AN OVERVIEW OF NEOTECTONIC STUDIES

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Translated from an original manuscript based on papers by the author published in various Soviet journals. Trifonov is with the Institute of the Lithosphere, USSR Academy of Sciences, Moscow. The paper reflects the broader Soviet usage of the term "neotectonics," encompassing activity back into the Neogene period. Trifonov presents a worldwide survey relating neotectonic activity, but the principal interest to non-Soviet geologists will probably be its use as an introduction to the literature on detailed investigations of current tectonic activity, particularly along fault zones. The following paper, by Fin'ko, is one such study, on a segment of the Indian-Eurasian plate boundary.

Three circumstances underlie the importance of neotectonics in present-day geologic science. First, Neogene-Quaternary, and particularly Holocene, tectonic movements at the surface can be studied, measured, and dated much more accurately than ones in the remoter past, because they are accessible and have not been distorted by subsequent processes. Second, the seismicity and the geophysical fields enable one to relate the surface neotectonics to current deep structures and transformations and thus to construct a three-dimensional model for the latest stage of tectonic development of the lithosphere. Third, one can compare Neogene-Quaternary, Holocene, and active tectonic movements in order to correlate the consequences of events in geologically averaged and real time scales and to identify the complicated and often extremely uneven course of events. Neotectonics thus provides a basis for testing tectonic concepts.

An important part has been played in plate tectonic theory by seismological and neotectonic data on deep structures, active-zone kinematics, paleomagnetic studies, and stratigraphic correlation of young deposits. The basic concepts of plate tectonics were formulated in the 1965-1970 period by J. T. Wilson, W. J. Morgan, X. L. Pichon, J. F. Dewey and J. M. Bird, W. R. Dickinson, B. A. Isacks, and others [58]. They used several sources of evidence. One of them was the observation that the lithosphere is divided into blocks: There are combinations of stable (slightly deformed) regions with mobile belts and zones of various orders, where faults and deformation are concentrated and one finds the highest gradients in the tectonic rates. Kheraskov [37, p. 391] considered that this feature is to be taken as the working of a general law of solid deformation at the earth's surface. This is most fully reflected in Peive's study [62, 63] on deep faults. Another source has empirical evidence on the considerable horizontal displacements in bodies, which had given rise to hypotheses of continental drift and related crustal deformations put forward by Taylor, Wegener, Argand, and Holmes, and which were supplemented by later research on the oceans, which led Hess and Dietz to formulate the spreading hypothesis, that oceanic crust expands in midocean rift zones and is subducted in island arcs and active continental margins. Decisive evidence for spreading has been provided by the Vine and Matthews interpretation of banded magnetic anomalies in the oceans.

In plate tectonics, the lithosphere, which includes the crust and the uppermost mantle, consists of several rigid undeformable plates, which move on (or along with) an upper-mantle layer of reduced viscosity, the asthenosphere, moving apart in some places and coming together in others. Where they move apart, the lithosphere expands, in particular in the oceanic crust, and where they collide, one of them is subducted beneath the other and descends into the mantle, which is accompanied by deformation at the plate edges and by magmatic and metamorphic transformations, which give rise to fold belts. The top of the descending plate is marked by a deep seismic focal zone, whose position and dip are related to regular variations in composition and metallogenic specialization in the eruptive rocks across the island-arc or active continental-margin zone, and the formation of paired low-temperature high-pressure and high-temperature metamorphic belts. The plate movements reflect mantle or at least upper mantle convection.

In the 15-20 years elapsed since the basic plate-tectonic concepts were formulated, new evidence has appeared that forces us to modify or supplement certain concepts. Here these are considered from the basis of neotectonic data. It is also necessary to establish how far these neotectonic structural and historical features are representative of earlier stages of earth history.

1. Neotectonic Research Methods

Neotectonics has four aspects: research on 1) the morphology, kinematics, and spatial relationships of the latest surface structures; 2) the current state and tectonic processes in the crust and upper mantle in relation to surface neotectonic formations, 3) Neogene-Quaternary structures and the correlation of neotectonic events in different regions, and 4) evaluation of deformations, displacements, and stresses, and analysis of the possible tectogenesis causes, i.e., topics combined under the name modern geodynamics. This wide range of topics requires various methods, not merely traditional geological ones, but also methods based in physics, mechanics, chemistry, geography, and history.

Research in morphology and kinematics is based primarily on traditional structural geology, but with one major supplement. The marker horizons are not exclusively or even mostly in stratigraphic sections but are rather the surface and existing relief forms. These are continually altered by exogenous factors, so it is an essential element to study those factors and the relief morphology and evolution for various landscapes, i.e., geomorphological research, where extensive use has been made of aerial and satellite imagery, which enable one to map various structures no matter what the landscape or accessibility. These images are particularly important in investigating spatial relationships and ranks for these structures. Recently, methods of identifying large-scale horizontal neotectonic displacements by the determination of magnetic declination and inclination of rock bodies have become increasingly important.

Holocene tectonogenesis is well preserved, particularly displacements on faults, which enable one to distinguish them from earlier neotectonic phenomena and to determine the horizontal and vertical movements. Points of reference are provided not only by Late Quaternary beds and relief forms but also by manmade features: ancient structures and irrigation systems, whose analysis requires not only geological and geomorphological evidence but also archaeological and historical data.

Ocean neotectonics is based on the bottom relief and deposits; the scope for observing these directly (e.g., from photographs) is limited, and seismic methods are decisive, among which multiple-beam echo sounding is promising, which enables one to determine morphology and areal occurrence more accurately than other methods. Investigation of magnetic anomalies retains its value, as supplemented and checked by dredging and deep-sea drilling.

Research on deep-level neotectonics and current tectonic processes within the lithosphere is based mainly on geophysical methods: gravimetric, geothermal, seismic surveying, and particularly seismological. The latter includes a new method of interpreting earthquake recurrence plots in terms of seismic viscosity, deformation, and stresses acting in seismic zones. Information on deep-level neotectonics is provided by the geochemistry and petrology of Neogene-Quaternary volcanism, as well as hydrochemical and gas and isotope analyses on underground fluids. Research on the physical states and transformations occurring at high pressures and temperatures sheds much light on processes in the lithosphere. Space pictures assist in evaluating these very varied data, as they often show indirect signs of deep structures in the present landscape, particularly in certain aspects of the relief. Such signs are also detected in geomorphology.

One uses biostratigraphic, lithologic, facies, and formation studies on young beds to correlate tectonic events in the recent past and to deduce the history of young structures; those methods may be supplemented with geomorphological ones, unconformity analysis, and changes in structure and deformation style. Biostratigtraphic and lithologic or facies methods are widely used also in research on older formations, and the specific feature for neotectonics lies merely in the scope for more detailed subdivision, wider territorial coverage, and correspondingly more comprehensive interpretation of the structure, as various methods of correlation are applicable only to the recent tectonic events. Particularly, one can correlate recent beds with synchronous relief forms such as terraces and peneplains. Specifically neotectonic methods include archaeology, thermoluminescence, and thermochronology, which enable one to date formations. Although paleomagnetic stratigraphy is now applied to Cretaceous and Paleogene beds, it remains reliable only for Neogene-Quaternary beds. Much increased detail has become available on the dates of Holocene movements, which has extended as far as recording individual seismic pulses. Statistical processing is applied to observed displacements in conjunction with radiocarbon, archaeological, and historical dates.

Neogene-Quaternary structure morphology can be determined accurately and in detail, as can the amplitudes and rates in the tectonic movements; one can determine the structure history, the seismic parameters, and other geophysical features, which makes it possible to evaluate quantitatively the distributions of displacements, deformations, and tectonic stresses at the surface and at various levels below it. This enables one to determine, at least hypothetically, the origins of those structures and to resolve local and global problems in neotectonic geodynamics by the use of mathematical and physical methods.

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Neotectonics thus uses many methods, and the range is more extensive than in any other branch of geology. Neotectonics demonstrates the necessity and effectiveness from applying advances in various sciences. Only such combination of approaches can assure progress in research on the earth, particularly the lithosphere.

2. Tectonic Features of the Present-Day Lithosphere

Two concepts are basic to the dynamic and kinematic aspects of plate tectonic theory: vertical homogeneity of plates in respect to lateral displacements or deformation and rigidity in conjunction with negligible deformability outside narrow zones of interaction. Recent neotectonic data cast doubt on these.

2.1. Neotectonic layering of the lithosphere

In 1967, when plate tectonics was being established, Peive [63] demonstrated the existence of tectonic layering, differential lateral displacements of lithospheric layers. He wrote [64, p. 7] that one can conclude that individual parts of the tectonosphere move differentially in the lateral direction, i.e., at different rates. If one assumes that the main tectonic flow and displacement zone is the asthenosphere in the upper mantle, we may take it that differential lateral displacements occur equally at the base of the crust and within it.

This idea was confirmed from research on thrust sheets, which arise from layering and lateral displacements at various levels in the crust and upper mantle [65, 95]. This layer motion and twisting is decisive in continental crust accretion. However, research on ophiolite complexes has shown that tectonic layering occurred in them before continental crust accretion, during the stage of spreading or formation of marginal seas. This is indicated by the marked disharmony between the dunite-harzburgite mantle assemblage and the overlying succession, or between gabbroids and dike complexes and basalts [95].

Neotectonic and seismological results were important to the foundation of plate tectonics, and they also contributed to models of layering. Neogene-Quaternary (N-Q) structures of various types occur at various levels in the lithosphere. One can examine their reflection in the relief and the subsurface crustal layering, and in N-Q volcanism, facies, and thicknesses, as well as in seismicity and in geophysical fields and on satellite imagery, from which one can distinguish the relative depths of the various structures and sometimes precisely in what depth range a structure is most active. Then one can compare the patterns of active N-Q structures at various levels. The differences in pattern indicate disharmony between deformation and faulting at various levels, which are signs of tectonic layering [49, 96].

The N-Q surface structures in the East Caucasus (Fig. 1) have longitudinal elements with west-northwest trends. These predominate in the map of

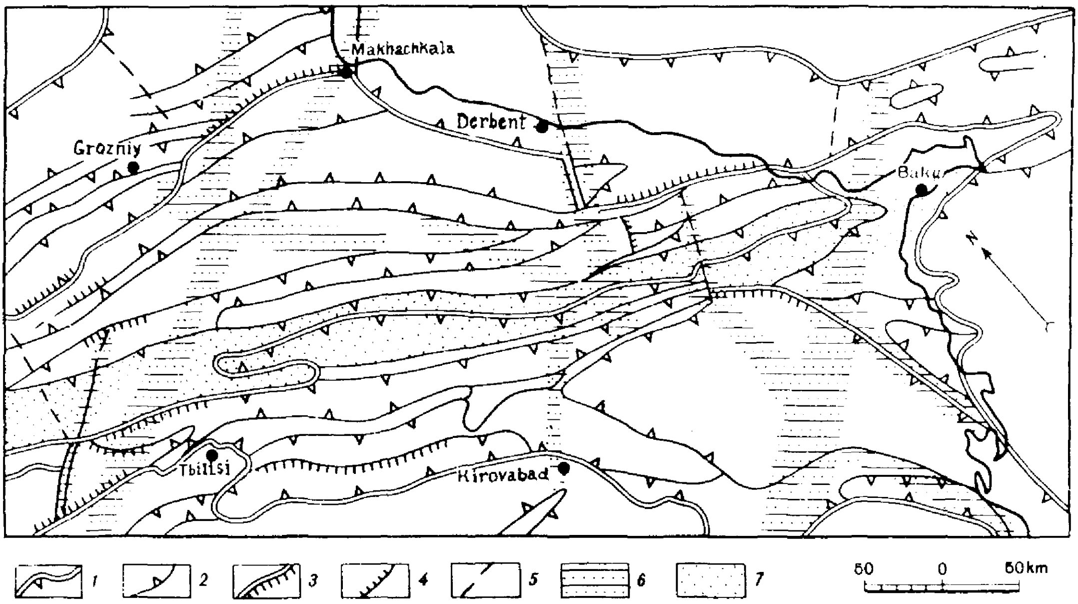


FIGURE 1. Neogene-Quaternary structures in the eastern Caucasus: 1-5) elements of the surface structure [54]: 1) boundaries of major neotectonic elements, 2) boundaries of secondary elements, 3) principal N-Q faults and flexures; 4) secondary faults and flexures (4), 5) N-Q faults having unknown displacement directions; 6 and 7) deep zones showing N-Q deformation and displacement [49, 96, 103]: 6) at 10-24 km, 7) at depths up to 60 km (southern flank of the High Caucasus).

the overall deformation subsequent to the Middle Miocene [54] and in the neotectonic structure and facies zoning. The surface less often shows linear neotectonic formations with other directions, which are not very prominent; many of them correspond to flexures having a long history, sometimes covered by alluvial deposits; there are also zones where the facies and thicknesses of the Neogene and Quaternary beds change, zones showing structural anomalies intersecting N-Q folds of the Caucasus trend, and boundaries between regions with various types and/or different directions of folding. The diagonal and transverse elements correspond to zones where there are considerable gradients in the gravity field, as well as to elements at the surface of the Pre-Jurassic crystalline basement in areas where that is deeply buried, and elements on the Conrad surface. Along these one often finds clusters of shallow-focus earthquake epicenters. Deep seismogenic dislocation zones are also similar in trend and location to the diagonal and transverse zones, because they are regions where seismic waves are anomalously damped, and which are probably identical in most cases with seismogenic faults. These element having north, northwest, and northeast trends reflect linear fault and dislocation zones, which are most active at depths of 10-20 km [49, 96].

The continuation of the southern High Caucasus flank has a particular structuring role near the Moho (Fig. 1), which dips steeply to the north; along it, one finds the deepest-focus earthquakes in the region, and a sharp change in crust thickness: from 50-60 km in the north to 40-45 km in the south.

There are three crustal levels in the East Caucasus having differing N-Q element sets and orientations: the sediment cover, in places together with the upper part of the crystalline crust; deeper crustal horizons; and the lower part of the crust and possibly the very uppermost mantle. This vertical disharmony may reflect differences in reaction to the transverse Caucasus compression [103]. Neotectonic volcanic structures having a deeper (mantle) source are indicated by the submeridional volcanic chains and the faults most fully represented in the central segment of the Caucasus and which continue southward to Lake Van. They are discordant to all the Caucasus N-Q crustal elements and suggest an essentially different setting in the upper mantle.

An even better example is represented by the N-Q structure in the western USA (Fig. 2), which is governed by three fault systems: right-lateral shears on the San Andreas system (northwest trend), left-lateral shears that overthrust in the transverse ridges of northeast and east trend, and northerly-trending faults in the Basin and Ridge Province. The relative orientations are typical of the continental crust and indicate that they were formed under the same conditions of north-south relative compression and east-west relative stretching. The San Andreas system has the greatest displacement, exceeding 300 km, an effect of the North American plate moving to the northeast relative to the adjacent Pacific plate [71].

Geological and geophysical data [96, 103] show that the main N-Q fault systems in the western USA do not extend deeper than 15-20 km. Beneath the San Andreas fault, there is a sublatitudinal zone of high seismic wave velocity in the upper mantle in the area of the Transverse Ranges, which has not been displaced by the fault [32]. The heat fluxes and the $^3He/^4He$ ratios in springs characterize only the southernmost part of the San Andreas fault (Salton Sea area) as a mantle boundary and indicate that the more northerly segments do not penetrate deeper than the granite-metamorphic layer. In the north, the San Andreas fault is not continued by the Mendocino oceanic fault but instead gives way to it without emerging from the continental flank, whereas the Mendocino fault can be traced to the east deep into the continent. Finally, recent reflected-wave studies have given direct evidence that the San Andreas fault is almost vertical in the upper crust, but at 10-15 km it bends sharply and gives way to a horizontal reflecting surface (shear zone?) and is not identified deeper [26].

The average dip of the Great Basin fault is ~60°, and this and the block sizes there, where the blocks can be traced along the dip, define a prism 15-25 km high. If the faults level out deeper down (there are signs of this), the prism height would be less. It is evident that the faults are shallow from the lack of isostatic compensation in the horst ranges they delimit. Additional evidence that the faults do not correspond to deep tectonic elements is the fact that the faults are very seldom employed by basalt magmas as eruption channels, although there are many eruptions in the region. One assumes an equally shallow root for the N-Q Sierra Nevada uplift, as the granitic and metamorphic rocks appear to be underlain by altered oceanic sediments and serpentinites, which occur at a comparatively shallow depth, to judge from the compositions of the abundant xenoliths in the Sierra Nevada pluton and the water and gas compositions in springs.

The deeper-horizon neotectonics in the region can be reconstructed hypothetically from geophysical data, surface structure anomalies, distribution of N-Q basalts, and the present-day heat flux [96, 103]. To the north of the Salton Sea trough, the boundary between the Pacific and the North American continent now lies considerably to the east of the San Andreas fault system at depths >15-20 km (Fig. 2); it runs along the eastern edge of the high-velocity upper mantle zone and continues to the north along the western and northern margins of the Great Basin. The latter is similar to rift zones in that the crustal thickness is reduced to 25-30 km and under it there is a thick lens of deconsolidated mantle with P-wave velocities of 7.5-7.8 km/s. Here one can identify several rift-type deep dislocation zones extending to the northeast showing volcanic eruptions and elevated heat fluxes. Between them, on the surface, are northwest-trending right lateral shears, which possibly correspond to transform-type deformation zones. This reconstructed N-Q structure has features in common with the oceanic structures in Baja California.

The N-Q structure under the western USA, as under the Caucasus, differs from the structure in the upper crustal layer in that there have been extensive horizontal movements and that tearing and slipping appears to have occurred along the boundaries between the disharmonic layers. In the south of the Basin and Ridge Province, where these N-Q forms ceased to develop in the Miocene, the base of the fault-disrupted upper crustal layer is exposed in young uplifts. It consists of rocks that in the Miocene were subject to greenschist dynamometamorphism with indications of flow in directions perpendicular to the faults [34].

The Himalaya, Karakorum, and Pamir mountain system comprises tectonic zones, distinct and varied in history, which were brought together as the Indian and Eurasian plates collided. The Indus zone is a major suture, where the collision has thrown the Cretaceous-Eocene beds into folds and overthrust them on the Tibetan Himalaya. The initial nappe formation, metamorphism, and granitization in the

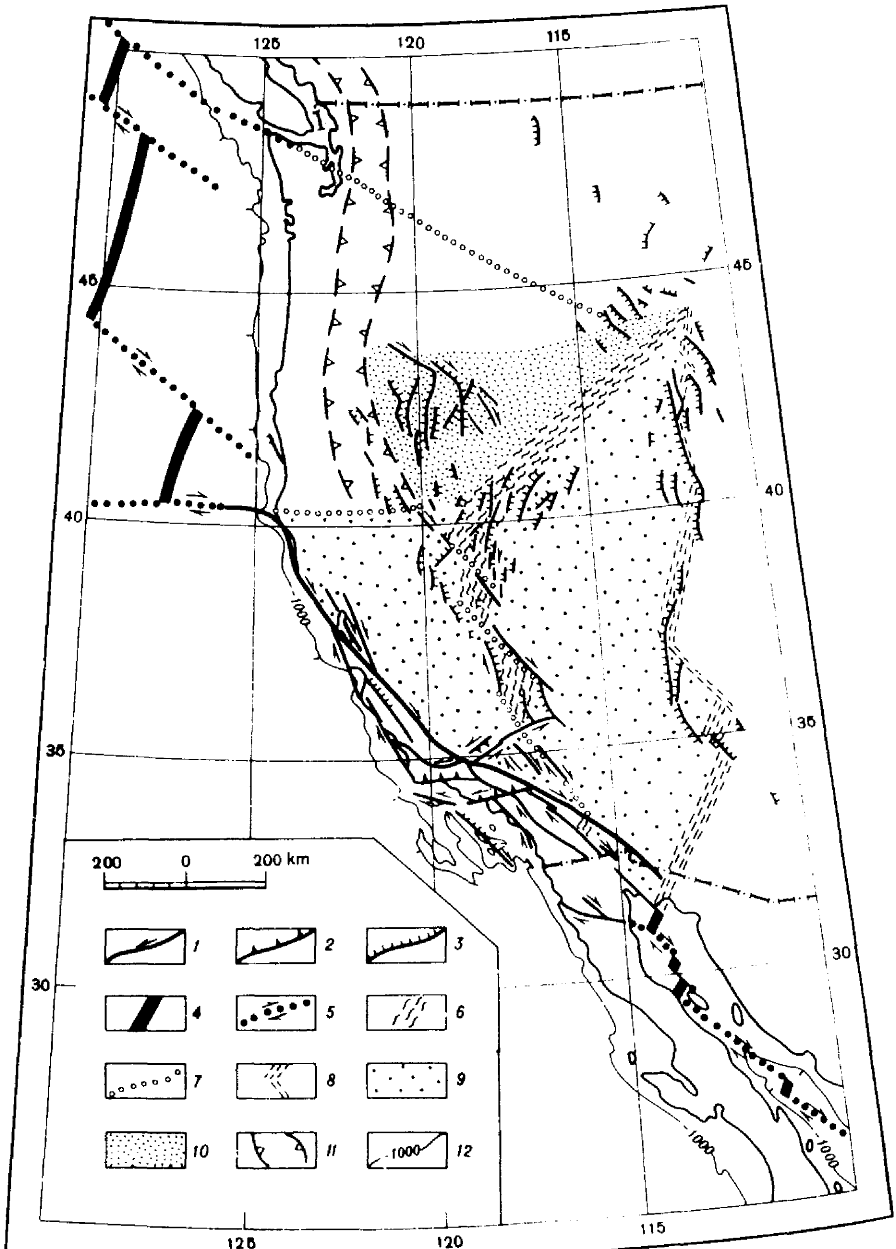


FIGURE 2. Neogene-Quaternary structures in western North America: 1-3) faults in upper continental crust: 1) shear, 2) overthrust, 3) normal faults; 4) rift zones, 5) transform faults, 6-8) deformation zones buried under the upper continental crust: 6) main rift-type zones, 7) main transform-type zones, 8) secondary rift and transform type zones; 9) region of shearing in upper crustal layer; 10) diffuse transform zone on the continental continuation of the Mendocino fault; 11) region of andesite volcanism in the Cascade Range; 12) 1000-m isobath.

Himalayas is associated with the closure of the Tethys, as is evident from the Cretaceous and Paleogene ages of certain metamorphic and plutonic formations [7]. It is possible that this was also the time at which the granite-metamorphic layer in the southwest Pamir was sheared off and compressed [65, 95] and the recumbent folds in the Central Pamir were formed [82]. Similar deformation in upper crustal strata of Tibet may have begun then also [28]. There are Oligocene-Miocene conglomerates lying unconformably on the folded Cretaceous-Eocene strata in the Indus zone [99], which indicate the end of that stage.

The subsequent neotectonic evolution proper in the Pamir-Himalaya region occurred with consolidated and partially deformed continental crust. The N-Q stage involved accelerating uplift [28] and the formation of arcs by structure deflection, which reproduce the angular form of the northern margin of the Indian plate, which is known as the Punjab syntaxis [113].

The structural environment of a syntaxis is exemplified by a system of faults active in the Late Pleistocene and Holocene (Fig. 3); this begins in the southwest in the vicinity of Karachi as a series of meridional faults extending into the continent from the floor of the Indian Ocean. The northernmost of the faults, the Chaman fault, extends almost 100 km and north of Kabul joins up with the Darvaz-Alai young fault zone, which extends from south to north across Pianj River and then turns to the northeast and east-northeast and reaches the southern margin of the Alai Valley. In the east, the end of this zone adjoins the Pamir-Karakorum fault zone, which extends to the southeast and there adjoins the Main Himalaya boundary fault. The directions of the young displacements have been determined for all these faults and indicate that India-Pamir is moving to the north relative to adjacent parts of the Alpine-Asiatic orogenic belt [102, 103].

The N-Q structures show that this basic pattern has persisted in the region throughout the neotectonic stage, but that there have been substantial changes in the structural realization. Up to the end of the Oligocene, the northern front of the moving Indian plate was in the Indus zone. During the relative movement, the part of the orogenic belt ahead of the front was subject to extensive folding and overthrusting, which led to its distinction from the blocks in the zones to the north arising from shearing in the western and northeastern environments of the moving masses, which tore away the deformed upper crustal formations. Consequently, they lost the capacity for extensive dislocation and were involved in the general northern drift, whereas the zone of greatest displacement and deformation along the moving mass front migrated stepwise to the north. Age determinations have been made for the most extensive overthrusting and folding in various parts of the Pamir-Himalaya region, and also for the displacements in various parts of the surrounding shear zones, which show that there have been repeated steps in the front. In the Oligocene period, the front lay in the Indus zone; in the Miocene, it was in the Central Pamir zone; in the Late Miocene and Pliocene, in the region of the Karakul overthrust; currently it is at the southern margin of the Alai Valley, which it occupied only since the Pleistocene (Fig. 3). Consequently, one has a series of tectonic nappes having various ages to the north of the Indus zone, which arose in the N-Q period and which have been torn away from their basement and are disharmonic with respect to the deeper elements in the N-Q structure (Fig. 4). In the outer Pamir zone, in the Peter I and Trans-Alai ridges, this is demonstrated by direct observations on the overthrusts and displaced folds, which suggest a general detachment of the sediment cover from the surface of the crystalline basement [30, 85].

Seismological data confirm this vertical crustal disharmony in the Pamir and the subhorizontal N-Q displacement zones. The crustal thickness increases to 70 km, and waveguides can be identified at the base of the granite-metamorphic layer and within it [10, 96, 106]. Scobelev's interpretation of the rock volumes in the upper crust for the various velocity characteristics shows that the guides correspond closely to boundaries in the overthrust nappes. On the other hand, Shchukin [10] [*sic*] distinguished a layer at ~30 km where earthquake hypocenters concentrate, which links up with a hypocenter zone dipping to the south, corresponding to continuation at depth of the active overthrusts at the northern India-Pamir front (Fig. 5). That inclined zone and the horizontal layer of hypocenters continuing it probably correspond to the zone of active shearing for the N-Q upper crustal structures. The number of hypocenters decreases considerably at the base of the crust, and even farther down, at 70-270 km, lie the Pamir and Hindu Kush branches of the mantle

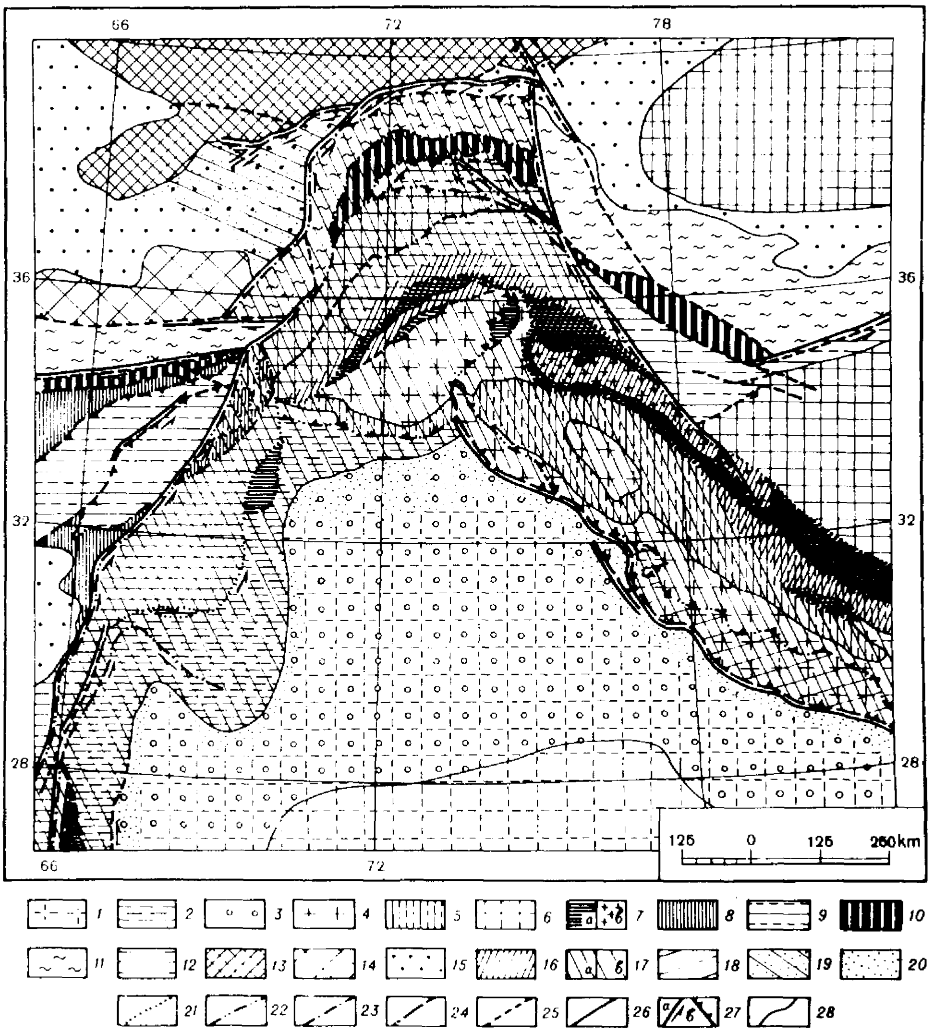


FIGURE 3. Tectonic classification of the Pamir-Himalaya region, compiled by the author on the basis of data from various sources: 1-5) Indian platform: 1) part of Indian platform not showing considerable differential movement in the Late Alpine period, 2) Mesozoic-Cenozoic miogeosynclinal trough, 3) Late Alpine foothill trough, 4) Himalayas, 5) northern part of platform bearing Late Paleozoic and Mesozoic beds from southern margin of Tethys (Tibetan Himalayas and analogs); 6-12) internal zones in the Alpine-Asiatic orogenic belt: 6) ancient massifs, 7) Indus and Quetta ophiolite zones (a) and granite batholith in Indus zone (b), 8) Farahrud ophiolite zone and possible analogs, 9) Karakorum and southeastern Pamir zone and analogs, 10) Central Pamir zone and analogs, 11) western Hindu Kush, North Pamir, and Kun Lun zone, 12) Beluchistan flysch zone; 13 and 14) southern Eurasia: 13) Tien Shan, 14) Paleozoic basement of Turanian platform; 15) Cenozoic and Mesozoic troughs in inner part and northern edge of Alpine-Asiatic orogenic belt on consolidated continental crust; 16-20) times of maximum occurrence of differential Late Alpine horizontal movements in Pamir-Himalaya region: 16) Oligocene, 17) Late Oligocene and Miocene (a), Miocene and probably Pliocene and Quaternary (b), 20) Quaternary; 21-25) start of extensive Late Alpine horizontal movements on faults: 21) Oligocene and earlier, 22) end of Oligocene and Miocene, 23) end of Miocene and Pliocene, 24) Quaternary, 25) age unknown; 26) faults active in the Late Pleistocene and Holocene; 27) strike-slip zones (a) and overthrusts (b) (symbols 26 and 27 shown together with 21-25); 28) boundaries of tectonic zones not containing faults.

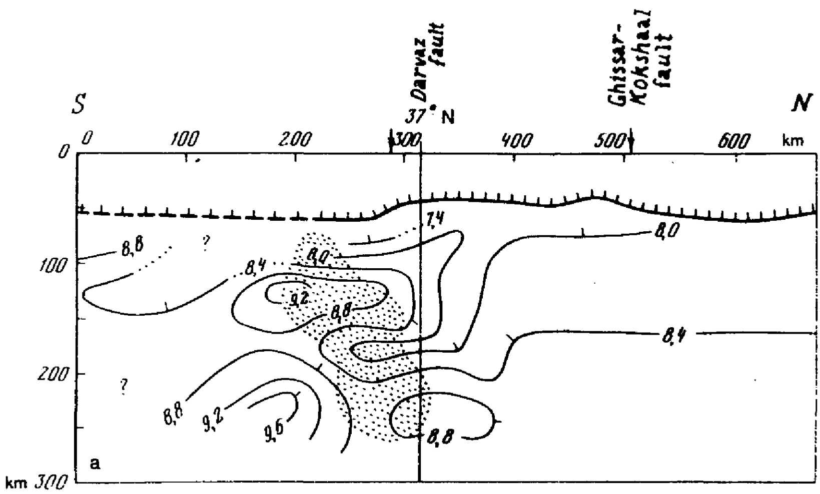
seismic focal zone. These dip steeply with the Pamir branch lying 150 km to the south of the dipping upper crustal zone, which confirms the disharmony between the upper crustal and mantle structures in the region.

The seismic characteristics of the Pamir-Hindu Kush focal zone shed light on the mantle structure. First, distribution of hypocenters of mantle earthquakes shows a vertical nonuniformity (Fig. 6). Second, Vostrikov's [106, 112] method of interpreting earthquake recurrence plots is used to evaluate the relative shear stresses acting in the focal zone, the seismic deformation rates, and the effective seismic viscosity, and has shown that there are considerable vertical variations in those characteristics (Fig. 7). Finally, transit-time discrepancies for *P*-waves from local mantle earthquakes as recorded at stations in Central Asia have been used by Nikolaev and Sanina [59] to detect considerable vertical and lateral inhomogeneities not only in the seismic focal zone but also to the south of it at depths down to 300 km (Fig. 5). Tectonic layering and depth-differentiated displacements equivalent to those observed in the crust of the Pamirs evidently occur to the same extent, or more so, in the mantle.

In the N-Q period a series of north-verging overthrusts was formed in the Himalayas, which show the same order in neotectonic history as in the Pamirs, but with the opposite direction: the region showing the most extensive horizontal displacements has migrated to the south. These overthrusts are not traced deeper than the upper crustal layer. Other faults occur at the level of the Moho, and lower the base of the crust steps to 70-80 km under the Himalayas (Fig. 4) [52], so here again there is marked disharmony between the faulting at the top and bottom of the crust. It is notable that both of these are lost sight of in the lower part of the crust, being probably replaced by other forms of rock deformation.

The neotectonic layering found in the Pamir-Himalaya region, East Caucasus, and western USA can also be observed in other folded-mountain structures, where there are differences in structure and position for the N-Q structures at various lithospheric levels. Makarov [49, 50] examined the neotectonics of Central Asia from this viewpoint and concluded that the N-Q faults that disrupt subsurface horizons are accompanied by deeper latent ones,

FIGURE 4. Pamir-Himalaya lithosphere succession: 1) ophiolites; 2) sediment cover; 3) consolidated part of crust, cataclastic in lower part, but at the top of the cataclastic part, there is a subhorizontal layered zone, with most of the upper crustal faults not extending beyond it; 4) upper mantle.



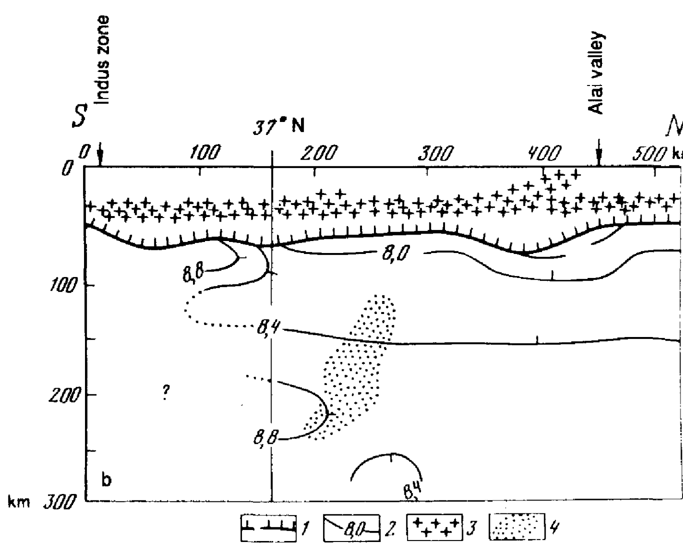


FIGURE 5. Geophysical sections for the Hindu Kush and Tadzhik depression (left) and Pamir (right): 1) Moho [10]; 2) lines of equal P-wave velocity in upper mantle, km/s [59]; 3) crustal seismic focal zone [10]; 4) mantle seismic focal zone [5].

which are indirectly reflected at the surface in geomorphological, geochemical, or hydrochemical features or only as geophysical anomalies.

Shchukin [96] has demonstrated crustal layers beneath the East Carpathians, the Crimea, and the Tien Shan with substantially differing numbers of hypocenters; the general trend for strong shallow-focus tremors is notable: The number decreases below 20-30 km (Fig. 8), which is characteristic of all intracontinental and continental-margin active zones, as well as of island arcs where there is an extensive-granite-metamorphic layer, but is absent in those island arcs where the latter is slight or absent. The upper and lower parts of the continental crust differ most markedly in deformation features.

In an island arc having an intermediate crust type, hypocenters cluster in a subhorizontal zone at the boundary between crust and mantle (Fig. 8f). What

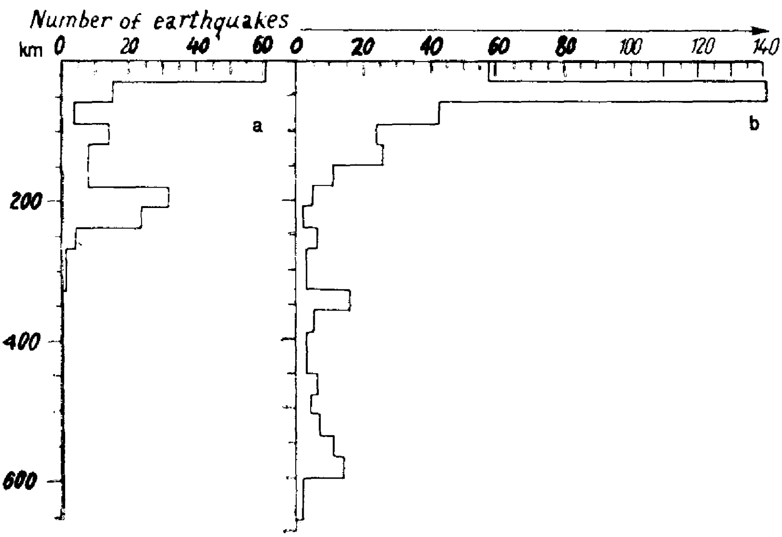


FIGURE 6. Hypocenter depth histograms for $M \geq 6$: a) Pamir, Hindu Kush, and Tien Shan; b) Kurile, northern Japan, Sea of Okhotsk, and Sea of Japan [57].

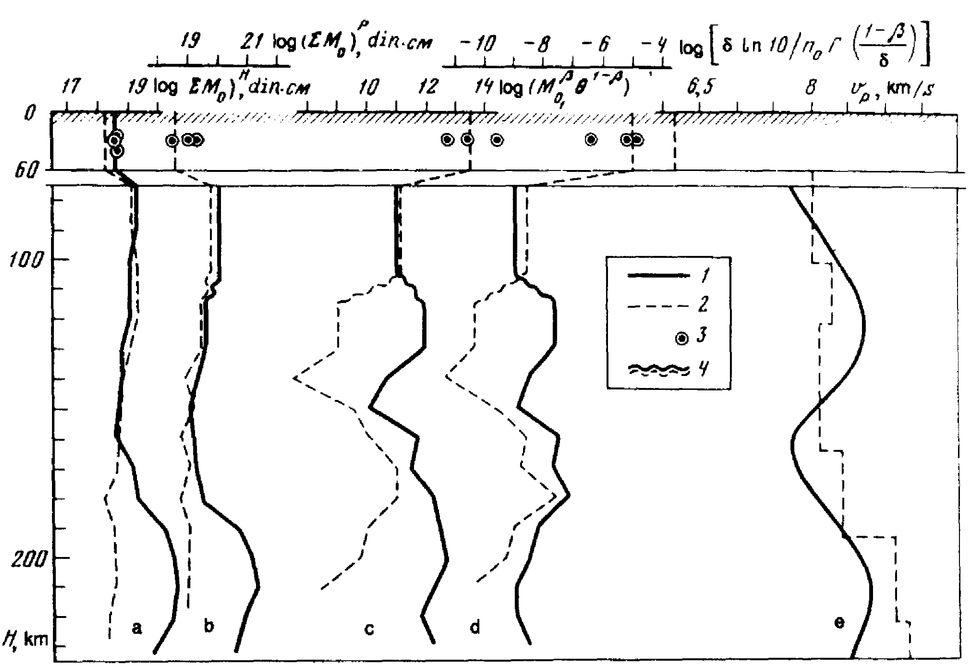


FIGURE 7. Depth dependence for the Hindu Kush and Pamir branches of the Pamir-Hindu Kush focal zone for the observed (a) and calculated (b) sums of the seismic moments which characterize the seismic deformation rates; parameters $M_0^{\beta}1\theta^{1-\beta}$ (c) and $\delta \ln 10/\eta_0 \cdot \Gamma((1-\beta)/\delta)$ (d) characterizing the state of stress and the seismic viscosity; P-wave velocity (e) [106]: 1) Hindu Kush branch; 2) South Pamirs branch; 3) Kamchatka region; 4) unreliable measurement ranges.

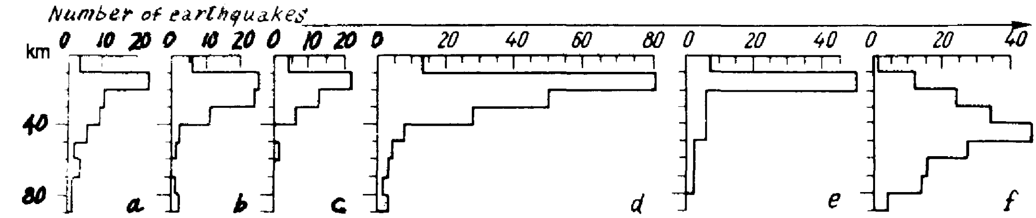


FIGURE 8. Depth distributions down to 90 km for hypocenters of earthquakes in the USSR and adjacent areas having $M \geq 6$: a) Near and Middle East, Caucasus, and Kopet Dag; b) Pamir, Hindu Kush, and Tien Shan; c) Altai-Sayan region, Baykalia, Transbaykalia, Mongolia, and adjacent parts of China; d) continental part of the USSR and adjoining foreign areas as a whole; e) Kamchatka; f) Kurils and northern Japan [57].

appears in the first approximation to be a single inclined seismic focal zone on more detailed examination frequently splits up into distinct subhorizontal lenses [93], which are separated by weakly seismic or aseismic parts. Such a pattern occurs in the Kuril and other arcs and in the West Pacific, from which Tarakanov [93] deduced that the presence of several asthenosphere layers. Such lenses up to 250 km long are evident also on the transverse profile for the Lesser Antilles arc at 30-40 and 100-120 km [100].

Crustal waveguides also indicate tectonic layering in active regions; many of them show reduced effective viscosity, which facilitates disruption and displacement. Sometimes, more detailed judgments can be made on these waveguides. For example, the waveguide in the Southern Alps is overlain by a plate with P-wave velocities of 7.2-7.38 km/s, which dips to the southeast and joins up with the mantle (8.3 km/s). The few mantle exposures in the Ivrea-Verbano zone are marked by lherzolites. It is quite clear here that the velocity inversion is related to the mantle plate being overthrust together with the Southern Alpine section resting on it onto the Pennine and East Alpine zones [65]. Krasnopevtseva and Shchukin [42] examined areas showing strong earthquakes in the Caucasus and assumed that the crustal waveguides found there can have arisen as a result of dynamic equilibrium recovery in a medium disrupted by the earthquakes. In other words, the waveguides may have formed when the rocks weaken during detachment at plate boundaries.

We have at present little evidence on possible tectonic layering in oceanic active regions, although it is indicated by discrepancies in the structural patterns found in transform zones to the north and south of Iceland and in certain parts of the Mid-Atlantic ridge [103], particularly the waveguides in the lower parts of the crust north of the Azores, as well as in the East Pacific and elsewhere in addition to the gently dipping seismic reflection surfaces found in the third oceanic layer in the Mid-Atlantic ridge at 20°N, which can be traced across strike for 200 km [74].

In a theoretical respect, it is very important to establish whether this layering extends outside the active belts. Pavlenkova [86] examined deep seismic sounding data for the USSR and found crustal waveguides in virtually all types of continental structure: ancient shields, various Phanerozoic folded regions, and ancient and young platforms. Continuous reflected-wave profiling has given new information on these. Such measurements on basement faults in the Ukrainian shield have shown that systematic changes in dip from steep at the surface via medium (30-35°) at depths from 2 to 15 km to gentle or almost horizontal at the top of the crustal waveguide [88]. The Kola superdeep borehole data [39] show that the reductions in rock density and seismic velocity at 4.5 km (signs of the crustal waveguide) are related not to changes in primary rock composition but to an increase in shearing on passage from the greenschist to the epidote-amphibolite facies; there is an increase in water content due to the increased fracturing.

This indicates that the reduced-velocity layer is of tectonic or dislocation origin; Nikolaevsky, Sharov, and Sherman [60, 87, 89] consider that fairly thick crust subject to tangential tectonic stresses reacts in a differential fashion controlled by the lithostatic pressure, i.e., the depth. Shearing occurs in the upper crustal layer, whereas deeper horizons develop numerous small fractures, which lower down pass into mylonitization and blastesis. At intermediate depths, bulk disruption leads to deconsolidation, which is recorded as reduced seismic velocities [60, 87]. Layering and the production of a dislocation waveguide are thus regular results from the tangential forces. Rocks are displaced along the waveguide on subhorizontal tears and fractures, which is facilitated by the elevated water levels. In deeper horizons, quasiplastic and plastic strains predominate. With vertical composition inhomogeneity, the layering pattern becomes more complicated.

This layering outside active regions is due to ancient tectonic processes, but it provides potential for depth differentiation in N-Q lateral displacement, i.e., present-day tectonic layering under various conditions.

2.2. Crustal deformability

The lower crustal and mantle parts of the continental lithosphere are capable of considerable plastic or quasiplastic strain, as has been demonstrated in laboratory rock measurements, and by direct studies on rocks reaching the surface, whereas the upper part of the crust is considered as stronger and brittle. However, here also, the assumption of negligible deformability called for by classical plate-tectonics can be adopted only with considerable reservations and additions.

Figure 9 shows major active faults in Asia between 20° and 60°N, they confirm previous conclusions [56, 102, 103] that horizontal displacements predominate over vertical ones as regards amplitude and in particular that Arabia and the Indian-Pamir region are approaching more northern regions in Eurasia, as is evident from the left-lateral displacements on the western and northwestern flanks of Arabia and the Indian-Pamir region and the right-lateral ones on the northeastern flanks. There are sublatitudinal shear lines in northern Anatolia, Iran, Afghanistan, and the Tadzhikistan depression, as well as Tibet, the Kun Lun, and the Altyntag, which reflect compression from the northward movement of the Arabian and Indian-Pamir plates [56, 102, 103]. However, there are also two new circumstances.

First, the relative displacements in the major plates extend to mobile belts hundreds of kilometers wide, so the plate edges are subject to considerable deformation and internal displacement. The Pamir-Himalaya edge regions shows Holocene deformations and is comparable with the relatively slightly deformed part of the Indian subcontinent. Farther east, the mobile belt becomes even wider. It is best to consider the numerous active fault lines with their various directions, which are combined with folding in the cover and basement, not as due to interaction between rigid plates or microplates but as a consequence of plate deformation.

Second, displacements on the Asian active faults reflect predominantly shear. Most of the faults run along the boundaries between mountain chains and adjacent depressions. Field studies confirm that there are vertical displacements on the faults, which maintain the relative mountain uplift. However, many of these boundary faults have a shear component in the Holocene displacements, which is comparable with or often larger than the simultaneous vertical component. The calculated Late Quaternary shear-displacement rates are millimeters a year, and now they exceed centimeters a year (Table 1). These longitudinal shears are thus of much greater kinematic importance than had previously been thought, which may be due to the energy economy in these shear displacements, as it is not necessary to overcome gravity.

On the other hand, there are only quite small areas showing mainly folding and overthrusting in the Holocene faulting (such as in the outer zone of the Pamir or the Himalayas) or downfaulting and rifting (such as in Baykalia or the Shanxi graben). There is also a further feature: The vertical component in most of these Asian shears is an upfaulting one, no matter what the strike. Downfaulting occurs only in coastal parts of East Asia, in the Baykal rift system, in the Lut block in Iran, and in a few other places. In other words, most of the intracontinental active zones involve compression.

Belts can be distinguished where there are mainly sinistral or dextral offsets. For example, faults with dextral offset are important in the North Tien Shan, Dzhungaria, and the Mongolian Altai. Another belt with right-lateral offsets extends from eastern China (the Tanlu fault) to Kamchatka (the eastern margin of the Central Kamchatka depression). Between

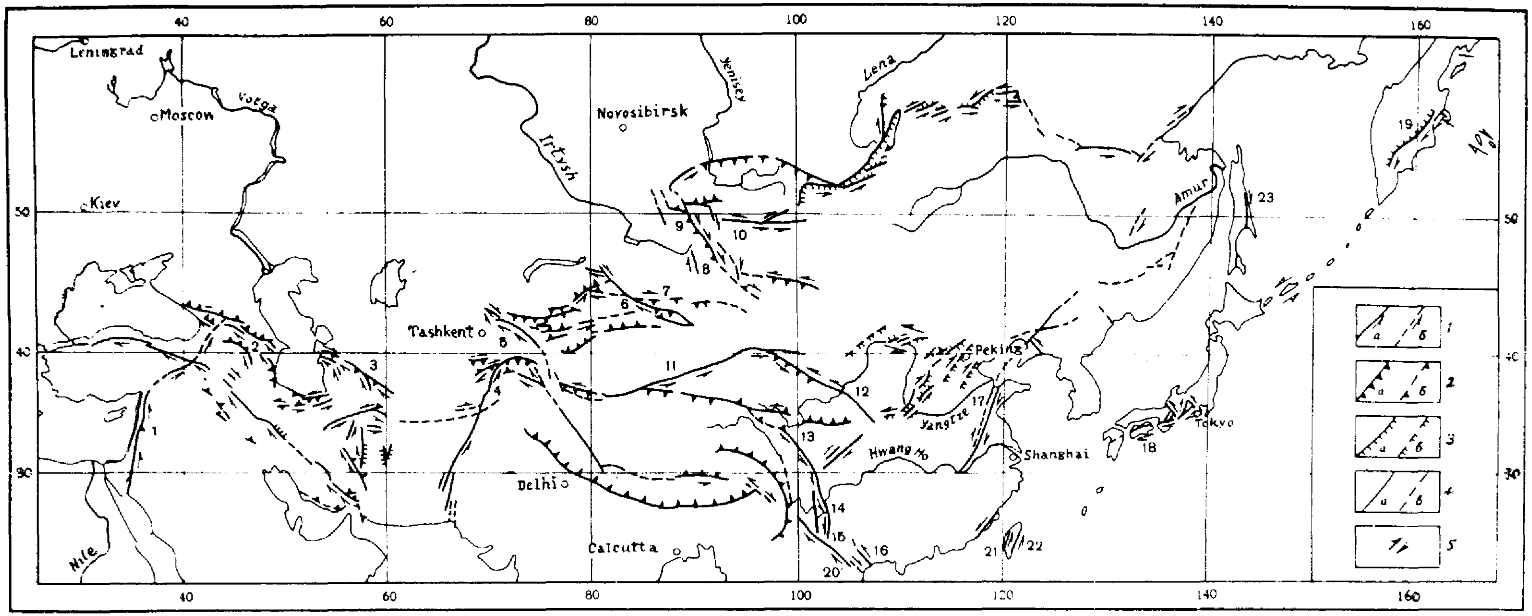


FIGURE 9. Map of Asian faults active in the Holocene between 20° and 60°N compiled by the author [104] and published data from various sources: 1-4) active faults (a reliably established, b supposed): 1) strike-slip; 2) overthrusts and reverse faults; 3) normal faults; 4) with unknown displacement directions; 5) shearing zones. The numbers for faults are: 1) Levant; 2) Kobystan; 3) main Kopet Dag; 4) Darvaz-Alai; 5) Talas-Fergana; 6) Dzhungaria; 7) North Tien Shan; 8) Ertai; 9) Kobda; 10) Khangay; 11) Altyntag; 12) Nanyueshan; 13) Xianshui; 14) Zemu River; 15) Xiaochan; 16) Kaobang-Langshon; 17) Tanlu; 18) median line in Japan; 19) eastern edge of Central Kamchatka depression; 20) Red River; 21) West Taiwan; 22) East Taiwan, 23) East Sakhalin.

TABLE 1. Rates of Lateral Offsets on Active Faults of Asia, mm/yr

No. in Figure 9	Name of fault	Sinistral (L) or dextral (R)	Instrumental data	Geological and geomorph. data		Reference
				Late Holocene	Late Quaternary	
1	Levant (Dead Sea/NW Syria)	L		7.5/5		117/94
2	Kobystan	R			1	103
3	Main Kopet Dag	R			2	103
4	Darvaz-Alai	L		20	10-15	103
5	Talas-Fergana (SE part/ Central part)	R		5-7/15-20	6-8	94/17
6	Dzhungaria	R			9-18	21
7	North Tien Shan	R	9			21
8	Ertai	R			13-17	21
9	Kobda	R		5-6		19
10	Khangay	L		9		19
11	Altyntag	R	2		6-7	21
12	Nanyueshan	L			18	21
13	Xianshui	L	10		3-9	21
14	Zemu River	L			9	21
15	Xiaochan	L		5-6		21
16	Kaobang-Langshon	L			5-10	93
17	Tanlu	R	0.6		2	21
18	Median line of Japan	R			5-10	23
19	Eastern edge of Central Kamchatka depression	R			13-15	40
20	Red River	R			5-10	94
21	West Taiwan	L	6			21
22	East Taiwan	L	12		5	21, 41

these two belts lies a belt of left-lateral offsets in Cisbaykalia and the Stannovoye uplands, Mongolia, and China. The offset strike within a belt can vary substantially. With dextral offsets in the western belt, the strike varies from west-northwest or almost westerly at the southern edge of the Dzhungaria basin to north-northwest in the Mongolian Altai. Even more variation occurs in the sinistral offset belt in South China: from east-northeast on the Altyntag fault to north-northwest on the Xianshui and Xiaochan faults.

Holocene displacements of mainly shear type occur not only in intracontinental mobile belts; at active margins and in Pacific island arcs, the longitudinal faults often also have components of offset exceeding the downfaulting or overthrusting ones [41]. Dextral Holocene offsets are characteristic of the eastern part of the Central Kamchatka depression, the East Sakhalin fault, the median fault and the fault in the Sagami trench in Japan, the Alpine and other faults in New Zealand, the Atacama fault (?) in the Andes, the San Andreas fault and the ones parallel to it in western North America, and faults in Alaska and the western Aleutians. Holocene sinistral offsets occur in Taiwan and the Philippines. These areas have extensive crust of continental, or occasionally transitional, type. Where an island arc has crust of close to oceanic-type, the offset on longitudinal faults tends to be reduced, with the shearing replaced by structures representing oceanic masses subducted under the island arc or, in regions with N-Q volcanism, longitudinal extensional structures. A characteristic example is provided by the Tonga-Kermadec arc along the continuation of the Alpine shear in New Zealand. The dextral offset median fault in Japan is not traced on seismic data to depths greater than 20 km, but at a depth of about 30 km there is a seismic generating structure having a different orientation, which reflects a geodynamic setting different from that in the upper crust. Thus the offset type of displacement is most characteristic of the upper continental crust.

* * *

This evidence extends the mobile theory of tectogenesis and modifies our concepts on the homogeneity of vertical deformation in the lithosphere and on the negligibly small deformability outside zones of direct interaction between monotonically moving plates. The lithosphere is layered, if not everywhere, then at least over considerable areas. There are two aspects to this. First, at different levels one gets largely independent N-Q structures and different forms of deformation and displacement. Second, plates and blocks with different types of deformation are separated not only by vertical and inclined zones but also by subhorizontal zones and show contrasting movements: asthenosphere layers and lenses. The latter can arise without external loading simply as a result of differences in internal stress between the rock levels. The potential for such layering increases considerably when there are external forces that produce lateral displacement, so such layering is detected not only in the crust but also the upper mantle, where the boundary between the lithosphere and asthenosphere becomes highly arbitrary.

There are signs of such layering in the oceanic lithosphere, but they are more prominent in the continental; the latter is divided at least into upper crustal and upper mantle relatively rigid plates, with a layer in the lower crust separating them (this includes the crustal waveguide), whose strength with respect to lateral displacement is reduced. In regions of N-Q compression, such layering is more pronounced, and, consequently, the directions and rates of the N-Q relative displacements calculated in plate tectonics in many places characterize only the crustal plates, or even only the upper crustal ones in regions where the crust is thick enough, and they may differ from those in deeper parts. At the same time, plate-tectonic kinematic calculations are inaccurate even in relation to the upper crustal plates, because they do not incorporate the deformations and internal displacements in the plates, which extend to the vast mobile belts.

In continental mobile belts, offsets predominate in fault displacements, which has been underestimated in kinematic constructions; zones where active compression or stretching predominate cover only small areas in continental belts by comparison with those showing shear tectonics, which shows that upper crustal plate collision or separation has a restricted structuring role. These plates are deformed by numerous offsets and cannot transmit structuring forces to large distances, whereas the offsets themselves are not due to such interactions but rather to the displacements of deeper masses.

3. Neotectonic Systems

3.1. The tectonic-system concept

In statistical physics or thermodynamics, any physical body, no matter what its size, composition, and state, is considered as a system whose state is determined by a set of parameters (temperature, pressure, density, chemical potential, energy content, etc.). These parameters are related, so that change in any one of them involves linked processes affecting the others, which restore equilibrium. In tectonics, we are usually dealing with more complicated processes, which involve links between parameters, each of which is a system from the viewpoint of, for example, physics. Nevertheless, the coupling principle still applies. A tectonic system is a set of natural processes linked in a particular volume of the geological environment and leading directly or indirectly to movement in the lithosphere and the development of structural forms [70, 107].

In engineering, there is the concept of structural stress, in which stresses within a certain volume are, in the absence of loads, balanced at the boundary surfaces. These stresses provide a measure of the deviation from the equilibrium state and are closed within microscopic volumes or at most within machine components. In the geological environment, the stresses also form closed systems, but they may be balanced not only within local areas but also over substantially larger volumes. A tectonic system is a structural-stress system developed at various levels of organization in the medium as a result of deviation from the equilibrium state for any of the parameters characterizing it as a thermodynamic system [70]. The system rank is indicated by the size of the region in which the links between the elements are closed. In that sense, one can speak of systems of a global scale and of local ones of various ranks.

3.2. The global neotectonic system

Any process affects all the others in a system and under certain conditions may be the primary cause of all subsequent changes, large or small. In a global tectonic system, the decisive processes governing the consequent changes involve displacements of mantle material, which are caused by deep-seated differentiation and are affected on a global scale by mantle convection.

Seismic tomography [5, 22] shows that mantle volumes differing in velocity have a complicated distribution; there are regions where hot material is rising, the largest of them being under the Pacific, which can be traced from the deepest mantle levels and even shows a relationship to the irregularities of the core surface. In the upper shells, the main ascending flows are associated with the oceanic rift zones, although the two do not always coincide exactly or completely. The differentiation products influence the oceanic lithosphere and produce differential movements at various levels toward the adjacent continents (Fig. 10).

The seismic focal zone dipping toward the continent plays an important part in island-arc and active continental-margin structuring; the rock chemistry varies regularly in that direction, which indicates that the composition is at least partially related to reworking of oceanic lithosphere descending along the focal zone, although there is also other evidence, particularly strontium isotope ratios, indicating that eruption products are related to wedges of lithosphere rocks above the focal zone.

The upper part of the Japanese focal zone has two branches defined by hypocenters of weak earthquakes: one, which is in places split into two close parallel bands, descends without gaps to depths of 150-170 km, and the other, which is almost horizontal, is associated with the top of the crust and the lower mantle [23]. In both branches, the numbers of hypocenters decrease sharply as active volcanic regions are approached. The hypocenters of strong quakes show similar trends, but with the difference that some of them lie in the wedge above the descending branch traced from the weak tremors. This combination of subhorizontal shallow and dipping branches is found, according to Fedotov and Boldyrev [93], in the Kuril-Kamchatka and other island arcs of the Pacific margin. Vostrikov employed a method of analyzing earthquakes recurrence plots [106] to show that the focal region in the Kuril-Kamchatka arc comprises subhorizontal and dipping zones where the seismic deformation rates increase at depths up to 50 km, where there are also increased stresses and high gradients of effective viscosity.

In these regions, there are velocity inhomogeneities in the mantle down to 150-200 km, which can be related to heating and magmatic reworking above the focal zone [5]. Fedotov [93] directed attention to the anomalous wave damping under major volcanics in Kamchatka and the almost complete lack of seismicity in the mantle under the active volcanic belt at depths of 20-100 km. The vertical projections of the volcanoes on the dipping branch fall in the depth range 100-200 km, where the number of tremors decreases sharply, with andesite volcanoes often projecting to smaller depths than basalt volcanoes. There is a layer of deconsolidated low-velocity mantle under the Kuril and Kamchatka volcanic belt, which Boldyrev [93] found to dip about 20° toward the continent and to be traceable to 250 km.

Below 200 km, lateral velocity inhomogeneities that can be related to reworking of oceanic lithosphere in the focal zone are much fewer; the hypocenters do not form a continuous belt but instead form groups lying approximately on the continuation of the zone [72]. For example, large deep-focus earthquakes under the Sea of Japan form two groups: a small one at 300-350 km and a larger one at 500-600 km. This suggests that much of the lithosphere descending along the zone to about 200 km has been reworked so extensively that it approximates to adjacent mantle in physical properties. The earthquakes with deeper foci in these groups may be different in nature from those down to 200 km [72] and be due in part to local increases in deformation rate at the ocean-continent boundary. On the other hand, the wave propagation anomalies for deep focus earthquakes under the Sea of Okhotsk and the Sea of Japan [20] suggest inhomogeneities possibly related to residual products from reworking of oceanic lithosphere down to 900-1000 km, which may also serve as sources for some of the earthquakes.

This evidence leads us to represent the movements at the margin as follows. Part of the oceanic lithosphere approaching the boundary is subducted

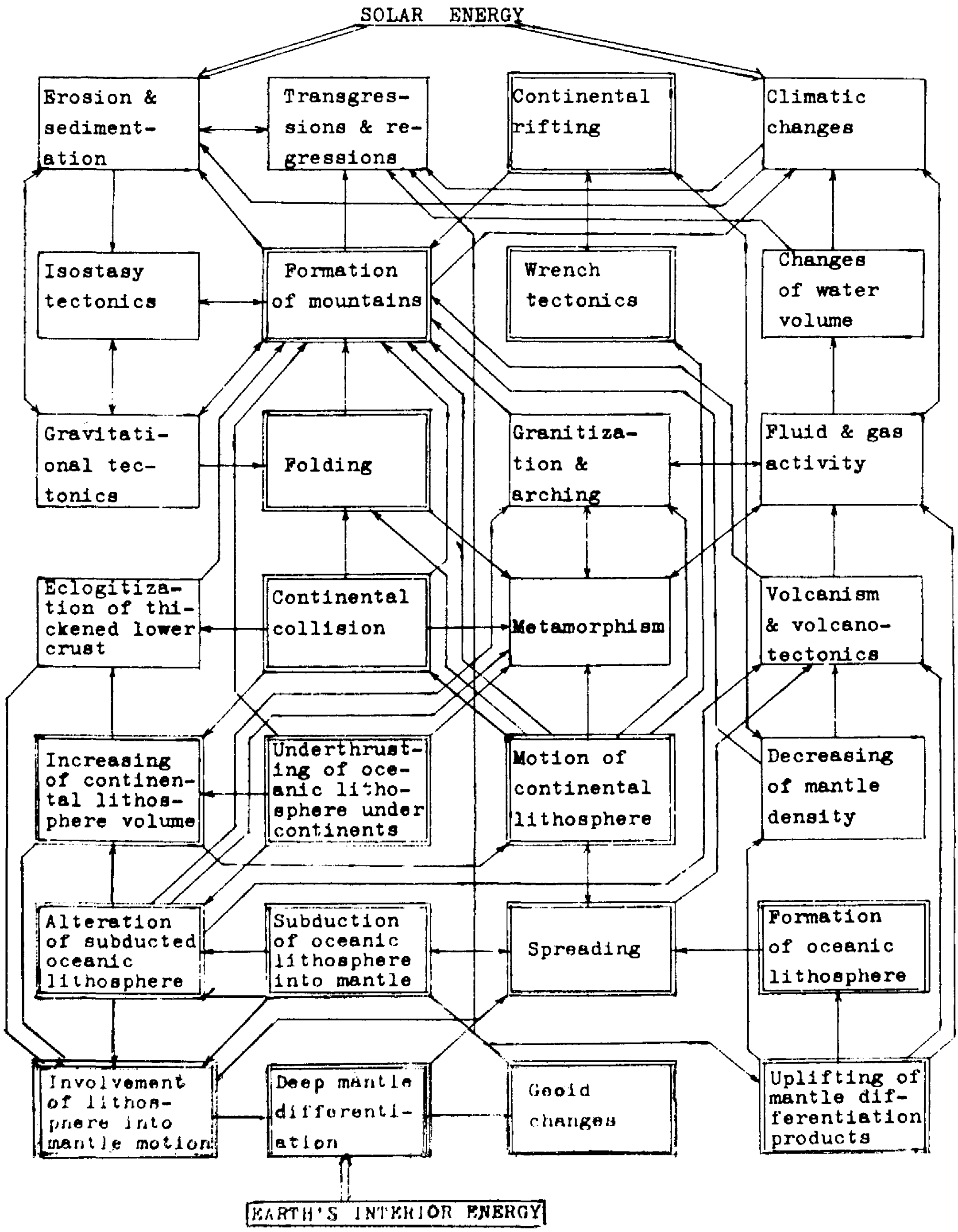


FIGURE 10. Relation between neotectonic processes, with elements in global neotectonic system outlined by double lines.

under the island-arc crust or continental margin along subhorizontal planes, while another part, probably the large one, descends along the focal zone. It is largely reworked down to 200 km. The products are accreted to the lithosphere at the active margin, and only the fragmentary residual masses may preserve their distinctive existence and continue to descend to considerable depths. An active margin is therefore not merely a zone of descending flow but also, and perhaps more so, a region of oceanic lithosphere reworking.

The upward-migrating materials may be partially erupted on the surface or otherwise affect the crust in the margin or island arc, but a large amount will concentrate under the crust as a deconsolidated lens. This lens dips gently and extends toward the continent, which may indicate movement in that direction. Such movement would explain marginal seas as results of continental-crust thinning and breakage above mantle bodies moving under the continent. Parts of the granite-metamorphic layer of the former continent edge may be torn off the moving lithospheric masses as a consequence of thermomechanical weakening of the crust related to island-arc volcanism, and these often persist in island-arc successions, as in Kamchatka or Japan. As the masses move farther under the continent, there is differential buckling in the continental lithosphere, which may be related to "intraplate" orogeny, such as extends to vast areas in East Asia and in the western parts of North and South America.

At existing passive margins, there are no specific boundary N-Q structures apart from longitudinal normal faults, which may be seen, for example, in some parts of the Atlantic coast. The moving oceanic lithosphere at a passive margin therefore appears to entrain the adjacent continent, which is displaced in the same direction and at the same rate. There are, however, certain differences in the rates of lateral displacement for continental and oceanic blocks recognized as parts of individual plates [96]. For example, the rate of collision between Eurasia and the northern part of the Indian Ocean as indicated by the youngest banded magnetic anomalies is 5-6 cm/yr [45], whereas the collision rate for the subcontinent obtained by summing the Late Quaternary deformations and displacements at the surface in the Tien Shan-Pamir-Himalaya region is ~4 cm/yr, and that found for the Arabian subcontinent by similar summation for Asia Minor and the Caucasus is less than 3 cm/yr [103]. The rate calculated from the banded anomalies in the ocean is probably close to the rate in the upper mantle, which thus moves more rapidly than the upper crustal continental formations.

A similar situation occurs in Iceland, which is composed of anomalously thick (over 20 km), although oceanic crust; summing the Late Quaternary displacements here gives a surface extension rate of ~1 cm/yr [103], whereas in adjacent parts of the Mid-Atlantic ridge, the magnetic-anomaly pattern indicates a value exceeding 2 cm/yr [111], which suggests that the various lithosphere layers move at different rates, generally increasing with depth. However, there is another important point: the general tendency, at active and passive margins alike, for the oceanic lithosphere reworked to various degrees, and the mantle material beneath it, to move under the adjacent continent, which may lead to buckling and thickening of the continental lithosphere.

The greatest amount of buckling occurs in areas of continent-continent collision such as the Pamir-Himalaya region. Tectonic layering leads to nonuniform buckling at the different levels, which on the whole leads to thickening in the crustal and mantle lithosphere sections. The buckling is a major factor in orogeny, supplemented by isostatic uplift, which compensates for surface erosion, as well as by other secondary processes. Folded sheet structures occur widely in the upper crustal layer. At 40-70 km, crustal rocks are subject to partial melting, which leads to granitoid magmatism and metamorphism in the overlying rocks, accompanied by stress metamorphism in extensively deformed zones. The residual melting products may eventually be converted to eclogites [7] and thus come to approximate mantle rocks in physical properties. Such transformations (along with surface erosion) may explain why the present depth of the Moho beneath the Paleozoides of Eurasia is usually not more than 40-45 km, despite the fact that geological data indicate they underwent buckling comparable to that in the Pamir-Himalaya region.

The Pamir-Hindu Kush mantle focal zone thus formed under these conditions in part by buckling in relatively cold upper mantle sections capable of brittle failure. However, it is more likely, particularly at substantial depths, that there is a relation between mantle seismicity and increased rates of strain in

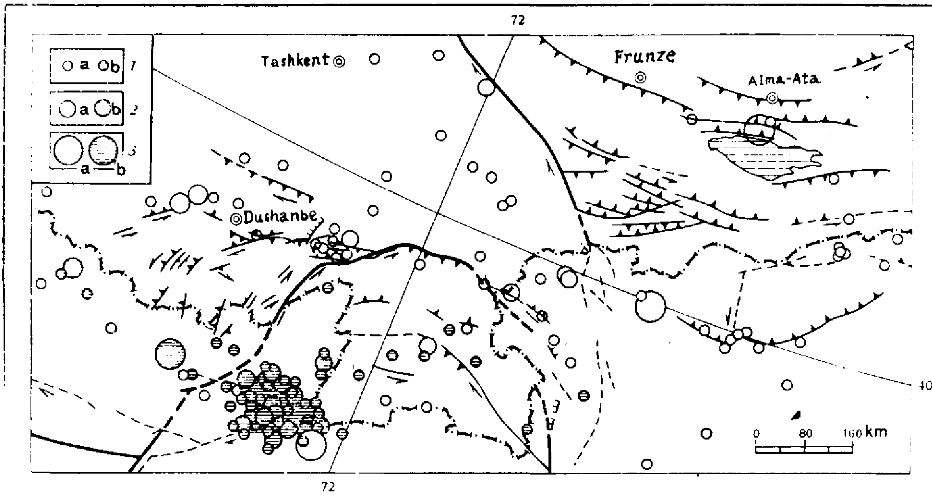


FIGURE 11. Pamir and Tien Shan active faults [51] and epicenters of earthquakes in 1900-1974 having hypocenter depths: a) up to 70 km; b) between 70 and 300 km; magnitudes M: 1) $6 \leq M < 7$; 2) $7 \leq M < 8$; 3) $M \geq 8$ [57]. See Fig. 9 for other symbols. Major faults are shown by bold lines.

restricted volumes (the funnel effect), as is clearly illustrated from the areal distribution in the Pamir-Tien Shan region of epicenters for strong crustal and mantle earthquakes (Fig. 11). The same origins may apply for other intracontinental mantle seismic focal zones such as the Vrancha zone in the Carpathians.

The N-Q tectonics in the upper crust in a mobile belt gradually changes away from the regions of maximum compression and buckling; this is best seen in Central Asia. There are large basement folds in the Tien Shan, which extend to the entire upper rust and which are complicated by overthrusts, reverse faults, and folds in the sediment cover [48]. There may have been some mixing between the crustal and mantle rocks, which in part may be related to deconsolidation in the uppermost mantle, as has been suggested for the Tien Shan. The basement folds correspond to ridges and intermontane depressions. Crustal buckling decreases away from the Pamir to the northeast, and basement folds of the Tien Shan type are replaced by horsts and ramp grabens, which are merely dissected, and not everywhere at that, by flexures and peneplains. Offsets along faults increase. This is found in the Tien Shan and also farther from the Pamir in Mongolia and in western and central China, where these offsets become predominant, and vertical displacements are very much subordinate. In Mongolia and China, the offsets and faulting shears form groups in extended belts, whereas in the Soviet part of the Altai and in Tuva, they are more varied in direction and result in a complete mosaic of blocks. Flexures *en echelon* arrangements of sheared belts are associated with tensional depressions such as the Baykal and Shanxi graben.

It appears that the N-Q structuring in the continents is not restricted to the above processes in intracontinental and marginal mobile belts; under all the continents, there are high-velocity and relatively cold volumes of mantle material of lithosphere type that can be traced down to depths of over 150 km [5]. These include the ancient cores: much of northern Eurasia, the southwestern half of Africa, the northern part of the Indian subcontinent, Australia, and the Canadian and Brazilian shields (Fig. 12). Two interpretations can be given.

On the first explanation, the elevated velocities in the upper mantle and accompanying reduced tectonic activity in cratons and in other parts of the continents outside the mobile belts must reflect the cooling in the upper mantle extending to considerable depths, and also reduced or lacking asthenosphere, and consequently links between the crust and upper mantle to deeper horizons in strong lenses with uniform deformation parameters. Evidence against

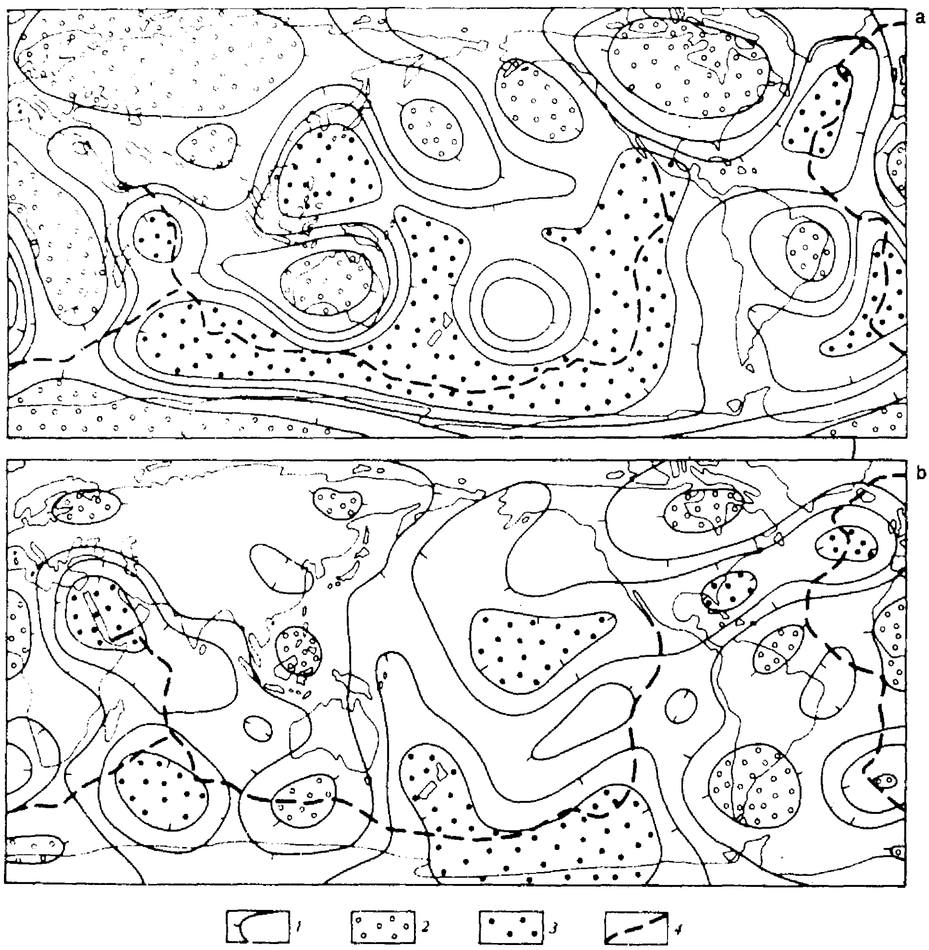


FIGURE 12. Distributions in nominal isoline terms for the volumes of various velocities in the mantle at depths of 150 km (a) and 350 km (b) [5]: 1) *P*-wave velocity isolines, with hatching on the side of lower velocities; 2 and 3) regions of highest seismic-wave velocity (2) and least (3); 4) major elements in oceanic rift system.

that explanation is provided by the tectonic crustal layering in the continents outside the mobile belts, as well as the rapid isostatic compensation for lateral displacements in mantle masses related to melting of the icecaps after the last glaciation in the Baltic and Canadian shields [6].

Therefore, the more likely explanation is the second one, according to which mantle buckles under the continents, this being supplemented by reworked oceanic lithosphere migrating toward them [104]. Such buckling in the mantle may be reflected in tendencies for the crust to thicken from the edges toward the interior in the continents [9] as well as by compression in the upper crustal lithoplates exceeding the lithostatic pressure [43]. An example of the latter is provided by the North American craton. Under the Basin and Ridge Province in western North America, there appears to be (see above) a buried continuation of the East Pacific rise, which is similar in present-day kinematics to the rift-transform system in the Gulf of California, but one disharmonic with respect to the N-Q upper crustal structures of the Province [103]. The influx of hot mantle material into the buried uplift region is reflected in thermal anomalies and in igneous-rock petrology and geochemistry, as well as the reduced mantle velocities. At the same time, it may cause compression and buckling in the upper mantle of the craton. The latter is seen as elevated velocities in the upper mantle, as well as in the northeast-southeast orientation of the axis of maximum present-day compression in the upper crustal plates, which can attain high values [29].

Compression in upper crustal continental layers is also indicated by the vertical components of the displacements on most active faults in continental mobile belts being of reverse or overthrust type [103, 104]. Recent detailed investigations show that gently dipping N-Q uplifts occurring outside mobile belts in the Russian platform are of folding type, whereas the faults bounding and dissecting them are reverse or reverse-shear faults [92].

This suggests that the present-day compression and thickening in the continental lithosphere may be related at least partially to transformed oceanic lithosphere moving under the continents. Only a few continental regions show different dynamic settings, particularly the East African rift system and the rift zones in the Baykal area, in the North American Basin and Ridge Province, and certain others. The lower horizons in the tectonically thickened continental lithosphere are at relatively high pressures and temperatures, where they acquire features found in the underlying mantle, and they mix with it and are involved in its displacement. Therefore, in the global system, the accumulation of lithosphere material in some places (mainly oceanic rift zones) is almost completely balanced by the return of some of the lithosphere material to the underlying mantle, and this occurs not only under active margins and in collision zones but also in other parts of the continents.

3.3. Local neotectonic systems

The processes in the global system act as external factors for local ones (Fig. 10); the mechanism and features for some of them have been described in detail, e.g., volcanotectonic and gravitational tectonic systems. There are also systems produced by tectonic stresses arising from phase and mineral transformations [9]. Some domal uplifts are related to granitization. Granulite and especially eclogite metamorphism in the lower thickened continental crust leads to densification, subsidence of orogenic zones, and development of basins. Such descending motions may have been responsible for depressions such as the Pannonian basin or the Lut block, which were subject to Alpine folding and orogeny before later downwarping. Fluid-gas activity has also played a role, with the local effects dependent on the tectonic conditions.

Erosion of uplifted areas and transport of sediments to basins disrupt isostatic equilibrium and cause compensating lateral flow at depth [69]. Consequently, the uplifts rise farther and the depressions sink. This is accentuated by dedensification during ascent (because of cracking, phase transformations, and near-surface oxidation) and densification during downwarping (by, e.g., lithification and transformation of clay minerals). The compensating deep flow has a dynamic effect on the overlying rocks and causes additional extension in depressions and compression in uplifted parts.

These compensating flows may occur at various levels in the layered lithosphere; in the Palmyrides of Syria, where the dimensions of the linked anticlines and depressions are a few kilometers, the isostatic compensation occurs at depths of 2-3 km in the plastic gypsum-bearing Triassic horizon and locally is accompanied by diapirism in the anticlines. In the Karategin Range (in South Tien Shan), the crustal low velocity zone swells to 10 km in contrast to that beneath the adjacent valleys of the Ilyak and Kafirnigan rivers, but in the Tien Shan as a whole, there is thickening in the lower crustal horizons and corresponding greater depths of the Moho under the ranges than under the adjacent intermontane and premontane depressions [96]. These thickness changes are due at least in part to deep isostatic displacements. Artyushkov [6] [*sic*, 7?] has demonstrated from the postglacial uplift in the Baltic shield that isostatic compensation occurs at the asthenosphere level in more extensive and gentle domal uplifts. The scale and contrast in the N-Q vertical movements in linked uplifts and depressions are thus related directly to the depth of isostatic compensation of exogenous transfers of material.

There are also very small tectonic systems due to strictly local factors, such as those causing unloading joints. The following explanation has been suggested [78]. Stresses in subsurface rocks, particularly those due to bulk compression, relax during erosion; vertical stresses normal to the surface particularly relax, and this increases the importance of horizontal ones, which give rise to joints subparallel to the surface, which form a type of dislocation layering.

Many processes convert various forms of energy to elastic form as a result of the close and long-acting links between system elements, and thus make the energy available for structure development. There are thus multiple elastic-energy sources distributed at various levels ranging from submicroscopic to global [68]. The total contribution from such sources governs the stress pattern responsible for structure development. In that sense, most structures are polygenic. This has given rise to various views on the causes and sources of tectogenesis, not only for particular structures but also for regional and global combinations. Usually, this is not a conflict of mutually exclusive views but the result of isolated consideration of particular process groups in a single system.

I now turn attention to the interaction between neotectonic systems with reference to the structures of the Peter I and Trans-Alai Ranges at the junction between the Pamir and Tien Shan [85]. These ranges developed from trenches filled with Mesozoic and Cenozoic sediments. Mountain formation on the whole followed folding, although some aspects were synchronous. It began when the folding and the related dynamometamorphism had largely homogenized the sediments and made them similar in deformation parameters to basement rocks. The orogenic folds were combined with tectonic thrust sheets, and on large scales are morphologically similar to the folded structures of the previous stage. Both are due to horizontal compression in the basin resulting from collision between the Pamir and Tien Shan, which in turn reflects the collision between the Indian subcontinent and Eurasia, one aspect of the global system.

The collision between the Pamir and Tien Shan was the external factor that caused various local processes. Ramberg [79] [*sic*, 75?] and Luk'yanov [95] gave data that Scobelev [85] used to conclude that the unevenly distributed N-Q molasse, which accumulated more extensively at the margins (near the eroding Pamir and Tien Shan arches) was unevenly thrust onto the underlying plastic Cretaceous and Paleogene rocks and flowed over them toward the center, which predetermined the axial uplift in the compressed trough. Later, when uplift was prominent in the relief, erosion and accumulation of the erosion products in adjacent residual troughs led to isostatic compensation, because the troughs were filled partly by sedimentation and partly by gravitational flow.

The compensation acted in the same sense as the growth of the uplift from lateral compression and flow of the compressed Cretaceous and Paleogene rocks. One can say that the same endpoint was attained in different ways.

In all these cases, large systems have been produced by stresses accumulating from large closure areas and have activated more local mechanisms. However, examples of the converse can be given. In the Tien Shan N-Q structure, the system formed by basement folds and related faulting is accompanied by extended lineaments oriented obliquely, or sometimes transversely, to the strike of basement folds. Makarov [49] found that the lineaments are a surface expression of the deep neotectonic crustal parting. At the surface, the lineaments are often associated with zones of jointing and small faults morphologically similar to elements in the planetary fracture pattern recognized by Shul'ts [84], with the lineament set in plan reproducing the planetary fracturing on a local scale. The joints appear due to local stress relaxation on very small scales in regions of structural-link closure. However, they are related to lines of deep division in the crust, so they have the same orientation over considerable distances and thus together form the tectonic system in the larger structures (lineaments), which are comparable in scale with the basement folds and are related to them by major faults. Thus energy accumulates at very local levels and causes deformation and failure in structures of substantially larger scale. There are differences in origin between the lineaments and the faults associated with the basement folds, but they partly coincide; a lineament may coincide with a fault if it has also been predetermined by the deep neotectonics, and a fault may use a lineament as a zone where the rock strength is reduced.

Tectonic systems develop by approaching equilibrium, but the latter is attainable only in an ideally isolated system. Rocks in their natural state represent open systems, which exchange energy and matter with their environment. Therefore, thermodynamic equilibrium is never attained in a geological environment.

* * *

The N-Q and active history of the lithosphere is determined by interactions in a set of neotectonic systems differing in scale and structuring. The decisive feature is the global neotectonic system, in which the lithosphere is displaced and partially rejuvenated on the scale of the entire earth. The global system includes structuring effects by lithosphere thickening in certain regions (mainly oceanic rift zones) due to ascending products of mantle differentiation, the transport (at rates varying with level) of oceanic lithosphere from such regions toward the adjacent continents, its transformation at active continental margins, and, as a result of all this, lithosphere compression and thickening within the continents and at the active margins, which involves the lower parts of the thickened lithosphere in subsequent mantle displacements.

Processes in the global system cause local systems of smaller scale, but sometimes very well expressed in their particular features, which may arise from strictly mechanical factors or from phase or mineral transformations. The local systems give rise to processes on various scales ranging from the fairly large (volcanic, gravitational, and isostatic tectonic aspects, or uplifting and downwarping arising from dedensification or densification at different levels in the lithosphere) and range onward to strictly local or even macroscopic deformation and failure forms related to extremely small-scale inhomogeneities and changes in the stress patterns.

Progressive mantle differentiation provides the energy source for the global system and correspondingly for most of the local systems it produces. There is probably an additional source in the upper layers from radiogenic heat released by the differentiation products. A particular case is the neotectonic system involved in compensation for disturbed isostatic equilibrium. On the one hand, this can occur only from nonuniformities in surface relief related to the global system and some local systems such as volcanotectonics. This is predetermined by the energy sources in those systems. On the other hand, the lithosphere movements occurring in isostatic tectonics produce exogenous redistribution, which involves solar energy to a considerable extent, and thus the latter is also a source of tectonic processes. This is not the only way it influences tectonics. There is also supergene mineral formation and the formation of sedimentary rocks, where endothermic reactions occur that store solar energy and predetermine the scope for its release in subsequent transformations, as in clay minerals [11, 44] buried in subduction zones.

There are several groups of factors responsible for these tectonic systems in the N-Q stage; the first consists of the present-day structure and physical properties in the lithosphere, namely the sediment cover, which is subject to folding, and the brittleness in the upper continental crust, which results in extensive faulting; also the variation in physical properties in the lithosphere with depth, which facilitates the tectonic layering, and, finally, there is the fairly large average thickness for the continental crust, which can thus buckle and thicken under compression and produce tectonic forms seen in the relief. The second group consists of the differentiation rate and the distribution of the products from it in the upper parts of the earth. In the N-Q period, this distribution has been such as to tend to move the lithosphere toward the continents, which has maintained mountains at extensive and high levels. The third group is exogenous, where importance attaches to the erosion rates in continental uplifts and the rates of debris transport, which are related to climatic differentiation as well as tectonic features, the extensive occurrence of arid conditions, or the reduced biomass volume associated with cooling in polar and temperate latitudes. The effects from exogenous factors determine the importance of isostatic tectonics in the current relief.

6. The Course of Development of Neotectonic Processes

6.1. Tectonic movement regimes

Existing methods of recording tectonic displacement in the past are based on variations in facies and thicknesses and on paleomagnetic reconstructions; they allow determination only of rates averaged over long intervals. However, there are formations arising with variable tectonic rates, with the changes occurring rapidly and repeatedly, such as flysch series and dike-type volcanic vents. The findings discussed above on structural stresses related to tectonic systems suggests that the stress pattern is highly variable and that the response is the same. To elucidate such variable effects and the various structures associated with them, it is necessary to examine regimes of tectonic motion.

It is not possible to determine the motion regimes in the geological past with adequate accuracy, so there is considerable interest in current movements as examined by geodetic and seismological methods. For example, geodetic observations have been made over many years on the junction between the Pamir and the Tien Shan, in the regions around Garm [31] and Fayzabad [108], frequent changes in displacement rate where have been found for leveling markers, often with changes in sign, which must be summed to yield the long-term trends. The trend in the annual displacement is less in magnitude, sometimes by a large factor, than the individual oscillations.

However, in many active zones, the periods covered by instrumental observations are insufficient to indicate the true movement picture. It is necessary to examine a longer interval: the Holocene or sometimes the entire late Pleistocene. This can be done by geomorphological methods [19, 103], with the most interesting results obtained for strike-slip zones. These show less effect from local factors causing irregular oscillations transverse to the direction of motion than any other type of fault, so the general trends are seen more clearly. Such investigations of Holocene and Recent displacements on faults has shown (in so far as the accuracy permits) that during the last few thousand or tens of thousands of years, the displacements have been in a single sense, but with rates varying time. Three states are distinguished: pulsed, pulse-creep, and creep.

The pulsed state is characterized by rare displacements in catastrophic earthquakes ($M \geq 7.5$), where a displacement of several meters occurs in a fault zone hundreds of kilometers long in an almost instantaneous fashion. Such a pulse is preceded by a period of quiet lasting from a few hundred years to thousands or more, during which no appreciable displacement occurs.

The total displacement in the pulse-creep state is also made up mainly of shifts during strong earthquakes ($M \geq 7.5$), but a more or less considerable proportion occurs in weaker shocks and in places as slow movements (creep).

The periods of relative quiet between strong pulses are usually not so long as in the first. In the Pacific island arcs, which show pulse-creep movement, those periods do not exceed 200 years [25].

Continuous slow movement predominates in the total displacement in creep, as in the Surkhob-Ilyak fault zone at the northwestern margin of the Pamirs [103], where the Surkhob overthrust has a current creep rate of 2 cm/yr, with considerable temporary oscillations [66]. Similar oscillations occur in the right-lateral offset of the Calaveras fault zone in California: in 1910-1929, the creep rate was low, but it then rose to 0.8 cm/yr, and after 1961, to 1.7 cm/yr [80]. Weak earthquakes are very frequent in these creep regions, whereas moderate quakes occur at intervals of decades. On the part of the San Andreas fault in California between Los Gatos and Cholame, where such motion occurs, creep has accelerated for several years before and after the moderate earthquakes ($M = 5$-5.5) in July-August 1966 [3] and after moderate ones ($M = 4$-5) in 1971-1973 [71].

Such movements may occur on an entire active fault or only part of it; correspondingly, one distinguishes faults with the same regime of Holocene and Recent movements throughout their length (synchronous development) and ones where the parts develop differently (asynchronous).

The Khangay fault in northern Mongolia (Fig. 13a) is a typical synchronous one, along which there are Late Quaternary left-lateral offsets extending for 450 km. On July 23, 1905, the Khangay earthquake occurred, with a magnitude of 8.7, in which there was left-lateral offset of up to 5.5 m on a 370-km part of the fault (Fig. 13b). The displacement covered over 80% of the fault. There have been no signs of activity on the fault since 1905.

To establish how far such catastrophes are characteristic of that fault zone, a 15-km segment on the northern flank of the Dagan Del ridge was used in determining the amplitudes for the Holocene displacements on all the gullies, ravines, and other forms of relief intersected by the fault. The histogram (Fig. 13c) shows that the displacements are 5.5 ± 0.5, 11 ± 1, 16.5 ± 1.5, 22 ± 0.5, 28.5 ± 1.5, 33 ± 1, 40 ± 1, and 45 ± 1 m, while intermediate values are few or lacking. The first of these peaks corresponds to the offset in the July 23, 1905 earthquake. The other peaks represent the sum of it with preceding pulses related to similar seismic events as reflected in the relief forms that existed at those times. The movement in each earlier pulse increased the total displacement by about the same amount of 5.5 m, i.e., the geological effect and

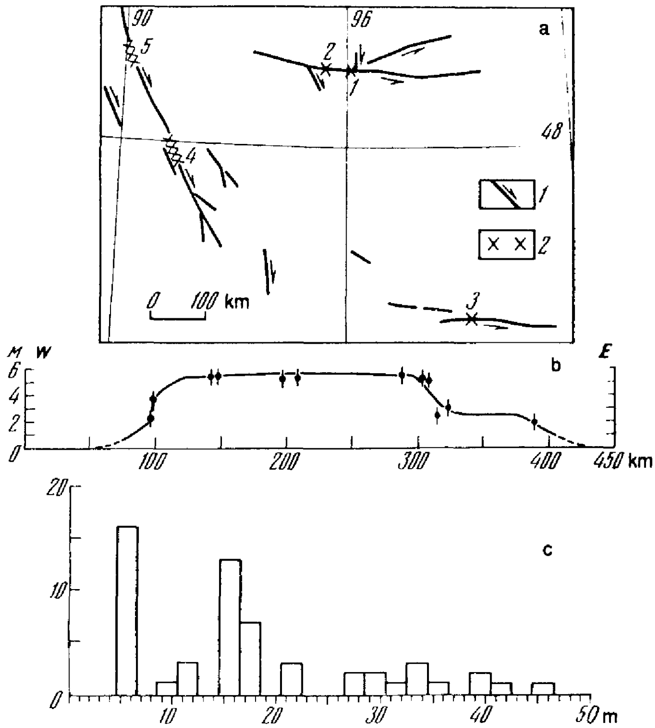


FIGURE 13. Activity characteristics for Khangay fault, northwestern Mongolia: a) faults in western and northwestern Mongolia active in the Late Pleistocene and Holocene [1-5) active faults, left reliably established, right supposed: 1) strike-slip, 2) overthrusts and reverse faults, 3) normal faults, 4) rifting, 5) with unknown displacement directions]; 6) detailed observation areas: A) Khangay fault, northern flank of Dagan Del ridge, B) Khangay fault southeast of Dzun Khangay, C) Dolinoozero fault to the east of Ulan Bulak, D) Kobda fault between the valleys of the Dund Us and Tsagan Burgas Gol, E) Kobda fault in the region of the valley of the Khavtsalyn Gol and the Chikhteyn Bulak spring; the numbers on the map denote the numbers of the sections shown in Figs. 14 and 18c; b) displacement distribution for earthquake of 1905 along the Khangay fault (abscissa extent of fault from west to east, ordinate amplitude of sinistral offset in 1905 in m); c) Late Holocene sinistral offsets amplitude distribution in small forms of relief on a 16-km section of the Khangay fault on the northern flank of the Dagan Del ridge (abscissa amplitude in m, ordinate numbers of displaced streams and other relief forms).

probably the energy parameters in the ancient earthquakes were comparable with those in 1905. A check was made on another 10-km segment southeast of Dzun Khangay. Most of the peaks in the Holocene displacements were confirmed.

In some places, the Khangay fault deviates from the general east-west trend to a northeast-southeast direction, and in these segments there is an extended component to the displacements; one finds grabens and normal-faulted steps, which are associated with closed depressions and deflection in brooks and gulleys. During quiet periods, these steps are cut into by the brooks, and the depressions are filled with clastic material from the slopes. After the next pulse, the steps are rejuvenated and the closed depressions deepen. These often becomes small lakes, in which fine clastic material rich in organic matter is deposited. The age of the organic matter can sometimes be determined by radiocarbon methods and is close to that of the seismic pulse. For example, the small lake Urtyn Nur on the eastern part of the fault occupies a graben-type depression, which has influenced a brook with a system of normal faults, and where signs of the 1905 earthquake are accompanied by marshy beds related to the four previous pulses. The radiocarbon ages have been determined by L. D. Sulerzhitskiy at the Geological Institute, USSR

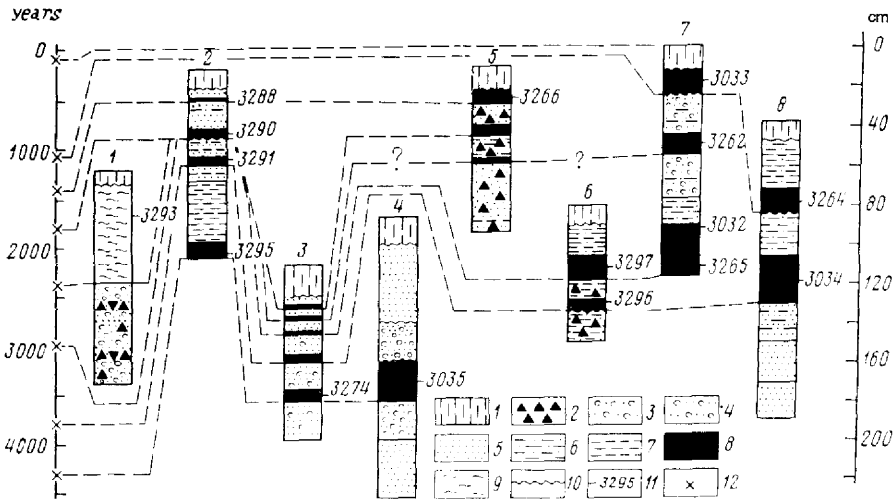


FIGURE 14. Comparison of sections in excavations on flanks of small dammed lakes and closed depressions in the Khangay fault zone: 1) soil layers; 2) rock rubble; 3) grits; 4) coarse-grained sand; 5) medium-grained and fine-grained sand; 6) loam; 7) sandy loam and clay; 8) loam or clay, less often sandy loam, rich in organic matter; 9) peat bed; 10) erosion surface; 11) radiocarbon ages determined by L. D. Sulerzhitsky at the Geological Institute, USSR Academy of Sciences, in 1982: 3032 = 2690 ± 110, 3033 = 1090 ± 50, 3034 = 3720 ± 160, 3035 = 4280 ± 250, and in 1983: 3262 = 2370 ± 80, 3264 = 920 ± 60, 3264 = 2990 ± 90, 3266 = 1300 ± 250, 3274 = 4210 + 80, 3288 = 1400 ± 100, 3290 = 1780 ± 200, 3291 = 3870 ± 180, 3293 = 2360 ± 100, 3295 = 4340 ± 20, 3296 = 3280 ± 180, and 3297 = 2950 ± 150; 12) supposed major earthquakes on time scale.

Academy of Sciences: 920 ± 60 to 1090 ± 50, 2380 ± 80, 2690 ± 110 to 2990 ± 90, and 3720 ± 160 yr. The Urtyn Nur successions may be compared with those in other graben-type and lake basins in the fault zone (Fig. 14), which suggests that the pulses similar to the 1905 earthquake occurred in the Khangay fault zone about 1050, 1400, 1800, 2400, 3000, 3800, and 4300 years ago. Thus the mean recurrence interval is 600 yr, which gives an average displacement on the fault of 0.9-1 cm/yr.

Similar Holocene movements occur on the sublatitudinal Dolinoozero left-lateral strike-slip fault in the Gobi Altai (Fig. 13a), where the last major earthquake (M = 8.3) occurred on December 4, 1957, and in which a left-lateral offset and oblique displacement of up to 5 m occurred along a 270-km segment [47]. In a 12-km segment to the east of Ulan Bulak spring, where the displacement amplitude in 1957 increased from west to east from 2.7 to 3.3 m, measurements were made on all displacements in the gullies and other young relief forms intersected by it. Displacement peaks were found differing by 3.0 ± 0.5 m (Fig. 15). The displaced-gully morphology suggested that major earthquakes occur more frequently on this fault than i the Khangay zone.

Somewhat different but essentially similar Holocene movements occur on the 1400-km North Anatolia zone of sublatitudinal dextral strike-slip faults. Although in many parts of the zone, there are Late Quaternary dextral offsets, there were no signs of movement for several centuries before the beginning of the 20th century [4]. In 1912, there was a strong earthquake in the west of the system immediately north of the Dardanelles [2]. Later, there were strong earthquakes in 1939, 1942, 1943, 1946, 1953, 1966, and 1976. The displacements occurred over distance ranging from 15 to 280 km (Fig. 16), and the movements produced a total dextral offset up to 4 m, with the southern flank of the zone rising by up to 1 m [61, 101, 114]. In its geological effects, this series was a pulse analogous to that in 1905 in the Khangay zone but extended over 64 yr.

In an asynchronous zone, in contrast to the above two, the movements accelerate at different times and with different periodicities. The time and type of movement in such parts enable one to distinguish three groups. Parts in the first group differ only in the times of the main pulses while showing similar pulse-creep behavior throughout the zone. In the second, the parts differ in mode of motion. In these two groups, the Late Holocene movements occur in some way throughout the fault lengths, whereas in the third group, the movements cover only part of the zone. Other parts currently show only slight displacements but are characterized by large movements in the Early Holocene or Late Pleistocene. Therefore, one suggests that long periods apply here (thousands or tens of thousands of years), during which the parts of largest displacement migrate.

Examples of the first group are provided by island arcs and active continental margins in the Pacific;

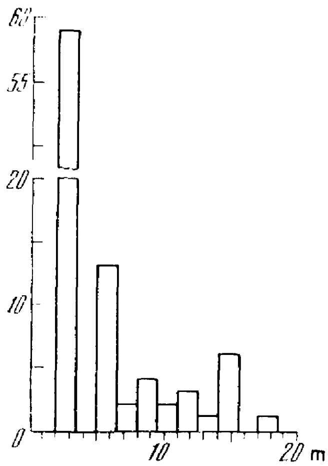


FIGURE 15. Late Holocene sinistral-offset amplitude distribution for minor relief forms on a 12-km segment of the Dolinoozero fault to the east of Ulan Bulak, southwestern Mongolia (see explanation to Fig. 13c).

most of the seismic energy at depths down to 70 km in each part is released in earthquakes having $M \geq 75$, and these are associated with the largest surface displacements (some meters) The segment involved in a major earthquake is usually 100-300 km long*. Fedotov [25] estimated the recurrence interval as 140 ± 60 yr for major earthquakes in each such segment in the Kuril-Kamchatka and Japanese island arcs. A similar 100-200 yr applies for other active structures in the Pacific environment. In the periods between major earthquakes, there are weaker ones, and possibly creep. Major earthquakes occur successively in different segments of the arc or margin. Some success has been obtained in forecasting future major earthquakes from the sequences in the various segments, which indicates that the current tectogenesis conditions are uniform. In essence, such segments differ only in seismotectonic cycle phases. It is not entirely clear whether these segments are stable over long periods, e.g., the Holocene. The data on historical Japanese earthquakes cover almost 1500 years [25] and to some extent indicate a stable situation.

The second group is represented by the 1000-km San Andreas fault (Fig. 17a). Holocene and even historical movements are known throughout the fault, but the parts differ in magnitude and time distribution [1]. There are two segments showing occasional major earthquakes ($M \geq 8$): the northern from Cape Mendocino to Los Gatos (400 km) and the southern from Cholame to Cajon Pass (308 km). In the

*In the Chile earthquake of 1960 (8.5) and the Alaska earthquake of 1964 (8.4), the movements extended to segments having lengths of 800-1000 km, while the displacements attained 20 m [67]. However, such events are exceptional.

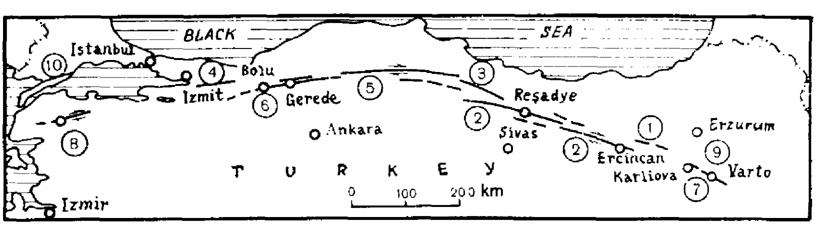


FIGURE 16. Parts of the North Anatolia fault zone on which movements have occurred during earthquakes: 1) November 21, 1939; 2) December 26, 1939; 3) December 20, 1942; 4) June 20, 1943; 5) November 27, 1943; 6) February 1, 1944; 7) May 31, 1946; 8) March 18, 1953 (1-8 from [61]); 9) August 19, 1966 [144]; 10) 1912 [2].

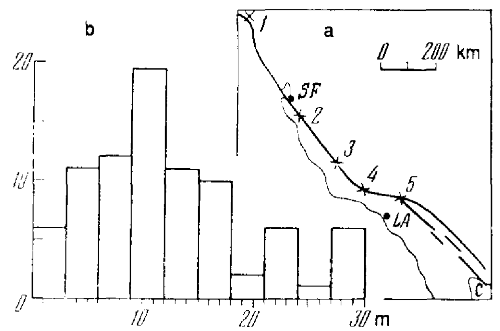


FIGURE 17. The San Andreas fault (a) and Late Holocene dextral offset amplitude histogram shown by small streams on a 110-km segment between Cholame and Camp Dix [115]; b) see explanation to Fig. 13c; 1) Cape Mendocino; 2) Los Gatos; 3) Cholame; 4) Damp Dix; 5) Cajon Pass; SF) San Francisco, LA) Los Angeles, G) Gulf of California.

southern one, the Fort Tejon earthquake of 1857 was a major one, with a right-lateral offset along the fault of up to 9-12 m [115]. After the earthquake, there was a period of quiet, during which observations over 30 years from triangulation networks revealed signs of creep [53], while movements in 1959-1973 using creepmeters showed a very low strain rate, or none at all in places [71]. The seismicity in the area is now extremely low [14].

The histogram for right-lateral offsets in gullies between Cholame and Camp Dix (110 km) shows not only a peak at 9-12m related to the 1857 earthquake but also peaks at 15-18, 21=24, and 27-30 m (Fig. 17b) probably related to earlier major earthquakes [115]. Each of them produced a right-lateral offset of ~6 m. Detailed studies have been made [90] on clastic sediments containing peat beds in the upper part of the succession in terrace 1 in the Pallet Creek valley, which have been produced by landslides on the fault line. Radiocarbon ages indicate that the sediments began to form more than 1400 yr ago and were completed not much more than 100 yr ago, when the pond was drained. During that time, there were repeated movements, which disrupted the beds that had accumulated up to that time. Later beds covered the displaced ones. The relation between the faults and the disrupted and overlying beds serve to date nine major earthquakes such as that of 1857. The periods between them varied from 50 to 300 yr, with an average of ~160. A similar period for major earthquakes is found in the northern segment [77, 83], where the last event of that type was the San Francisco earthquake of 1906. The two segments thus have typically altered regimes.

Creep occurs in two other parts of the San Andreas fault (from Los Gatos to Cholame and to the southeast Cajon Pass), where there are slow dextral offsets accompanied by movements in earthquakes of various strengths. The creep rate in the first segment, the northwest one, ranges from 2 to 4 cm/yr. The largest earthquakes here do not exceed magnitude 6. In the earthquakes in July-August 1966, M up to 5.5, there was up to 18 cm displacement along the fault [13]. The recurrence interval for such events is a few decades, and then the overall seismogenic displacement is less than the tectonic effect from the creep. The second, southeast, part has a more complex structure. Here the fault zone consists of several branches, which differ in seismic features. Earthquakes with M up to 7.1 have been recorded, which have caused displacements of tens of centimeters, or in some case a few meters. The fault segments involved in such movements are much smaller than in faults of the Khangay type. For example, the largest recorded offset was up to 5.5 m and had a vertical component of up to 1.2 m, which occurred in the earthquake of May 18, 1940 on the Imperial fault, but it extended only to a segment of 70 km [79]. After the earthquake, the movement continued in creep form with a rate of 3 cm/yr [53].

The third group is represented by the 950-km Kobda dextral reverse-strikeslip fault in Mongolian Altai (Fig. 13a), where I have examined in detail the northern (300 km) and the central (210 km) segments. The central segment shows evidence of a comparatively recent major earthquake, which produced a dextral offset of up to 5 m (Fig. 18a). A preliminary age has been determined from the fact that the fault has displaced graves dating from the sixth to eighth centuries by up to 4 m at Ar Khutel Pass [38] but does not affect the Late Mongolian graves and modern gullies on the fault near it. Drilling in the valleys of the Buyantu Gol and Tsagas Burgas Gol has reached alluvial and lake-marsh loam beds formed by valley blocking due to displacements on the Kobda fault, which contain much organic material (Fig. 18c). In both sections, the upper loam bed is of age 460 ± 100 yrs and is evidently related to the last movement, which thus

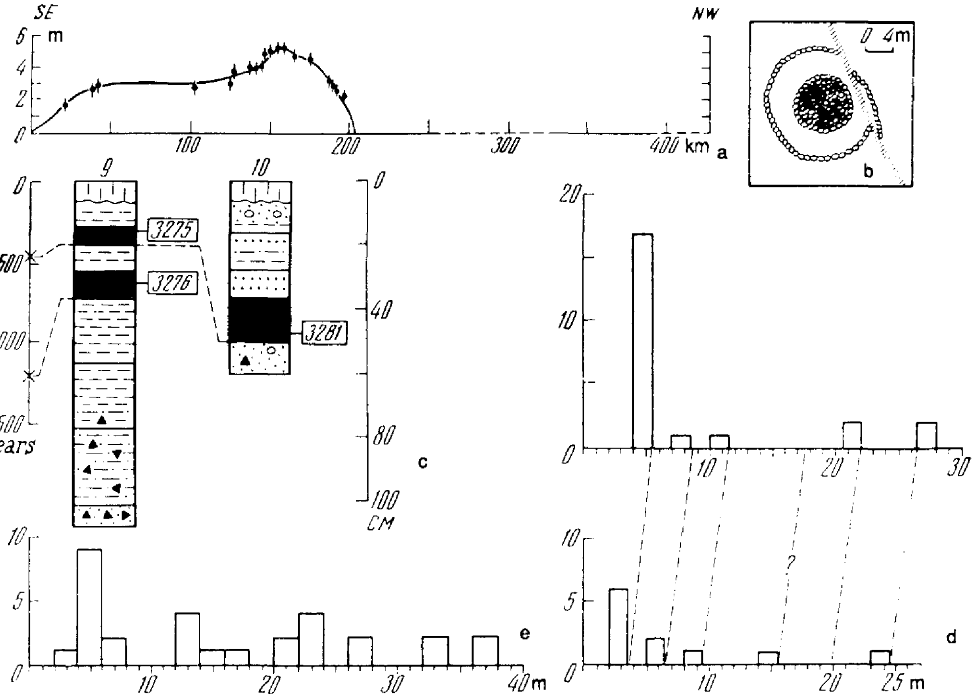


FIGURE 18. Activity features for Kobda fault, Mongolian Altai: a) offsets in earthquakes of early 16th century (?) along fault (see explanation to Fig. 13c); b) displacement of a Turkish (?) grave directly south of Ar Khutel Pass; c) comparison of sections in valleys of the rivers Buyantu Gol (No. 9) and Tsagan Burgas Gol (No. 10) showing radiocarbon ages determined by L. D. Sulerzhitsky at the Geological Institute, USSR Academy of Sciences, in 1983; 3275 = 460 ± 140, 3276 = 1190 ± 80, and 3281 = 460 ± 100; other symbols see Fig. 14; d) Holocene dextral offset amplitude distribution given by small streams and other relief forms between the Dund Us and Tsagan Burgas Gol valleys; the peaks do not coincide because the lower histogram is derived from the segment of the fault north of the Buyantu Gol River, where the seismogenic shifts from the early 16th century (?) gradually decrease to the northwest from 3.5 to 2 m; e) Early Holocene dextral offset amplitude distribution for small relief forms in the valley of the Khavtsalyn Gol and Chikhteyn Bulak (see explanation to Fig. 13c).

occurred approximately at the start of the 16th century. The previous pulse was 700-750 yrs before that. Major earthquakes are probably rarer than in the Khangay zone and particularly in the Dolinoozero fault. In many valleys, the relief forms are not of a single age (ancient beds, first terraces, fans) but are displaced by identical distances of up to 5 m, which also indicates relative rarity of such earthquakes. The histograms for the Holocene displacements in the central segment (Fig. 18d), however, show that strong pulses have occurred repeatedly and were the main form of movement on the fault. The amplitude ranges from 3 to 6 m with an average of 4.5.

In the northern part of the fault, there are no signs of major Late Holocene movements; the earlier displacements form peaks indicating pulsed or seismogenic movement (Fig. 18e). The amplitudes in these pulses on the whole are larger than in the later ones in the central segment: 5.2 ± 2 m. One has the impression that the lack of substantial Late Holocene movements is balanced by elevated activity in an earlier stage.

A similar situation is found in the latitudinal Gobi-Altai fault system in southwestern Mongolia where the Dolinoozero fault developed actively in the Late Holocene (pulsed), whereas more westerly faults in the system show only slight movements in the Late Holocene but were extremely active in the Early Holocene and Late Pleistocene.

More complicated relationships occur in different parts of the Talas-Fergana dextral strike-slip fault in the Tien Shan; here Holocene movements have occurred over a distance of 400 km, all of them being dextral offsets with very much subordinate vertical components. The fullest characteristics have been obtained in the southeastern and central parts [94], where from southeast to northwest one can identify six different segments: from the Kokkiya Pass to the upper reaches of the Pychan River (25 km), from the latter to the Kyldou valley (25 km), from the latter to the valley of the Kuroves River (30 km), in the basins of the Kuroves and Keklikbel (25 km), in the basin of the Karasu East River and onward to the Toktogul reservoir (75 km), and from the latter to the upper reaches of the Chatkal River (70 km).

The offset histogram for the first segment has 11 or 12 peaks in the amplitude range up to 50 m (Fig. 19a), which shows that pulses are dominant. The amplitudes in the individual pulses range from 3 to 7 m with an average of 4-4.5 m. The most representative offsets are 6-7.5, 15-17.5, 20-22, 27-28, 35, and 40-41 m. All these, apart from 15-17.5 m, have analogs in the second section (Fig. 19b), but the amplitudes in the corresponding maxima are there larger by 10-20%, which probably reflects an increase in intensity under similar conditions. The rates in the two parts have been deduced from sections in the adjacent depressions produced by valley damming by the displaced gully flanks or terraces formed during the offsets. The sediments at the base of the depression sections are closest to the times of movement and give the shearing rate in the first section as ~0.5 cm/yr (sediment age 3740 ± 40 yrs* with gully offset 19 m), and in the second, ~0.7 cm/yr (3740 ± 600 yrs, offset 27 æ 1 m) [94]. Three radiocarbon samples have been taken from higher horizons in depressions in the second (3150 ± 40 yrs with offset 90 ± 3 m, 2640 ± 600 yrs with 25 ± 1 m, and 2320 ± 40 yrs with 23-24 m), which do not conflict with that estimate. The first of these three determinations is interesting in that an offset of 7-8 m was measured in the same gully as arising after the depression was filled, which gives a lower imit to the offsets rate of 0.24 cm/yr.

Late Holocene movements are rare in the third section, and the amplitudes are substantially less; in the fourth and fifth, they again become comparable with the first or even larger. The peakedness in the amplitudes in the fourth section (Fig. 19c) is less pronounced. There the peaks do not have analogs in the first two sections. Either there were other pulses or the creep has been more important than in the southeast zone. The second assumption is more likely, as plastic rocks are abundant in the fourth and fifth sections.

Late Holocene movements occur everywhere in the sixth section. The most interesting results have been obtained in the northwest section, in the region of the Karakul'dzha Pass between the upper reaches of the Chatkal and Atoynok [17]. Here the predominant amplitudes are 30-40 m, and a radiocarbon age from a lower layer in one of the dispersions is 2020 ± 50 yrs, which gives a mean displacement of 1.5-2 cm/yr. It is evident that seismic pulses have been involved because this area is associated with the epicenter of the 1946 $M = 7.5$ Chatkal earthquake.

All six section show offsets over longer intervals in the Late Pleistocene and Holocene; in the fourth, a glacial trench and a bottom moraine dating from the end of the Late Pleistocene (20-25 thousand yrs BP) are displaced along the fault by 135 m, with the southwestern flank raised by 5-6 m, while the terminal moraine exposed somewhat farther to the southeast (Middle Pleistocene) is displaced relative to its glacial trench by 700-800 m [94]. In the upper reaches of the Karasu East River (fifth section), there is also a description of a moraine dating from the end of the Middle Pleistocene displaced by 750 m [76]. These three give the minimum average offset rate in the Late Pleistocene and Holocene as 0.6-0.8 cm/yr, which is the same as or slightly greater than the average Late Holocene shearing rate in the second section.

Geomorphological correlation has been applied to displaced Late Quaternary relief forms to the northwest of the Toktogul reservoir, i.e., in the sixth section, with displaced forms southwest of the reservoir; the amplitude of the overall offsets in the

*These and later radiocarbon determinations from the Talas-Fergana fault zone were made by L. D. Sulerzhitsky at the Geological Institute, USSR Academy of Sciences.

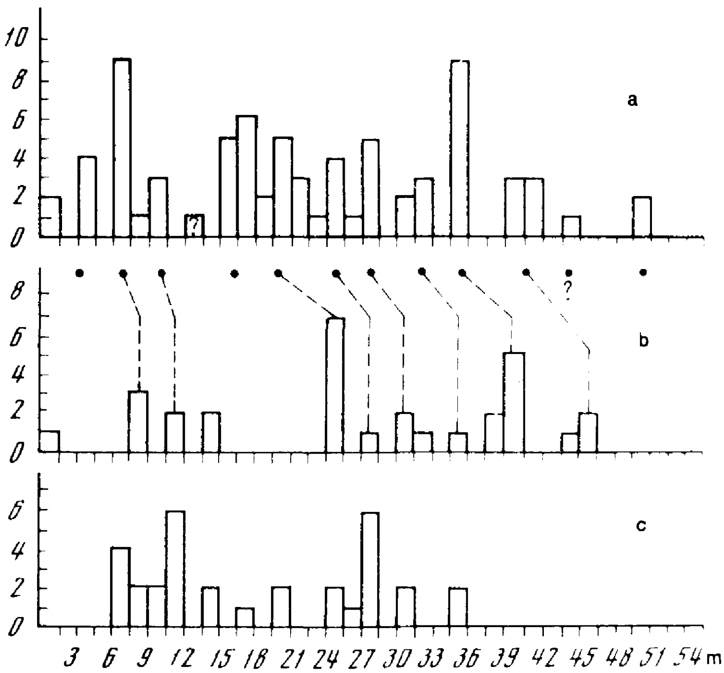


FIGURE 19. Holocene dextral offset amplitude histograms for various segments of the Talas-Fergana fault in the Tien Shan: a) from Kokkiya Pass to the upper reaches of the Pychan River (bold points possible traces of major earthquakes); b) from the upper reaches of the Pychan River to the Kyldou valley; c) in the basins of the Kuroves and Keklikbel (see explanation to Fig. 13c).

northwest direction for forms comparable in age increase by factors of 1.5-8, and so do the rates. Therefore, the tendency for the offset rate to increase the northwest existed in earlier stages in the Late Pleistocene and Holocene, but was less pronounced than in the Late Holocene. The southeastern and central parts of the Talas-Fergana fault was similar to the Kobda fault in that there were differences in Late Holocene offset rate in different parts, which have been compensated by differences in rate distribution in preceding epochs. However, that compensation is incomplete in the Talas-Fergana fault zone: on it is superimposed a general increase in rate to the northwest. Also, the various parts of the fault probably differed in development modes, i.e., larger or smaller proportions of creep in the overall motion.

The Holocene movement features in the various zones have been governed by the stress accumulation rates and the rock properties; with high accumulation rates, substantial rock volumes soon acquire a state of uniform high strain leading to a strong earthquake. Therefore, major earthquakes occur several times more frequently in parts of the San Andreas fault showing pulsed motion than they do in analogous zones in active faults in continental Asia, and the same applies to other fault zones along active Pacific margins.

The motion is governed by the rock strength, as is evident from the compositions and structures in lithosphere volumes making up and surrounding the various types of active zone. All zones around active faults in Asia showing the pulsed mode are in regions with thick continental crust. Major foci are associated with a granite-metamorphic layer, which can withstand high stress concentrations without failure in large volumes. Pulse-creep mode is most characteristic of Pacific-margin island arcs where one gets crust of oceanic and transitional types. With high stress accumulation rates, those ones differ from ones with the pulsed mode in retaining appreciable activity during periods of relative quiet. The approximately equal intervals between major earthquakes in the different parts indicate uniformity in strength, but the length of the segment failing in an earthquake usually does not exceed 300 km, i.e., is much less than the total zone length. The island-arc lithosphere thus does not usually attain a uniform state of high strain throughout the active zone or much of it, but instead fails earlier.

One can compare San Andreas fault sections showing pulsed and creep motion [1]; the first are composed down to seismic depths of granite-metamorphic formations, while the second are composed of the plastic Franciscan complex and serpentinites, with the southern one of the creep areas characterized also by reduced crustal thickness and elevated heat flux. In the first, the active zone is narrow and is represented by the single fault plane, with minimal branching and parallel lines. In the second, such fault lines are numerous and extend over a broad belt, particularly in the southern section adjoining Baja California. The modes are thus governed by the rock competence and degree of crushing.

The Late Quaternary movements in different parts of the Kobda and Talas-Fergana and similar faults places emphasis not only on strength but also state of strain in the rock volumes adjoining the fault; the differences suggest that those volumes are seismogenic, and the active fault zone consists only of the weakened medium where the accumulated stresses begin to be released and the most extensive deformation concentrates. An interesting point is that an earthquake occurred in the Borrego Hills in southern California on April 9, 1968 (6.4-6.5), and in the epicenter zone along the fault at Coyote Creek, a dextral offset of up to 38 cm occurred over a distance of 31 km; at the same time there were dextral offsets of 1-2.5 cm on faults in the southern Superstition Mts. (over a length of 23 km), on the San Andreas fault (30 km), and on the Imperial fault (22 km), at distances of 45-70 km from the epicenter [12]. Clearly, here the generating region is not restricted to the zone of the Coyote Creek fault.

6.2. General features of N-Q lithosphere evolution: tectonic phases and episodes

The N-Q stage may be defined as the epoch with extensive current orogeny and is characterized by the following features, which distinguish it from previous Mesozoic and Cenozoic stages. First, in regions with continental crust, regression is the greatest for any time in the entire Post-Variscan period. Second, there is maximal occurrence of mountain systems far outside regions in which the continental crust is of Alpine age. A third feature, which in part follows from the first two, is that there is considerable relief contrast at the surface in mobile regions. Finally, the fourth feature, also partially related to the previous, is the distinctive climatic situation, with marked zoning, the wide occurrence of arid conditions, and cooling in temperate and Polar latitudes.

These features arose and developed gradually; they can be detected at different times in the various regions [19]. At the same time, there are extensive segments in mobile belts where the neotectonic events are regularly related in space and time, as can be seen clearly by comparing them in the rift zones of northeast Africa and the Anatolia-Caucasus-Iran part of the Alpine-Asiatic orogenic megabelt. The start of calc-alkali volcanism in Iran (55 Ma) coincides with the start of traprock volcanism in Ethiopia. The Red Sea and Aden rift zones were formed ~40 Ma, which coincided with the epoch of folding in Anatolia and directly preceded the vigorous Early Oligocene folding in the Adzhar Trialet zone, at El'brus and more easterly structures in northern Iran. Mountain ranges began to form in these regions. Kaz'min [35] related the events at the end of the Eocene and start of the Oligocene to accelerated collision between Africa-Arabia and Eurasia at 40-35 Ma.

The comparison in the above paragraph can be continued and shows that two types of structure were activated at the boundary between the Oligocene and Miocene, but that activity fell off in the Early Miocene. The latter prepared major structuring that made itself felt at the boundary between the Early and Middle Miocene (~15 Ma). Kaz'min found that the Afar and Ethiopia rift zones developed from that time, whereas the Kenya area of alkali-basalt volcanism arose somewhat later. Folding began in the marginal folds in eastern Anatolia and in the Outer Zagros. At the boundary between the Middle and Late Miocene, the compression in those folded zones was accentuated. The folding was completed and the Crimean Highland began to rise, along with the axial parts of the High Caucasus and Kopet Dag. During the Late Miocene and Early Pliocene, rifting developed in northeastern Africa and orogeny in the Arabia-Eurasia interaction region, where several episodes of accelerated collision can be detected. The start of the last episode, the Early Pliocene one (Post-Pontian to Pre-Akchagyl), coincided with certain changes in the relative displacement of Arabia and the Somali block and with the formation of the Kenya rift zone [35].

It is clear that the synchronism has been predetermined by the coupling between adjacent lithosphere blocks and plates in a single region; however, the different segments show a more general regularity: extensive although perhaps not universal synchronism in major events [19]. For example, at the end of the Eocene to start of the Oligocene, there was a folding movement in the Indus zone (this is evident from surface deformation and the start of molasse accumulation over an extensive area extending to the Tien Shan), which occurred in Iran, Transcaucasia, Anatolia, the Alps, the Pyrenees, the Atlas, and Cuba. The current relief has been formed since that time in certain axial zones in the Alpine-Asiatic orogenic fault.

At the boundary between the Oligocene and Miocene, there were considerable episodes of folding, overthrusting, granitization, and metamorphism in the Himalayas and in the South and Central Pamirs; there was also folding in the Pontian-Caspian part of the Alpine-Asiatic belt, with thrusting and folding in the eastern Carpathians, Swiss Alps, the southern Pyrenees, and the Atlas. The current relief began to form in various parts of Iran and Anatolia. The San Andreas fault zone arose, and downfalling and volcanism began in the Basin and Range Province in western North America.

At the boundary between the Early and Middle Miocene, horizontal movements arose in the Pamirs, and fold-mountain systems began to develop in the northern and northeastern flanks of Arabia, with the Afar and Ethiopian rift zones arising, which transformed the double rift junction in northwestern Africa into a triple one, and the main lines of the neotectonic structures and directions of the N-Q movements arose in western North America.

Further activation occurred in the Late Miocene to Early Pliocene, when most of the current fold mountains in mobile belts were finally established. During that epoch, which lasted 8-9 m.y., there were several episodes of accentuated movement over extensive areas. The most important were those at the boundaries between the Middle and Late Miocene, the Miocene and the Pliocene, and the Early and Late Pliocene. Extensive horizontal movements and folding affected the peripheral zones in the Alpine-Asiatic belt: the North Pamirs, including the outer zone, the zone of the main boundary fault in the Himalayas, Beluchistan, and later the Suleiman and Kirtar mountains, the Outer Zagros, the zone of edge folds in eastern Anatolia, Kopet Dag, the southeastern and northwestern ends of the High Caucasus, the outer zone and the foothills in the eastern Carpathians, and the Jura mountains. At the boundary between the Miocene and Pliocene, the Levant sinistral offset zone at the northwestern margin of Arabia was reconstructed, in the region from the Dead Sea to Lebanon, where this zone existed already in the Miocene, from which it extended to the north and northeast into western Syria and southern Turkey. At the end of the Miocene, and in places throughout the Late Miocene, there was folding, and the current mountain relief began to develop in Kamchatka, Sakhalin, and Japan. In the Late Miocene, horizontal movements were accentuated on the San Andreas fault and the related fault systems, which remain active today. At the start of the Late Miocene, there was a peak in volcanism and rifting in Iceland. The rift zones began to expand in northeastern Africa.

The latter events reconstructed and finally fixed the present-day aspect of the mobile zones, with mountain ranges rising rapidly and intermontane depressions downwarping (in places such as the Tien Shan and the Tadzhik depression, with some acceleration), with increasing relief contrast in the rift zones, and rifting and orogeny involved in the Baykal region as well as West and Central Mongolia. Activation episodes identified in some zones can usually not be traced throughout an entire mobile belt or in different belts, exceptions being the episode at the end of the Pliocene to start of the Eopleistocene, which can be identified in the Tien Shan, Tadzhik depression, Baykalia, and the Pontian-Caspian regions, as well as in the Rhodopes and in Kamchatka, together with the episode at the end of the Middle Pleistocene, which has been detected in Central Asia, the Pontian-Caspian region, in Kamchatka, and in the transverse ridge system in the western USA [19].

These phases and episodes reflect the gradual origin and accentuation of features that distinguish the N-Q stage from the previous Mesozoic and Cenozoic epochs. Each new activation phase transformed the existing origins and involved extensive movements in new and previously stable regions.

The orogens developed in various ways at the sites of Mesozoic to Early Cenozoic troughs and regions where there was relatively ancient consolidated crust. In troughs such as in the eastern part of the Tadzhik depression, folds formed in the sediment cover and gradually became more complicated, more or less disharmonic with respect to structures of the same age in the crystalline basement. During the deformation in the sediments and basement, there was related dynamometamorphism in the sediments and homogenization in the deformable rock masses [85]. The larger mountain uplifts began to develop, as did the related intermontane depressions, whose axes may or may not coincide with those of earlier folded chains.

In areas, such as the Tien Shan, where there was ancient continental crust partly covered by thin Mesozoic to Early Cenozoic sediments, the early stages of the N-Q orogeny gave rise to weakly contrasting and often equant structures, low uplifts and shallow depressions. Later, the contrast in the uplifts and intermontane depressions increased; they everywhere acquired linear features. The Tien Shan ranges rose, while the downwarping in the Mesozoic-Early Cenozoic Tadzhik depressions was accelerated.

In the European part of the Alpine-Asiatic orogenic belt, and in the Caribbean region, there was extensive buckling and orogeny in the initial Eocene-Oligocene stages of neotectonic development, i.e, earlier than in Central Asia. In the later stages, the earlier mountain ranges were partly destroyed. An important role here appears to have been played by rifting and other forms of crustal destruction, which took various forms: linear grabens of the Rhine type, submarine rift-type depressions in the Tyrrhenian Sea, dispersed rifting structures of the type found in the Aegean and adjacent parts of the Balkan peninsula, superimposed volcanic belts of the Transcaucasian type, and finally equant depressions such as the Pannonian basin.

Rifting and crustal destruction set in the fold-mountain belt, which was split up into individual mountain systems, which, however, continued to rise, although less rapidly than the mountains in Central Asia. This is related to the position of at least the eastern part of Alpine Europe on the continuation of the rapidly developing rift megabelt covering the Indian and Pacific oceans and may be compared with the penetration of rifting structures into the North American Cordillera (the Basin and Ridge Province) at the other end of the same megabelt. At the same time, the distinctive neotectonic development in Alpine Europe can be interpreted as the start of a new stage in the orogen: dissection and ultimately destruction. It is notable that this occurs in areas where the most extensive orogeny occurred at the start of the N-Q stage.

The various phases and episodes differ in significance as regards the N-Q appearance of the earth [19]. The activation epoch at the end of the Eocene to start of the Oligocene (peak at 40-35 Ma) can be identified with the Pyrenean folding phase and appeared mainly in the Alpine-Asiatic megabelt, the Caribbean region, and in adjacent areas. As regards tectonic processes, it differed little from the Cretaceous phases of diastrophism. This applies also to the later tectonic phases: Late Oligocene to Early Miocene and beginning Middle Miocene. Each of these involved folding and thrusting in various areas, but not in all the mobile belts, and the areas covered by orogeny were only slightly increased. Up to the end of the Middle Miocene, the tectonic setting in the continents was not unique for the Mesozoic and Cenozoic. Similar conditions arose periodically in preceding tectonic phases. At the same time, from the Eocene to the end of the Middle Miocene, the main elements in the N-Q mobile belts were laid down, which developed actively later.

The N-Q continent structures were decisively affected by events and began at the end of the Middle Miocene and extended to the middle of the Pliocene [19]. This period, from ~12 to 3.5 Ma, gave rise to almost all the tectonic features distinguishing the N-Q stage from the earlier Mesozoic and Cenozoic epochs. Areas of continental transgression were markedly reduced. The mountain systems and rift zones formed and acquired their contrasting relief. Subsidence rates in ocean trenches increased, and from the Miocene, the trenches began to develop as uncompensated troughs, in spite of mountain growth and corresponding increases in clastic input from island arcs and active margins. Orogeny extended not only to the area of Alpine diastrophism and Alpine crustal consolidation but also to many ancient folded structures and even parts of Precambrian platforms. In the Late Pliocene to Quaternary period, those features became accentuated, and in particular mountain ranges grew rapidly and extended to large areas.

These Cenozoic phases were characterized by simultaneous compression in folded belts, stretching in rift zones, and horizontal displacements in shearing systems. However, there were also shorter episodes, which do not fit into this regularity, so far as one can judge from the current dating accuracy. For example, at the very start of the Late Miocene, before the Late Miocene rifting activation in northeastern Africa, in the Afar rift, and possibly in the Red Sea rift, there were brief differential movements, which make themselves felt in angular unconformities between the Middle and Upper Miocene beds. These signs of diastrophism coincide in time with folding activation in adjacent parts of the Alpine-Asiatic belt.

About 4.5 Ma, when extensive compression set in the Alpine-Asiatic belt after the brief Pontian transgression, there was rearrangement in the northeastern African rift systems (some change in the relative displacement of the Arabia and Somalia block and the initiation of the Kenya rift zone), in the Icelandic rift system (the northern part of the former rift became inactive and the northern part of the Mid-Icelandic neovolcanic zone arose), as also in the Central Atlantic rift (change in spreading direction in part of the Mid-Atlantic ridge near 37°N). At the same time, the current structure arose in the transform rifting system in Baja California, which was related to rearrangement in the San Andreas fault zone in southern California. Such rearrangements, at least in Iceland, were accompanied by a decrease in rifting activity. The next episode of diastrophism in the Alpine-Asiatic megabelt and in other fold-mountain belts occurred ~1.8-1.6 Ma at the end of the Pliocene to beginning Eopleistocene, and it also coincided with partial rearrangements in the Icelandic and northeastern African rift systems [19, 35, 103].

Some extensive-compression episodes in fold-mountain belts thus coincide not with active extension in rift systems but with episodes of rearrangement, possibly with decrease in rifting activity or even slight folding. This suggests that in the N-Q stage, there may have been alternating stages of relative compression and extension, as suggested by Milanovsky [55] and Bankwitz and Bankwitz [8]. If such oscillations actually occurred, they were of shorter period than the stages in the general tectonogenesis activation (compression and stretching) and to a considerable extent were suppressed by the latter.

The fold-mountain structures in the Pontian-Caspian region also involved changes in tectonic activity of higher rank, lasting for hundreds or even only tens of thousands of years, which can be established also less reliably in certain other segments in the mobile belts. As a rule, they do not extend beyond such segments and were probably due to local factors, one of which in the Pontian-Caspian region was the more or less prolonged climatically controlled breaks in the link to the ocean [19].

* * *

The main features in the N-Q stage arose gradually and at different times in the various belts; the structures arose unevenly, which can be represented as a wide oscillation spectrum. The highest frequencies accessible by geomorphological methods are provided by pulses on faults involving strong earthquakes, and with $M > 7.5$, one gets virtually instantaneous movements of several meters extending for distances of hundreds of kilometers, which alternate with epochs of decreased activity, and in regions with the most highly consolidated continental crust, even complete tectonic quiet lasting for hundreds of years. The weakest zones show a different mode of motion: creep, which is accompanied by relatively weak and frequent earthquakes. The creep rates may vary by factors of three or more, with increases before and after earthquakes. The Holocene movements have been determined by stress accumulation rates and rock properties, and they serve as a major tool in elucidating the current geodynamics.

Lower-frequency nonuniformities are represented by activation episodes and phases lasting from tens of thousand to a few million years. Episodes lasting 10,000 years or so as a rule occur in individual belt segments and are governed by local features. For many of the longer episodes, and for almost all phases, one can demonstrate a global pattern. The global neotectonic phases are: late Eocene to early Oligocene, from the late Oligocene to early Miocene, from the start of the Middle Miocene, and finally from the Late Miocene to Early Pliocene, all of which have been decisive in the N-Q stage, with the final emplacement associated with the last phase, which was 8-9 m.y. long and combines several episodes, which are similar to the last activation episode at the end of the Middle Pleistocene to Late Pleistocene.

All the phases and many of the episodes show simultaneous increases in process rates in compressional, extensional, and shearing zones, but there are also weaker episodes showing a different type of synchronism: activation in rift zones coincident in time with decreasing activity in orogenic regions, or, conversely, compression in orogenic regions synchronous with reduced rates of rifting and structural rearrangement, and sometimes even slight folding, in rift zones. These episodes may be considered as shifts between compression and extension affecting orogenic and rift belts. However, it is difficult to determine the periodicity in such shifts, because such episodes can be detected only among the youngest formations, because in older ones they have been suppressed by phases and episodes of general tectonic activation. The global extent of many episodes and phases is evident from the relationship between the tectonic processes in the global system.

7. Neotectonics and the Current Situation

Neotectonics is of particular interest because its regularities can be used in interpreting past structures, but the application is limited in that although present-day geological formations contain analogs for most past ones, the relationships and significance now may not have been the same as in the past, because the roles of the individual factors may have shifted depending on changes in the environment, although the laws of nature remain the same. In this section, I define the general features of interpretations for ancient tectonic formations in terms of the actualistic approach, i.e., I seek to elucidate how the various factors have altered or remained the same. I also identify characteristics occurring identically or nearly so in the N-Q and older stages, together with the differences in the effects from tectonic processes resulting from global changes, both directional and cyclic.

7.1. Characteristics that are largely unchanged

One of the largely unchanged characteristics concerns the order of rates of horizontal movement in the Mesozoic and Cenozoic. Banded magnetic anomalies in the oceans and deep-sea drilling indicate definite variations [46], but it seems that they differ from current rates by no more than a factor of two, so the relative rates reflected by N-Q structures can be taken as representative of older Phanerozoic epochs. The N-Q stage differs from the latter not so much in the rates of horizontal movement as in the vertical component.

An important group of characteristics has remained unchanged during the Phanerozoic: the combinations of tectonic forms and faults in the continents, which are called structural patterns. Luk'yanov [47] pointed out the similarities between N-Q and older structural patterns as the key to understanding the origins of the latter. It is often impossible to elucidate the chronological relationships in a structural pattern for a remote epoch, whereas it can be done reliably for N-Q formations and enables one to establish unique origins. The relationship is most evident for the damage done by recent major earthquakes.

One can investigate and interpret continental structures from the past from the general laws for N-Q structural patterns as one precedes from regions with maximum compression, i.e., where there is a shift from folding and overthrusting, with general but depth-differentiated buckling, to simpler forms of buckling and localized crustal thickening, and finally to block structures almost free from folding but with predominant shearing along faults, both cross-cutting and longitudinal in relation to structural forms seen in the relief. For a long time, kinematic reconstructions were made with too little allowance for ancient shearing epochs extending along structural zones, and only neotectonic investigations have demonstrated their true significance.

It is now widely accepted that geochemical and petrological comparisons can be made between ancient volcanic series and N-Q volcanism in various zones in order to elucidate past geodynamic environments: active margins, island arcs, rift zones, etc. However, some care is needed in using that method, which should be combined with other features, because research on N-Q volcanism shows that the features are often much the same in different geodynamic circumstances.

The different frequencies associated with nonuniformities in N-Q structures are important for research on ancient tectonics; pulse movement is now evident as alternation of strong earthquakes with epochs of relative quiet, which in the past may have been related to the formation of flysch suites and possibly sheeted-dike complexes. The lower-frequency features in N-Q history are expressed by global phases and episodes involving activation [19], and this throws additional light on whether those phases and episodes in the past relate to epochs involving general increases in process rates in kinematically inhomogeneous zones or whether events in compressional and extensional zones are activated in antiphase and reflect the alternation of such states in the earth's history [36, 46, 55]. Neotectonic data show that there have been phases and episodes of both types in the N-Q stage, with epochs showing general activation being more prominent and largely concealing episodes involving alternating and extension.

7.2. Consistent trends

The researches for the tectonic map for northern Eurasia [97] indicated a general trend in lithosphere evolution: steady increase in thickness and sialization in the crust accompanying differentiation of interior. The thickening was accompanied by flux of juvenile water to the hydrosphere. The salt composition also probably altered, and in particular, sea water in the remote Precambrian may have been more acid [18]. The CO_2 contents in the atmosphere and hydrosphere are largely regulated by living matter [110], although the component is derived from plutonic sources, and biologic control made itself felt almost from the start of the earth's geological history [118]. Living matter transfers excess CO_2 to sedimentary rocks and has maintained the CO_2 level in the range 0.03-0.4% at least since the Late Precambrian, which has provided a major surface temperature regulator [15, 16]. Living matter has also been involved in forming other sedimentary rocks [110], which has meant gradual growth in the stratisphere.

Consistent trends are reflected in certain sedimentary formations, which were produced most fully and widely, or sometimes solely, only in certain periods [116]. Similar differences for magmatic and metamorphic formations differing in age have been observed. It has not yet been finally established whether such trends changed gradually or whether there were comparatively brief transition periods. At present, most investigators are inclined to the second view. However that may be, it is clear that these trends have affected morphology, kinematics, and structural development when one compares the N-Q formations with the very remote Precambrian structures and geodynamic settings.

For example, the hotter crust in the remote Precambrian was favorable to plastic strain; the smaller stratisphere thickness did not allow folding of the modern type or shearing of the sediment layer relative to the basement. The thin and hotter crust could not have shown such substantial buckling as at present and, correspondingly, could not have produced such high mountains. The reduced relief contrast was also affected by the smaller amount of free water. It is readily shown that isostatic compensation associated with removing all the ocean water would reduce the contrast in level between the current continents and oceans by ~1.8 km. Precambrian conglomerates demonstrate smaller heights nd less differentiation in mountain relief. On the whole, they are much less common among sedimentary series of that age than are N-Q molasses, and they resemble the Witwatersrand conglomerates in being more extensive in area. The mountains were lower and less differentiated, and the generally low relief contrast consequently reduced the scope for exogenous transport and the occurrence of isostatic tectonics.

7.3. Cyclic changes

The largely unvarying processes and the consistent trends provide a background for the cyclic changes, which can be identified clearly in the Phanerozoic in various ways. Those features in the N-Q stage such as predominant regression on the continents, extensive orogeny, pronounced climatic zoning, including glaciations, occurred in certain

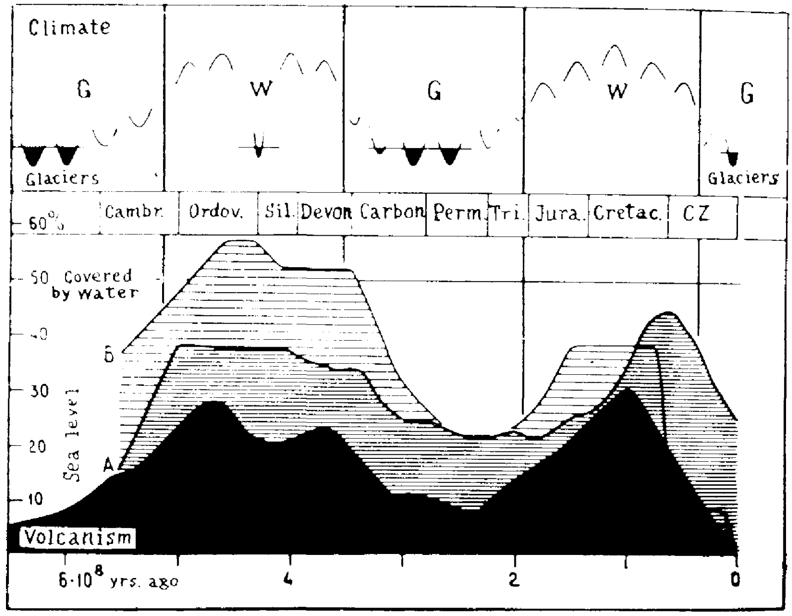


FIGURE 20. Marine transgression correlation scale A: from [109]; B) from [33], magmatism [24] and climatic changes [G) glaciation, W) warming] during the Phanerozoic [27].

other epochs: at the end of the Precambrian, in part in the Early Devonian, and particularly in the Permian and early Triassic (Fig. 2). Along with the N-Q stage, these may be called geocratic epochs.

One can contrast thalassocratic epoch with these, where there were transgressions, reduced transport of mountain erosion products, reduced climatic contrast, and a general increase in mean annual temperature. Several different evaluations have been given [33, 81, 109] for transgressions in various periods, but they agree in that typical thalassocratic epochs involving flooding of more than 40% of the area occurred in the Ordovician, the second half of the Devonian, and the early Carboniferous, as well as in the Cretaceous and possibly the Late Jurassic (Fig. 20).

The thalassocratic epochs differed from the geocratic one in other ways, such as the abundance of magmatic formations, Budyko, Ronov, and Yanshin [15, 16] examined the amounts of continental volcanics erupted in various Phanerozoic periods and found activity peaks in the Middle to Late Devonian and to a smaller extent in the Early Permian, Middle and Late Triassic, and Cretaceous (Fig. 21). The data, however, do not characterize the magmatic activity for the entire earth, because they do not fully incorporate oceanic volcanism and, of course, cannot incorporate the exogenous destruction of volcanics. Therefore, plutonic formations are preferable for such purposes. However, in so far as these are accessible, they characterize not so much the amounts of primary mantle products reaching the upper shells as the secondary reworking at active margins and in collision zones (Fig. 10). It has, however, been shown at least for the Cretaceous and Paleogene that there is a correlation between the spreading rate associated with primary mantle products entering and the intensity of intrusive magmatism at active margins [46]. It may be that the formation rate of plutonic rocks is indicative of total magmatic activity. Calculations for granitoids at the margins of the Pacific and North America (Fig. 20) show maxima for formation in the Ordovician, Devonian, and Cretaceous [24, 27, 46].

Thalassocratic-epoch transgressions are thus related to magmatism [27], which reflects mantle differentiation products entering the upper crust. There are numerous and often indirect links between these process groups (Fig. 22). The most obvious factor is

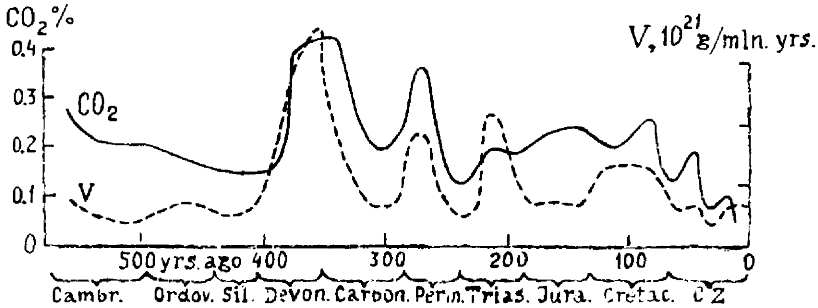


FIGURE 21. Atmospheric CO_2 levels and formation rates of volcanic rocks during the Phanerozoic as derived from continental beds [15].

that additional juvenile water accompanies the silicates, which itself may raise sea level. Another is the position of the continents relative to the entry points of mantle material. In the N-Q epoch, the continents are fairly close together. Eurasia and Africa, close together, constitute 56% of the area of the continents, and in the Late Miocene, they were even closer together than now, because the Mediterranean did not exist in its current form. In the Late Paleozoic, with maximum approach between the Pangaea continents, the closure parameter was even greater. These adjoining continents have the central parts remote from the principal source of erosion (the oceans), which hinders erosion and the removal of clastic material. It may be that the current heights of 1.5-2 km at the center of Eurasia are due not to tectonic uplift but remoteness from the principal base level of erosion.

The current continent grouping is related to the world rift system, with the lithosphere moving toward the continents, which maintains the compression and the elevated state of the latter. The world rift system with features similar to the current ones arose only at the start of the Cenozoic [73]. In the Cretaceous, the spreading zones were more widely separated and numerous. On the one hand, this may have increased the total extent of such zones of submarine volcanism, which reduced the ocean-basin volume and partially displaced the water onto the continents [27]. On the other hand, the separation between the spreading zones implies their partial extension into the continents, which tended to split them, with the current tendency for the continents to be compressed being less pronounced or not occurring at all, and correspondingly the relief-producing effect was less. The continental masses were split up and separated repeatedly during the Mesozoic. Much of the continental mass in the eastern hemisphere existed in the Cretaceous (apart from the very end of the Cretaceous, when opposite tendencies appeared) as distinct subcontinents and microcontinents. This meant that the upper reaches in rivers were close to the base level, which reduced the heights at continent centers. Another consequence was that the total perimeter in the continents was increased, i.e., the area with crust transitional from oceanic to continental, particularly parts under the oceans.

Numerous processes are associated with secondary volcanic effects: increased atmospheric aerosols and CO_2. The first raises the reflectivity and reduces the solar radiation and thus the surface temperature, whereas the second tends to raise it via the greenhouse effect, with the effects from CO_2 several times those from aerosols [15]. One can relate the amounts of volcanics on the continents to the CO_2 levels in sediments at the same time (Fig. 21); there is a positive correlation at various periods in the Phanerozoic [15, 16]. During increased volcanism, the climatic contrast was thus reduced, but the mean surface temperature increased by several degrees, as can be seen from marine-sediment paleotemperatures, sediment facies, and plant and animal assemblages. The abundant CO_2 and the more favorable climate favored increased biomass, which reduced erosion and thus contrast in relief and favored transgression.

There is a further trend probably confined to thalassocratic epochs. Hot mantle materials entering the upper crust may lead to the mantle circulating,

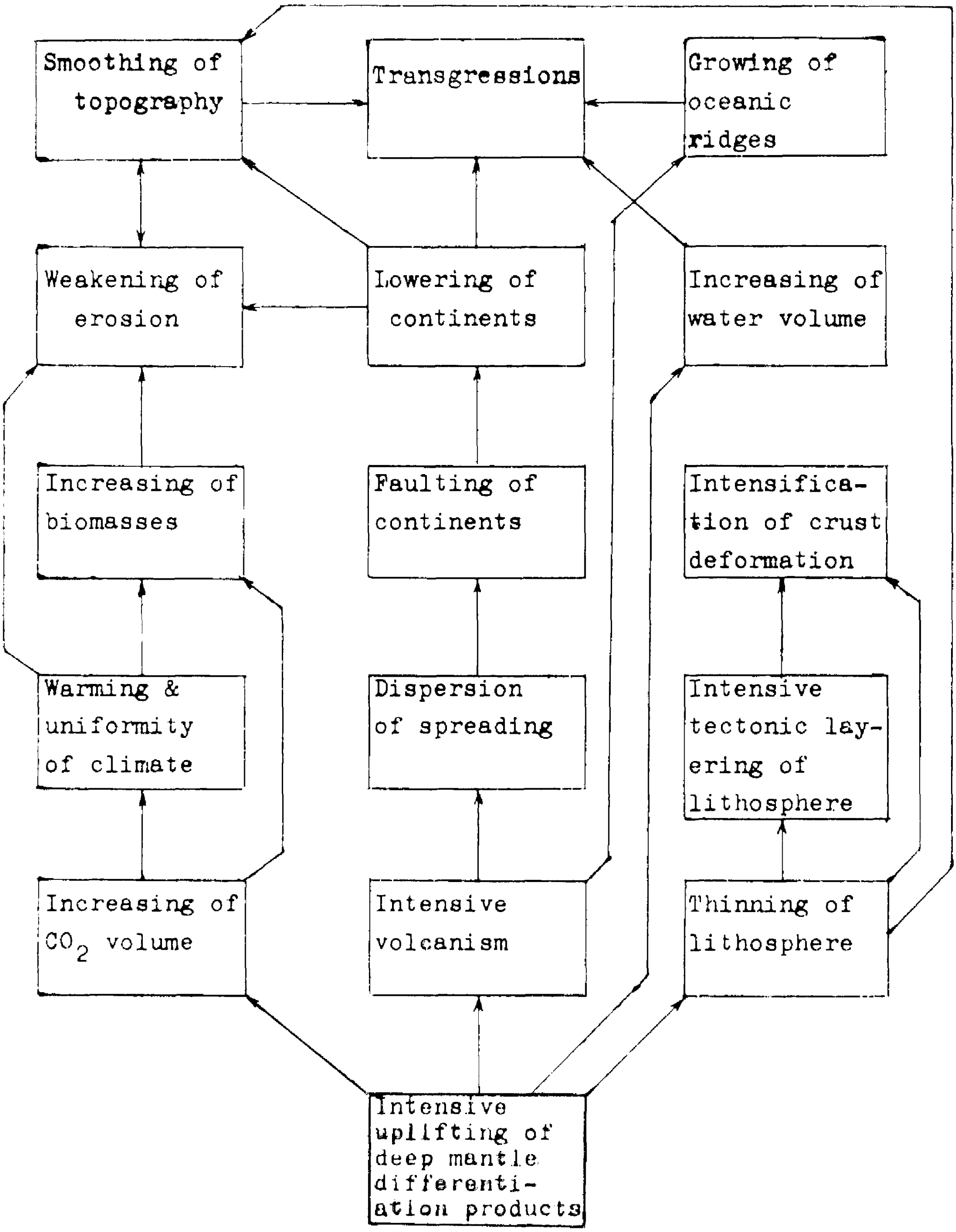


FIGURE 22. Effects of factors in the thalassocratic epoch by comparison with the Neogene-Quaternary geocratic epoch.

overturning, and descending; in other words, one gets a situation comparatively rare in the existing continents, where dedensified and hot mantle exist directly under the crust, which has the viscosity of the asthenosphere and may have been more characteristic of thalassocratic epochs. This would have heated the continental crust and raised its capacity for plastic deformation. Crustal layering levels may have approached the surface, while the velocity gradients in the differential lateral movements would have been increased. The thinner and less viscous lithosphere, however, would have been less capable of producing high mountains, which would have further smoothed the continents.

These various thalassocratic processes acted directly or indirectly to produce effects opposite to those from N-Q tectonogenesis: relief leveling and transgression, reduced mountain height, climatic equalization and general warming. Erosion and sediment redeposition were reduced, which minimized vertical movements associated with isostatic compensation.

* * *

Teilhard de Chardin [98] called the present an instantaneous section of unbounded time filaments. In a tectonic respect, that section has been examined much more fully than for preceding epochs, as it enables one to establish most fully the various and complicated relationships between phenomena. However, if one is to use them to interpret past tectonic processes, one has to correct for the changes in geological phenomena and in the lithosphere. It is necessary to distinguish ongoing processes, which produce similar effects at different periods, from tends and cyclic changes. The ongoing characteristics most fully defined for the N-Q stage include the mean rates of relative lateral displacements, structure combinations with definite origins, nonuniformity in structuring history, and pulses in tectonic movement.

The consistent trends are associated with crustal thickening and sialization, together with the entry of other mantle differentiation materials. Such changes can be detected when the N-Q stage is compared with remote epochs. The Early Precambrian lithosphere was thinner and more plastic, so it was less capable of producing large differential uplifts or brittle faulting.

The cyclic changes are seen during the Phanerozoic as alternation of geocratic epochs (such as the N-Q stage) and thalassocratic ones, with the latter differing from the N-Q stage in relief leveling and considerable continent flooding, as well as in less extensive and lower mountains, climatic leveling and warming, and possibly also partial degradation in the mantle part of the lithosphere, and higher deformability and tectonic layering in the crust. These features are associated with accentuated magmatism, which reflects the mantle materials entering the upper crust. One thus concludes that those materials rose to the surface in an uneven fashion during the Phanerozoic, which may have involved a self-excited oscillation.

8. Neotectonics and Adverse Consequences

8.1. Hazardous geological phenomena reflecting the current state and activity in the geological environment

Geological consequences are numerous and related within a complex system (Fig. 23); there are various phenomena hazardous to life, health, and economic activities, which are directly related to geology: earthquakes, volcanic eruptions, fluid-gas activity, landslips, rock falls, and mud slides, permafrost phenomena, karst and suffosion processes, groundwater breaking through into underground workings, and erosion and sediment accumulation. There is also a second group of phenomena hazardous or harmful to man but not related to geology although caused by geological factors. For example, uplifting or downwarping alters the surface runoff, as well as the hydrogeology and hydrochemistry, which affects the plant cover, the soil erosion, and the growth of moving sands or uplifts, as well as salinity or marshes in depressions. Major changes in landscape and weather are associated with volcanism. The more general neotectonic processes affect the climate. There is also feedback: changes in plant cover affect runoff and hydrogeology, whereas climate largely governs sedimentation, transgression, and regression.

Human activities cause changes that influence geological processes; the most obvious are the geological consequences of exploiting mineral deposits. The surface sinks where large amounts of groundwater are utilized or underground workings or boreholes exploit deposits. Downwarping larger in

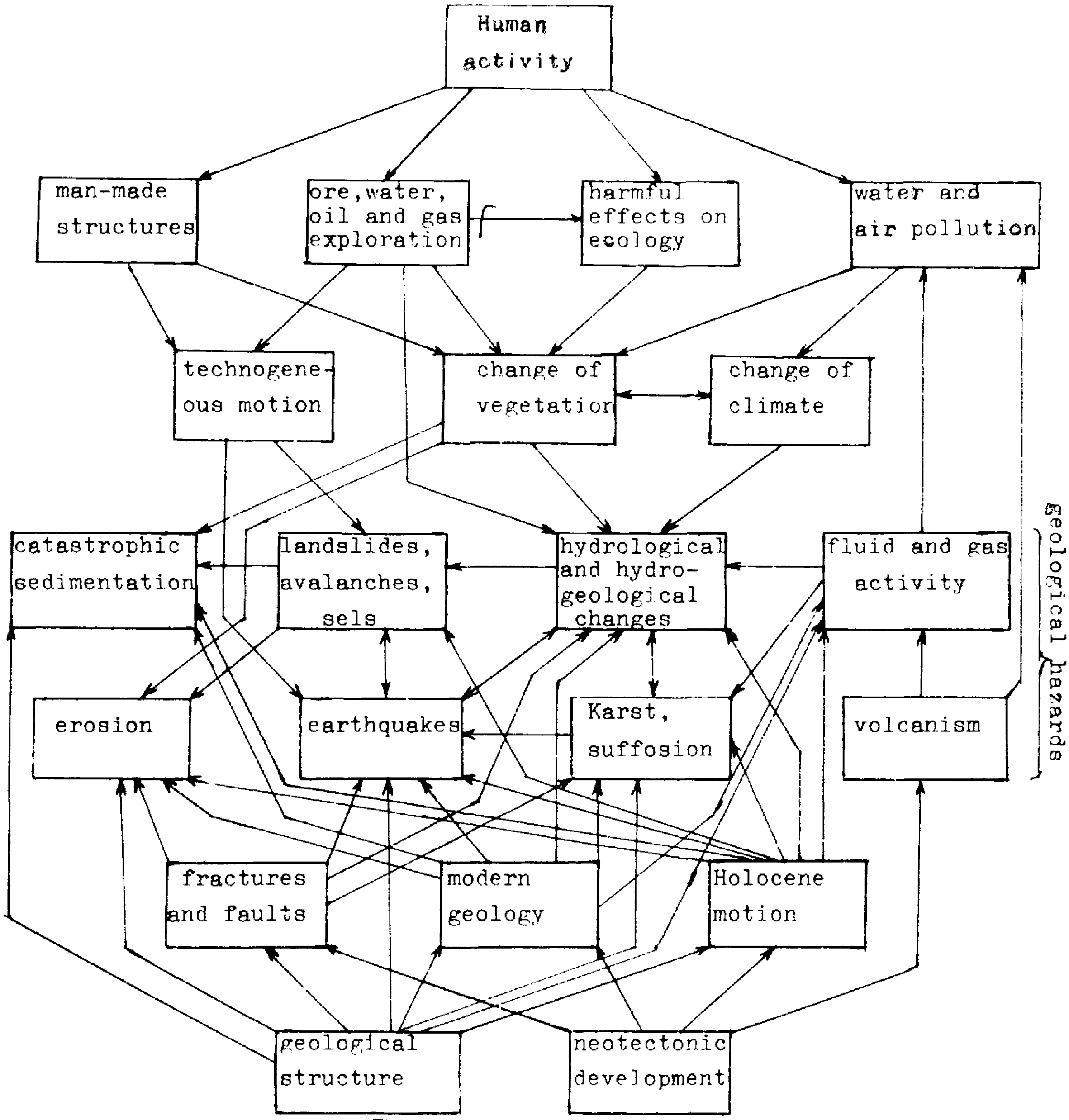


FIGURE 23. Interaction between factors governing geologic hazards.

scale although less in amplitude has been observed under all major cities. There are undoubted, although not unambiguous, effects from large reservoirs on seismicity and sedimentation. Cases have been described where hydrogeological conditions have been affected by building roads or other lines of communication, particularly groundwater pollution, flooding, and marsh extension. Under permafrost conditions, this may cause the ground to sink, ice barriers to break, and other adverse effects. As a rule, man-made ecological changes are also harmful in so far as they have geological consequences.

It is not possible here to consider in detail all hazardous geological phenomena, and I consider only one topic: geological evaluation of the hazards from strong crustal earthquakes.

8.2. Neotectonics and evaluating strong-earthquake hazards

Figure 24 [105] shows a plan for the geological research needed to evaluate hazards from strong crustal earthquakes; the first stage involves investigating the neotectonics and the N-Q trends of a region. The parts where the N-Q deformations concentrate may be considered as potentially seismic, but not all N-Q structures have remained active up to the present, particularly faults, so it is important to determine the development trends: accelerated or retarded motion at any point, structure rearrangement, and so on. For example, in northwestern Syria, neotectonic elements having northeast-southwest and north-south directions may be distinguished. The folds and faults in the first direction were particularly active up to the middle of the Miocene. Faults in the second direction form the complicated Levant zone and dissect it, and they show sinistral offsets [94]. A Pliocene-Quaternary age applies to that part of the Levant zone. The structures in the second direction represent a greater seismic hazard, and in particular should be the subject of detailed seismotectonic research.

In such potentially hazardous areas, one should map structures active in the Holocene, because most of the known strong earthquakes occur within them. One should determine the morphology, directions, and magnitudes and rates for the Holocene displacements. However, not all active zones and not all parts of them can generate strong earthquakes, which are dependent on the Holocene movements, where creep may accompany pulsed or pulsed-creep movement, in which case the probability of a strong earthquake is much reduced. There are two ways of evaluating the potential for strong earthquakes in an active zone.

The first is to identify traces of past strong earthquakes, where one estimates the magnitude and mean recurrence interval; Solonenko [91] and other Soviet researchers have shown that one can identify dislocations at the surface by deviations from gravitational equilibrium and by the shock waves in strong earthquakes, including ancient analogs. Methods have been devised for evaluating strong-earthquake parameters and recurrence intervals from such dislocations.

It is more difficult to identify the traces from strong seismic pulses in the morphology of the most active faults. Wallace [115] and Sieh [90] devised means of doing this for the San Andreas fault. I have used a similar method to demonstrate pulsed movement on faults in Mongolia and the Tien Shan [19, 94]. I was able to show that earthquakes with $M \geq 7.5$ occurred in the Khangay fault zone on the average every 600 yrs, whereas in the central part of the Kobda fault in Mongolia and the southeastern part of the Talas-Fergana fault in the Tien Shan, the recurrence interval is 700-800 yrs (see section 6.1). These earthquakes occurred on fault segments hundreds of kilometers long.

Another way involves estimating the potential for high stress concentrations occurring in large volumes, with subsequent rapid release; one needs to know the focal strength parameters. Competent rocks, the first being represented by such as the granite-metamorphic basement, or volcanics or limestones, are more capable of concentrating stresses than incompetent ones, such as serpentinites, marls, clays, and alluvium. The general crushing in the active zone is also important. These parameters can be evaluated from geological mapping. The current state of crushing can be evaluated from fault mapping. The more compact active zones represent more than ones where the faulting is distributed over large areas. Fault morphology is also important in the sense that strong earthquake hypocenters are usually close to large fault zones or directly in them in active shearing and downfaulting zones. In overthrust zones, they are more often associated with secondary faults (such as the White Wolf earthquake of 1972 and the San Fernando quake of 1971 in the Transverse Ranges in California), or else lie ahead of the main overthrust front (the 1897 Assam, 1918 Srimangal, or the 1930 Dzubri, and the 1934 Bihar earthquakes south of the main Himalaya boundary fault).

The behavior of faults at depth is important in evaluating potential hazards; if a fault penetrates

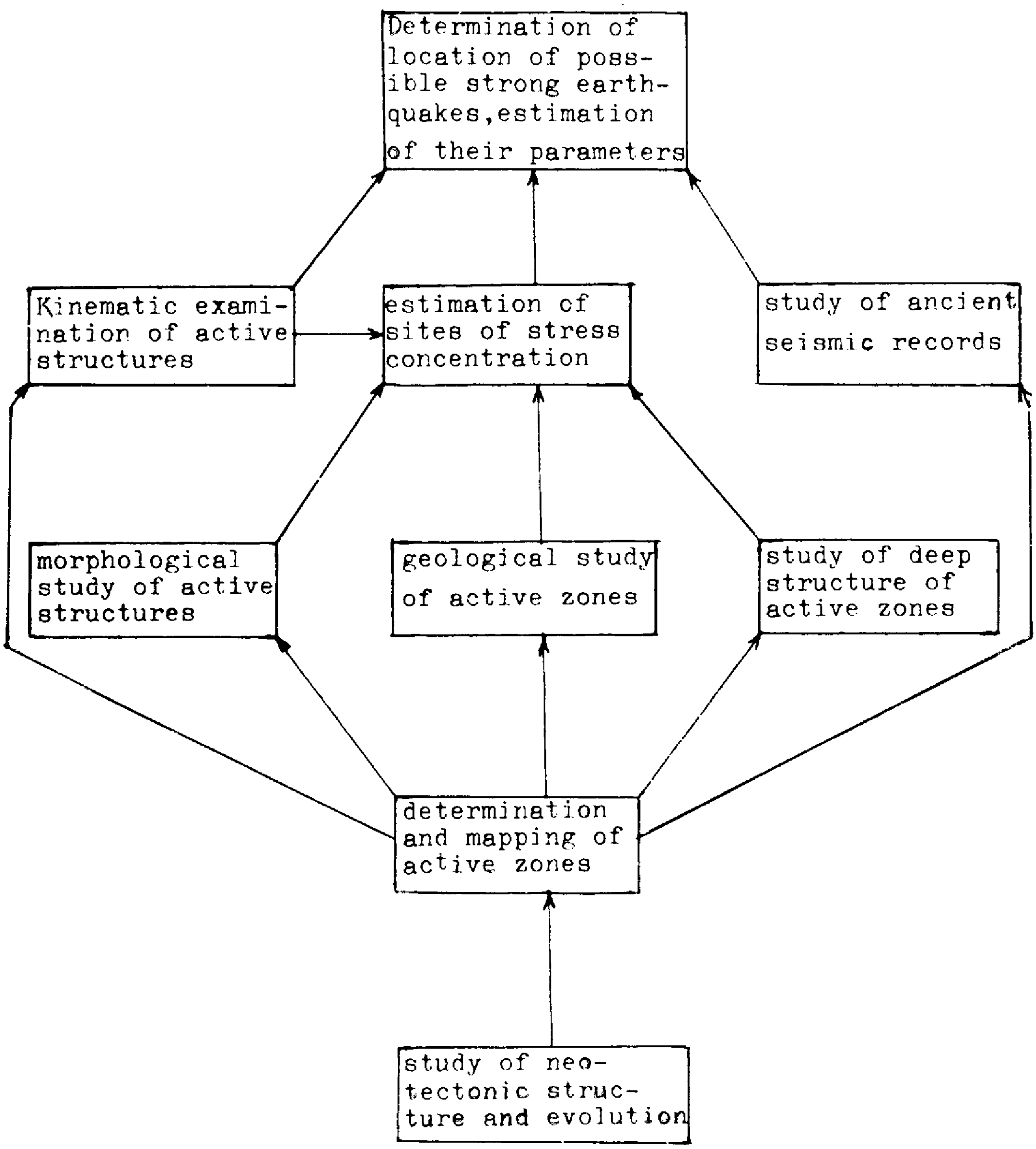


FIGURE 24. Geological investigation of strong earthquake hazards [105].

through much of the crust and emerges on the surface, the probability of stress concentration in that zone is less than in the one concealed at depth. The latter may be covered by undeformed rocks, tectonic cover, or a geophysical layer with different faulting. In the latter case, there is morphological or dynamic discrepancy between structures of the same age at different levels, which often leads to stress concentrations at the boundaries, with one layer slipping relative to another, i.e., tectonic layering. Such slipping on subhorizontal surfaces can be a source of strong earthquakes. Experience with the Gazli earthquakes of 1976 and 1984 in Central Asia [50] shows that the main hazard arises from areas of junction between faults differing in direction and active in different layers in the stratified crust.

The various factors causing concentration of stresses enable one to evaluate the probability of a strong earthquake in a given sector.

* * *

Adverse consequences arise from various processes linked into a complicated system; the decisive features are the structure, the N-Q and Recent tectonic activity, and human economic activity. A system approach can be applied to the hazard and provides a scheme for the necessary research. Where there is a likely hazard, as determined from the rates of the N-Q faulting and deformation, one should examine and map the fault zones active in the Holocene. Then within those zones, one establishes the most hazardous parts either from ancient dislocations and signs of ancient movements on faults or by evaluating the potential for stress concentration in large volumes. Such concentrations are deduced by examining fault morphology, zone geology, and deep structures.

Conclusions

Neotectonic enables one to examine details and complexities inaccessible for remoter epochs, partly because neotectonic forms are better preserved and occur everywhere but also because one can observe them directly and because one has a more appropriate range of methods for detailing geological data and supplementing them from geomorphology and geophysics.

Current neotectonics replaces a schematic model for the global lithosphere as based on the original form of plate tectonics by an improved model, which incorporates the differentiated conditions and movements at various levels (tectonic layering) together with rock deformability, as well as the various ways in which lithosphere thickening is balanced by subduction and consumption. Moreover, not only does neotectonics provide an improved global model but also indicates other structuring mechanisms (gravitational, isostatic, phase and other transformations), and it shows that they are all related to the global model and one with another within the tectonic system, which has multiple subsystems. The latter represent various levels of organization from the microscopic to global. The energy is provided not only by endogenous processes (mantle differentiation and radioactive decay) but also by cosmic factors, particularly solar radiation.

Neotectonics extends our view on structures and indicates a wide range of nonuniformities in development of structure, which range from individual seismic pulses alternating with quiet epochs to global phases of activation lasting millions of years. A bridge is thus built between the actual course of natural processes and the average as reflected in the geological record.

Geological effects harmful to man are at least related to neotectonics, and these can be understood and forecast only on the basis of the relationships.

Neotectonic experience should be more widely used in thinking about the past, but this needs to be done carefully, as one has to distinguish processes that vary little during the earth's history from those that show consistent trends and cyclic variations. A systems approach is required, as used in the relationship between neotectonic processes.

A more comprehensive approach is required to neotectonic formations, which should be based on all available methods, with cooperation between specialists; neotectonics provides accuracy unique in geology and gives a basis for evaluating geological processes in quantitative form, which should enable one to incorporate the methodology of the exact sciences more extensively into geology. Finally, neotectonics should extend our views on the conditions under which man emerged, his place and role in the past, present,a and future of the world.

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