4.7) aftershocks that caused weaker secondary aftershocks on the same day. The second stage was characterized by a more rapid decay described by the equation $N(t) = 72.16t^{-1.846}$, $R^2 = 0.77$. The decay rate was highest at the third stage, which terminated a month before the September 16, 1989, earthquake (M = 6.3) and is described by the equation $N(t) = 1E + 11t^{-4.064}$, $R^2 = 0.71$. No aftershocks with $M \geq 4.2$ were noted at this stage. At the end of the stage, complete quiescence lasted for nearly a month before the strong earthquake of 1989.

A characteristic feature of the first stage of the aftershock process following the Caspian, September 16, 1989, earthquake (M = 6.3) was the largest (M = 6.2) aftershock that occurred one day after this event. The repeated shocks of this day were more numerous than those noted immediately after the main shock. The earthquake of September 17, 1989, provoked a rapid stress release in the source area, after which the process began to decay less monotonically and, on the whole, more rapidly compared to the first 10 days after the 1986 earthquake; this decay was described by the equation $N(t) = 26.60t^{-1.066}$, $R^2 = 0.77$. A new activity burst was noted after a short period of quiescence (the second stage, $N(t) = 9.67t^{-0.499}$, $R^2 = 0.17$). Later, after nearly three months of complete quiescence, the aftershock process continued with the decay rate increasing at subsequent stages, which is possibly typical of the final period of the entire aftershock process (1992–1995). Note that earthquakes with M = 5 occurred in 1994 and 1995 outside the source area considered (Fig. 5). The spatial development of the seismic process, involving new areas, initiates the occurrence of rather numerous moderate earthquakes.

Figure 7 shows the depth distribution of aftershocks in the source zones of the Caspian earthquakes of 1986

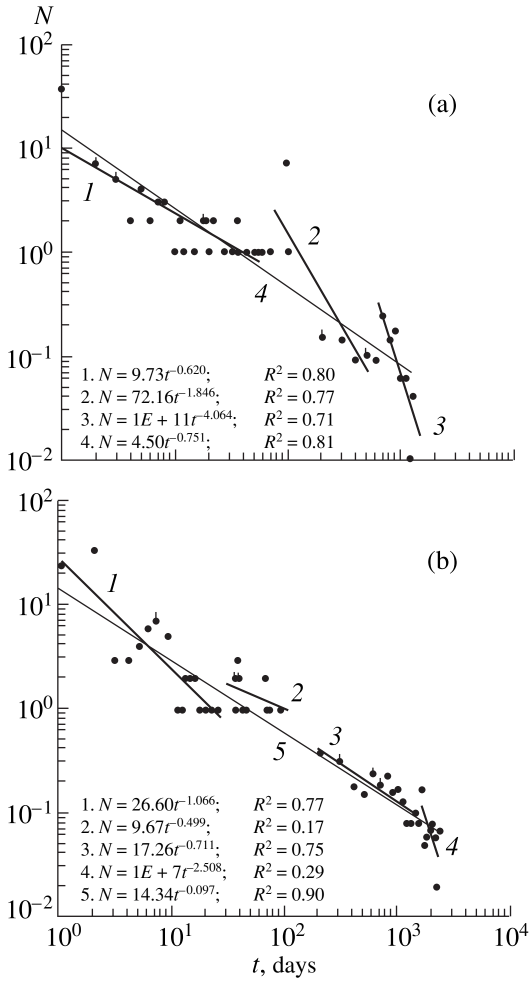

Fig. 6. Variations in the number (N) of repeated $M \geq 2.5$ shocks with time t after the main earthquakes: (a) 1986; (b) 1989–1995. Equations describing a decrease in the number of aftershocks N with time t are presented for every stage and for the entire process (the thin line); R^2 is the correlation coefficient. The barred points include $M \geq 4.2$ aftershocks.

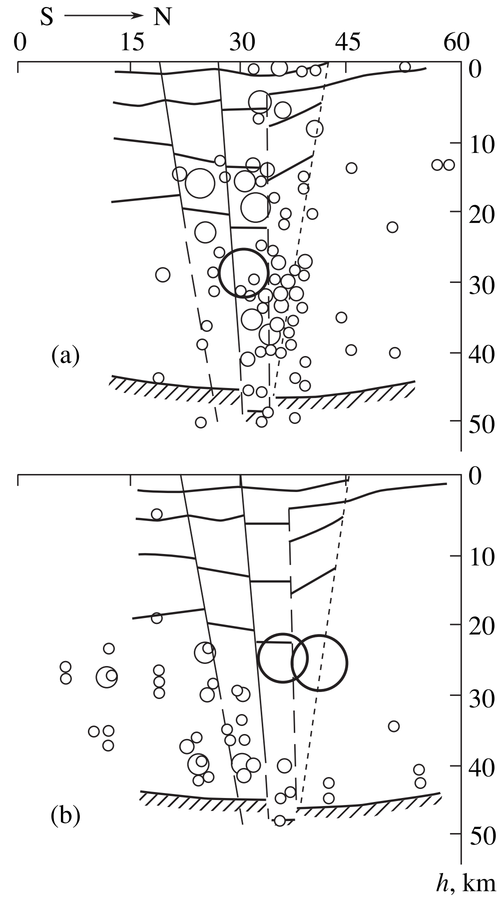

Fig. 7. Depth distributions of aftershocks of the earthquakes of (a) 1986 and (b) 1989 in the meridional cross section of the source zone. Sources of the main earthquakes and the strongest repeated shocks of the second event are indicated by large circles. The Moho is shown by a hatched line. Depths of the main events of 1989 are given according to macroseismic data.

and 1989. Crustal boundaries and sources of repeated shocks are shown according to [Golinskii *et al.*, 1989].

Nearly all aftershocks of the 1986 earthquake are seen to concentrate within the fractured zone of the crust, and the area of repeated shocks is obviously elongated in the vertical direction. The source of the main shock is located in the upper part of the basaltic layer, near its top. The strongest aftershocks are localized about 10 km above the main shock source and are confined to the metamorphic granitic layer. Weaker aftershocks migrate mainly downward in the fractured zone (down to the Moho) and, in a lesser degree, toward the Earth's surface. Thus, the stress release process of the 1986 earthquake is localized in space and encompasses the fractured zone of the crust, from its surface to its base.

Available data on aftershocks of the 1989 earthquake yield evidence of a stress release pattern basically differing from that of the 1986 events. Although, as noted above, their epicenters are close to each other, the depth distributions of their aftershock hypocenters are markedly different. The majority of aftershocks of the 1989 earthquake are localized in the lower crust and migrate southward toward crustal zones that were unaffected or weakly affected by the earthquake of 1986.

DISCUSSION

The spatial–temporal distribution and migration of seismic activity along the Greater Caucasus, central Caspian Sea, and Kopet Dagh, as well as the development of aftershock processes in the Caspian Sea after the two largest earthquakes of 1986 and 1989, indicate a structural and dynamic coherence of this region, including the zone of transition from Alpine structures of the northern Iran–Caucasus–Anatolia region to the young epi-Paleozoic Scythian-Turan platform. In other words, the development of seismic processes along the entire Caucasus–Kopet Dagh structure seems to “ignore” the presence of the Caspian offshore area, markedly differing in its geological and deep structure.

A more detailed study of the spatial–temporal Caspian aftershock activity has revealed certain regular patterns of the formation of aftershock sources. Aftershock epicenters of the 1986 earthquake are shown to group around the epicenter of the main shock, whereas the concentration area of aftershocks of the 1989 earthquake is displaced from its epicenter to the south by 30 km. Distinctions between the depth distributions of aftershocks of these events are more significant. Thus, the aftershock hypocenters of the first event are distributed vertically and encompass nearly the entire thickness of the crust. The strongest aftershocks are located above the source of the main shock, while weaker aftershocks migrate downward as deep as the Moho. Localization of aftershock hypocenters in the lowermost crust is typical of the 1989 earthquake. A noticeable migration of strong aftershocks toward a zone unaffected by the 1986 earthquake is observed.

The aftershock process of offshore Caspian earthquakes developed in separate stages, and this points to its similarity with processes forming source regions of continental earthquakes; the latter are primarily the strongest Gazli earthquakes of 1976 in the central Turan plate [Ulomova *et al.*, 1979] and the weaker Tashkent earthquake of 1966 in the zone of transition from the Tien Shan orogen to the Turan plate [Ulomov and Ulomova, 1971]. The aftershock activity of the 1986 earthquake is shown to have developed in three stages, each beginning with an abrupt increase in the number of aftershocks (by seven times at the second stage and by three times at the third stage). These bursts of activity were followed by its rapid decay. Note that, after the 1989 earthquake, whose source developed in a medium partially fractured by the 1986 earthquake, such intense bursts of aftershock activity were not observed at the third and fourth stages of the aftershock process. The general decrease in the number of repeated shocks with time was slower. However, as distinct from the 1986 earthquake, four, rather than three, stages were identified in the aftershock sequence of the 1989 earthquake. The features distinguishing the spatial–temporal development of aftershocks of the 1989 earthquake are corroborated by not only a greater number of shocks but also a significantly smaller value of elastic stresses released during the main event. These distinctions between the aftershock activity regimes of the Caspian earthquakes of 1986 and 1989 might be due to different strength characteristics of the medium (in the latter case, the medium was more fractured).

CONCLUSION

Investigations of the fine structure of seismicity, focal mechanisms of local earthquakes, neotectonic and recent crustal geodynamic movements in the zone of transition from the Crimea–Caucasus–Kopet Dagh fold system to the Scythian-Turan plate, migration processes of seismic activation, and specific features of seismic regimes open new opportunities for gaining more detailed constraints on the contemporary processes involving platform territories in geodynamic movements. The following will be beneficial to research in this area:

Acquisition of new information on the seismicity in the aforementioned transition zone. Updating of the instrumental earthquake catalog, with the inclusion of seismic events of very small magnitudes. Revision of the catalog of historic earthquakes and paleoearthquakes and evidence on their macroseismic effects.

Complementation of the electronic seismological database with other geologic and geophysical information with the wide use of modern technologies of geoinformation systems. Systematization of data on the deep structure and neotectonic and recent geodynamics of territories studied, including GPS data.

Studies of the fine structure and dynamics of regional seismicity and migration processes of seismic activation. More accurate delineation of earthquake source development zones and estimation of their seismogeodynamic parameters.

Development of an adequate seismogeodynamic model describing local seismicity effects within the zone of transition from the Crimea–Caucasus–Kopet Dagh fold system to the Scythian-Turan plate. Elaboration of methods for the identification of potential sources and the long-term prediction of strong earthquakes in the southern part of European Russia.

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REFERENCES

1. L. M. Balakina, A. I. Zakharova, A. G. Moskvina, and L. S. Chepkunas, “Focal Mechanisms of Strong Crustal Earthquakes in Northern Eurasia from 1927 through 1991,” in *Seismicity and Seismic Regionalization of Northern Eurasia* (IFZ RAN, Moscow, 1993), Vol. 1, pp. 123–131 [in Russian].
2. S. F. Dotsenko, I. P. Kuzin, B. V. Levin, and O. N. Solov’eva, “Possible Tsunami Implications of Seismic Sources in the Caspian Sea,” Fiz. Zemli, No. 4, 49–55 (2003) [Izvestiya, Phys. Solid Earth **39**, 308–314 (2003)].
3. J. K. Gardner and L. Knopoff, “Is the Sequence of Earthquakes in Southern California with Aftershocks

Removed Poissonian?" Bull. Seismol. Soc. Am. **64** (5), 1363–1367 (1974).

4. G. L. Golinskii, N. V. Kondorskaya, A. I. Zakharova, *et al.*, "The Caspian Earthquake of March 6, 1986," in *Earthquakes of 1986 in the USSR* (Nauka, Moscow, 1989), pp. 58–77 [in Russian].
5. G. L. Golinskii, Ch. M. Muradov, N. V. Petrova, *et al.*, "The Caspian Earthquake of September 16, 1989," in *Earthquakes of 1989 in the USSR* (Nauka, Moscow, 1993), pp. 44–61 [in Russian].
6. T. K. Karzhauv and V. I. Ulomov, "Evidence for Recent Tectonics and Seismicity in the Kyzyl Kum Area," Uzb. Geol. Zh., No. 3 (1966).
7. T. G. Rautian, "Determination and Interpretation of Earthquake Subsource Parameters," in *Engineering Seismology Problems* (Nauka, Moscow, 1988), Vol. 29, pp. 21–29 [in Russian].
8. V. I. Ulomov, *Dynamics of the North Asian Crust and Earthquake Prediction* (FAN, Tashkent, 1974) [in Russian].
9. V. I. Ulomov, "On Seismogeodynamics of the Transition Zone between the Tien Shan Orogen and the Turan Plate," in *Seismogeodynamics of the Transition Zone between the Tien Shan Orogen and the Turan Plate* (FAN, Tashkent, 1986), pp. 3–9 [in Russian].
10. V. I. Ulomov, "Seismogeodynamic Activation Waves and Long-Term Prediction of Earthquakes," Fiz. Zemli, No. 4, 43–53 (1993a).
11. V. I. Ulomov, "Global Ordering of Seismogeodynamic Structures and Some Aspects of Seismic Regionalization and Long-Term Prediction of Earthquakes," in *Seismicity and Seismic Regionalization of Northern Eurasia* (IFZ RAN, Moscow, 1993b), Vol. 1, pp. 24–44 [in Russian].
12. V. I. Ulomov, "Seismogeodynamics and Seismic Regionalization of Northern Eurasia," Vulkanol. Seismol, Nos. 4–5, 6–22 (1999).
13. V. I. Ulomov, "A Three-Dimensional Model of the Lithosphere Dynamics, Seismicity Structure, and Variations in the Caspian Sea Level," Fiz. Zemli, No. 5, 5–17 (2003) [Izvestiya, Phys. Solid Earth **39**, 353–364 (2003)].
14. V. I. Ulomov and N. V. Ulomova, "Formation of a Hypocentral Region and the Seismic Regime of Repeated Shocks," in *The Tashkent Earthquake of April 26, 1966* (FAN Uzb. SSR, Tashkent, 1971), pp. 122–138 [in Russian].
15. V. I. Ulomov, T. P. Polyakova, and N. S. Medvedeva, "Seismogeodynamics of the Caspian Sea Region," Fiz. Zemli, No. 12, 76–82 (1999) [Izvestiya, Phys. Solid Earth **35**, 1036–1042 (1999)].
16. V. I. Ulomov, T. P. Polyakova, and N. S. Medvedeva, "On the Long-Term Prediction of Strong Earthquakes in Central Asia and the Black Sea–Caspian Region," Fiz. Zemli, No. 4, 31–47 (2002) [Izvestiya, Phys. Solid Earth **38**, 276–290 (2002)].
17. N. V. Ulomova, R. P. Fadina, and T. P. Merkulova, "Regular Study of the Aftershock Regime of the Gazli Earthquakes in Relation to Earthquake Prediction," in *Seismological Investigations in Uzbekistan* (FAN, Tashkent, 1979), pp. 64–75 [in Russian].