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Abstract

The origin of the Himalayas, the highest mountain chain in the world, has been the most interesting problem from the geodynamic point of view. Plate tectonics, which considers the lithosphere as a number of rigid plates which are in a contact along their boundaries, can explain many of the features of the Indian subcontinent. Molnar and Tapponier (1975) and then many authors consider most of the large scale tectonics of Asia to be a result of the India-Eurasia collision. The collision of Indian plate and Eurasian plate involves the large scale of active continental deformation and therefore a rather diffuse seismicity prevails in this region. Moreover, the volcanism in its separate parts is indicated. To analyze these regions from the geodynamic point of view we shall construct the geodynamic models of these regions in the form of vertical profiles. For this reason the map of the lower boundary of the Indian lithospheric plate as well as new geodynamic models of the Himalayas and the subduction of Indian plate below the Burmese plate will be given and discussed. In the paper the stress-strain analyzes across the Nanda Devi profile as well as the profile across the Andaman Islands will be numerically computed and then the obtained results will be analysed. While geologist presume the contact boundary between the Indian plate and the Eurasian plate to be the Indus Tsangpo Thrust, the obtained results of the numerical computations indicate that the contact between the Indian lithospheric plate and the Eurasian plate is the Main Boundary Thrust and the Main Central Thrust is the deep fault from the Himalayan evolution period before 40 Ma.

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Keywords
Geodynamics, mathematical simulation of geodynamic processes, Himalayas, Andaman Islands

1 Tectonic Evolution of Himalayas

The origin of the Himalaya, the highest mountain chain in the world has been the most interesting problem from the geodynamic point of view. Plate tectonics can explain many of the features of the Indian subcontinent. The collision of Indian plate and Eurasia involves the largest scale of active continental deformation and therefore a rather diffuse seismicity prevails in this region. At present authors consider the collision of the Indian and Eurasian plates, with Indus-Tsangpo Thrust marking their contact, to give rise to the Himalaya. The evolution of the Himalayas is characterized by several stages of evolution, starting from the Aptian up to present. During the first two stages of evolution of the Himalayan region, the subduction of oceanic lithosphere under the continental one of Eurasia was characterized by different speed of movement of the Indian plate relative to the Eurasia one.

The Aptian (Middle Cretaceous $\sim$ 110 Ma) is the time when India, together with Madagaskar, splits apart from East Africa, and as a result the Somalian basin within the Indian Ocean arise. At the period of Cretaceous-Paleocene ($\sim$ 65 Ma), India moves to the north at very high velocity of about 18.4 cmyr^{-1} with the India-Eurasia pole of relative motion in North Africa. In the Late Eocene and Oligocene between 65 and 54 Ma, it moves in the north direction at the velocity of 12 cmyr^{-1} and between 54 to 35 Ma by the same northward motion at the speed of 8 cmyr^{-1}. During Late Eocene-Oligocene, oceanic areas in the Neo-Tethys are shortened by 300 km to the west, 600 to 1000 km off Iran and by 3000 km between India and Euroasia. Then, between 35 and 20 Ma (Early Miocene $\sim$ 20 Ma), the speed of motion of the Indian plate begun to slow down to 5 cmyr^{-1} due to the collision of India with Eurasia. The India-Eurasia pole of relative motion is affected in the same way as in the preceding period. However, it abruptly moves east of the horn of Africa, which results in dextral northwestward motion of India, due to the collision of the two continental plates, Indian and Eurasian plates. During this period, the whole Alpine-Himalayan system along approximately the same latitudes was created. This is the time when continental collision operates along all the Eurasian margins. As a result, a complex system of intra-continental strike-slip faults begins to appear. During Early Miocene ($\sim$ 20 Ma) India moves northwestward. The flysh sedimentation were observed, the basin in thrust over the Indian plate where it forms the present Waziristan, Zhob and Quetta arcs. During the Middle Miocene ($\sim$ 10 Ma), the India-Eurasia pole of rotation moved back from the eastern Indian Ocean into North Africa, into the region of northern Egypt. The speed of motion of the India plate to the north is about 4 cmyr^{-1} and it becomed north-northeastward at about 5 cmyr^{-1} after the Upper Miocene. The continent-continent collision operates along the Neo-Tethys margin with the rate of convergence increasing from west to east. Folded mountain ranges develop: the Pamir-Hindu Kush and Himalayas, in the region of collision of the Indian plate with Eurasian plate.

The tectonic evolution of the region has continued since the last 10Ma up to present. To understand the stress-strain state in the northern part of the Indian plate-the Indian subcontinent, we can divide it into four different regions:
(i) the Himalaya, (ii) the Indus-Gangetic basin, (iii) the Peninsular Indian Shield, and (iv) the Andaman-Nicobar arc region.

The Himalayas are geografically defined as the 2500 km long mountain range 200-250 km wide, from Nanga Parbat (8125 m) in the west-northwest to Namche Barwa (7755 m) in the east and being convex towards the south. South of the Himalayas lies the Indus-Ganga-Brahmaputra basin, and south of the Ganga basin lies the Peninsular Indian Shield. Western part of the Himalayas is bounded by Sulaiman and Kirthar ranges; by Hindu Kush ranges on the north-west and by Tibetan Plateau on the north. To the east, Himalayas are bordered by the Arakan-Burma ranges. Andaman-Nicobar ridge system forms the southward continuation of the Arakan-Burma arc and it represents the boundary between the Indian and Eurasian plate in the east. The present situation in the collision zone between the Indian lithospheric plate and the Eurasian plate is figured at Fig. 1.

The Himalayas consist of sedimentary rocks, deposited over millions of years in shallow seas. The border between the Indian lithospheric plate and the Eurasian plate, according to Ghosh (1988), Kaila and Narain (1976), Verma (1991) and others, is situated in the Indus-Tsangpo Thrust (ITT). They belive that the ITT (they denote it as Indus Suture Zone (ISZ) or the Indus-Tsampo Suture (ITS)), with the presence of ophiolitic suite, is possibly the relict of old Tethys and the junction of two continental lithospheric plates. It might represent earlier subduction zone and a collision boundary. The stage of the subducted Indian plate below the Eurasian plate during the period between the Late Cretaceous to Palaeocene is presented in Verma (1991, p.282). The process of subduction led to the formation of granitic plutons along the southern margin of the Eurasian plate (Tibetan Plateau). The granitic plutons form a magmatic arc along the north margin of the Indus-Tsangpo Thrust (ITT). In addition, a volcanic arc developed north of the trench axis, and it extends from the Kohistan region, situated west of Nanga Parbat, to Shigatse, situated northwest of Lhasa, in the east part of this volcanic arc. In the author's geodynamic model the border between the Indian lithospheric plate and the Eurasian plate is situated in the Main Boundary Thrust (MBT), which is explained by the numerical results and by their good correlations with geological observations.

Valdiya (1984a), (1984b) supposes that actual collision between India and Eurasia took place during the Eocene to Oligocene, and it leads to the obduction of deep oceanic crustal material in the form of ophiolites (in this area represented by a suite of rocks including harzburgite, pillow lavas, dunite, sheets of dolerites, layered gabbro, etc.). South of the Indus-Tsangpo Thrust, plutonic bodies of quartz diorite-granodiorite-granite association were also formed. The rise of the crystalline basement occurred south of the Indus-Tsangpo Thrust.

2 Geological and Tectonic Characterization of the Himalayan and Tibetan Plateau

Based on geological and tectonic studies the Himalayas can be subdivided into the following main areas:

The Siwalik Belt

Due to Valdiya (1976) it is characterized by rugged topography, being built by Tertiary and Early Pleistocene molassic sediments. The belt is bounded by the Main Boundary Thrust (MBT) in the north and separated by the Himalayan Frontal Fault (HFF) from the planes in the south. The MBT consists of a series of thrusts that separate the pre-dominantly pre-tertiary Lesser Himalayan sediments from the Tertiary Siwalik sedimentary belt (the Sub-Himalayan belt). The MBT has developed during Pliocene and the thrust zone was active through the Pleistocene (Valdiya, 1981). However, results of modelling, presented by the author firstly in 1996 at the Indian School of Mines, Dhanbad and at the Wadia Institute of Himalayan Geology, Dehra Dun, indicate that the MBT represents the main contact boundary between the Indian lithospheric plate and the Eurasian plate.

According to Gansser (1964) the Siwalik belt is built up of structures with a normal and not a reversed sedimentary succession. More or less wider synclines alternate with steep and often faulted asymmetric anticlines. The strike-slip faults in this area are steep at the surface, but probably dip towards the mountains at depth (i.e. northwards). In the upper Siwalik belt, the younger layers outcrop in the north along the Main Boundary Thrust (MBT), whereas monoclinally folded older Siwalik members outcrop in the south. Their southern limit is mostly erosional and their course is interrupted by the large fans of the Ganges alluvial deposits. Folding and faulting in this area are of the early to middle Pleistocene age, and probably have preceded the overthrusting of the Lesser Himalayas along the MBT. The strike of the Siwalik does not always conform to that of the fault. North of the Main Boundary Thrust, the Siwalik does not overlay the Lesser Himalayas. They are overthrust by the pre-Siwalik rocks of the Lesser Himalayas and therefore, they do not form a normal cover for these ranges. Moreover, the northern hinterland of the Himalayas is a highly elevated mass, which is drained by southwards-flowing rivers.

The Main Boundary Thrust forms the southern limit of the Lesser Himalayas. Secondary thrust or faults branch off the Main Boundary Thrust (e.g. in Punjub, where they separate the Siwalik and Murrees). These secondary thrust zones diverge westwards and then joint with the main fault towards the east. One of them is the Krol thrust sheet. The Krol thrust sheet coincides, for over 400km, with the Main Boundary Thrust. On the other hand in some places, as in south Nepal, the higher, internal crystalline thrust sheets lie in contact with Siwalik. Gondwana formations are situated between the MBT and the next higher, internal thrusts, as it was observed in the Damudas. Along the frontal outcrops, the MBT and the internal thrusts show a similar dip and run parallel to each other.

The Lesser and High Himalayas

The topography of the Himalayas (Gansser, 1964, Ni and Barazangi, 1984, Verma, 1991) is characterized by
(i) an abrupt rise of the Sub-Himalayas and the Lesser Himalayas above the close-to-sea-level Ganga basin,

(ii) a gradual increase in elevation across the Lesser Himalayas,
(iii) a steep slope between the northern Lesser Himalayas and the High Himalayan range and
(iv) a uniform elevation of about 5 km north of the High Himalayan range. The uplift of the Himalayas is a consequence of the horizontal convergence between India and Eurasia. The Lesser Himalayas and Sub-Himalayas represent ranges with elevation from about 1500 to 3000 m with dissected valleys. The Lesser Himalayas consist of various types of sediments and metamorphics, which have been strongly folded and thrusted along several North-South thrusts.

The Lesser Himalayas are separated from the High Himalayas by the Main Central Thrust (MCT), which has a northward dip of about $30 - 45^o$. The MCT is known under different names in different parts of the Himalayas, i.e. as the Panjal thrust in the Kashmir Himalayas, the Vaikrita thrust in the Kumaun Himalayas and the Darjeeling thrust in the Eastern Himalayas. The thrust is characterized by a zone of an abrupt change in the structure, intense shearing, and in the grade of metamorphism represents the main manifestation of the presence of the MCT. The MCT has existed since mid-Tertiary time, and there are some geological indications of minor recent movements along the MCT (Gansser, 1964, Valdiya, 1980, Verma, 1991). According to the author's results, it originated during the subduction of the oceanic (Tethyan) lithosphere under the Eurasian plate, and therefore it is younger than MBT and older than the Indus-Tsangpo Thrust.

The Lesser Himalayas on their southern side are limited by the Main Boundary Thrust. As it was observed in situ, a border facies of the Peninsular Shield makes up relatively a large volume of the formations of the Lesser Himalayas, which can have their origin in the young Himalayan phase during the mid-Tertiary age. Gansser (1964) noted that the primary differences in metamorphism, reflecting the history of the shield rocks, were overprinted by the metamorphism of these young Himalayan phase. Moreover, the strinking changes in metamorphism are mostly restricted to the older rocks. In the Himalayas no higher grade metamorphism has been observed in formations younger than Ordovician. Only along the ophiolite belts such as the Tibetan zone and in the Karakorum, younger beds of the Jurassic age (e.g. in the southern Tibet) and Permo-Carboniferous age (e.g. in the Karakorum) were affected by higher-grade metamorphism. The Himalayan metamorphism which produced migmatization and finally young mobilization were restricted to the previously metamorphosed sections of the crystalline thrust sheets. The youngest granites which intrude even Mesozoic sediments have not produced any marked contact metamorphism.

Along the MBT, a strong tectonization was observed, but this structural zone is not a sharply delimited feature. The strong tectonization is manifested in the foothills of Bhutan. Within this tectonized zone along the Himalayan south border, independent thrust sheets, as the Krol thrust, are developed. Where the Krol thrusts are absent, a segment of steeply thrust Gondwana rocks occur. They extend practically along the whole length of the Nepalase foothills, and continue through Sikkim and Bhutan upto the frontal thrust region in the NEFA Himalayas.

Between the Main Central Thrust and the Indus-Tsangpo Thrust is situated the 2500 km long mountain range of **the High Himalayas,** which is about 200-250 km wide. It starts in the area of the Nanga Parbat on the west-northwest and ends in the area of Namche Barwa on the east.

The High Himalayas are built by monoclinal, in the centre, and/or gently folded, in the western and eastern part, Precambrian metamorphics intruded by middle Tertiary granites. Valdiya (1979), (1981) shows that the High Himalayan rocks represent basement rocks. A previous study of elevation in the High Himalayan-Tibetan region (Bird, 1978) indicates that the mean elevation of the High Himalayas is about 5000 m, similarly as in Tibet, but with maximal elevation more than 8000 m. In the northern half of the Himalayas there are fewer earthquakes. In contrast, there is a maximal concentration of seismicity in the southern half.

North of the High Himalayas lies a vast stretch of the Tethys Himalayas, made up of fossiliferous sediments deposited in the basin of the Tethys Sea during the Late Precambrian to the Upper Cretacerous period. The Tethys sediments occur in several basins that extend along almost the entire length of the northern margin of the Himalayas, and which are suddenly cut off by the Indus-Tsangpo Thrust (ITT). The Tethys sediments are preserved in several basins all along the length of the Himalayas i.e. basins: the Kashmir and Spiti basins in the NW Himalayas, Kuman basin in the central part, West Nepal basin in the Nepal Himalayas and Kampa Dzong basin in the eastern part of Himalayas.

According to our numerical results we indicate that the Indus-Tsangpo Zone is the thrust. The Indus-Tsangpo Thrust according to (Gansser,1964, 1980), denotes the northern limit of the Indian subcontinent following the Late Cretaceous-Early Tertiary closure of the Neo-Tethys. The ITT comprises pillow lavas, volcanics, and basic and ultrabasic rocks as well as radiolarite. The author, due to the own numerical results, showed that the Indus-Tsangpo Thrust is the secondary deep fault probably of the third period of evolution that originates in areas with traction stresses.

North of the Indus-Tsangpo Thrust, a long belt of Trans-Himalayan plutons is situated. The belt extends from Pakistan i.e. from Swat 36°N,65°E, situated west of Nanga Parbat-Hamosh massif, to the eastern margin of the ITT $\sim$ 30°N,90°E, where it forms an Andean-type magmatic arc. This magmatic arc system follows directly from the conditions on the contact boundary discussed in detail in the APPENDIX and it is indicated from the obtained numerical results. The plutons located along the belt are known as Swat, Landkh, Kailas, Tsangpo, Kyichu (Lhasa), and Eastern Tsangpo. The Swat and Ladakh plutons are separated by a transverse structure, the Nanga Parbat high, which makes part of the western Himalayas. The Trans-Himalayan Plutons have a genetic link to the deep fault zone, which suggests that magma production is related to the processes that occurred during the subduction of the Indian plate.

The High Himalayan pluton belt extends over a large part of the Himalayas and it runs close to Kedarnable, Badrinath, Mustang, Mugu, Manaslu, Makalu, Kangchenjunga, Thirupu and Chekha in Bhutan on the east. The plutons are of two types. The older: 580 ± 10 Ma - the Manali-Rohtang area, 495 ± 16 Ma - Lahsel-Spiti area, 467 ± 46 Ma biotite granite from Manikaran, 486 ± 10 Ma - Nepal Himalayas and the younger (Upper Tertiary): Manaslu granite - 28 Ma-18 Ma, Badrinathh granite - 18 Ma, 19 Ma - biotite from Everest granite. These younger plutons were formed by remelting of the preexisting older rocks. The mechanism is mathematically described and discussed in Nedoma (1997a). Mehta (1980) shows that the younger ages fall into three groups related to different phases of the Himalayan orogeny. The first group 75-50 Ma is related to folding and metamorphism; the second one 40-25 Ma, appears to be related to the uplift of the Himalaya; the third one 25-10 Ma, is related to the major uplift, formation of nappe structures and regional metamorphism.

The Tibetan Plateau

The Tibetan Plateau is situated north from the High Himalayas and represents the region formed predominantly during the first two periods of evolution of the Himalayas. The southern and main Kun-Lun borders the High Tibetan Plateau to the north. The geology of the region of the Central Tibet is not so complicated. It is characterized by elongated ranges striking mainly E-W. Marine Jurassic is succeeded by Cretaceous and widespread, Tertiary formation of mostly continental type. The distribution of young volcanic rocks in the Tibetan Plateau and along the Kun-Lun Range is quaint. As often suggest, the youngest basaltic lavas together with relic volcanic cones are probably of Pleistocene age, which closely corresponds to the mechanism of the volcanic chambers located at the boundary between the lithosphere and asthenosphere near the subducted (Tethian oceanic and colliding Indian plates) lithosphere. In the southern Kun-Lun, there are indications of the post-Pleistocene volcanism (Gansser, 1964, Leuchs, 1937).

The highland to the west is situated between the Karakorum in the SW and the Kun-Lun in the north. Tethys-type marine formations are widespread in the Inner Tibet. The Tibetan high plateau is limited to the east by the Eastern Himalayan syntaxis. Its north-east border is formed by the Nan-Shan ranges. The Nan-Shan ranges end in the Kuhn-Nor depression. The E-W striking element is the Tsingling-Shan Range. A major structural element is formed by the Tsingling-Shang Range of metamorphic and basic rocks.

The Northeast and East Tibet are built by older formations. In the Tang La Range and NW Yunnan, the Devonian formations and Hercynian core of the range are observed. From the Lhasa-Bomi folded belt, thick Devonian and Carboniferous deposits with intercalated submarine volcanics representing the deeper elements of the Tibetan platform are described. In the eastern Tibet (Chando region), deposits of the Lower Palaeozoic are situated. Structures of the Tibetan/Tethys Himalayas are of several types. The crystalline thrust sheet of the Higher Himalaya is normally covered by a great thickness of mostly pelitic sediments. South-vergent folds and south-directed thrusts are mainly observed, but in the northern Nepal, the north-vergent folding play an important role. Towards the western Himalayas, where the range widens and the elevations of the main range decrease, the folding of the Tibetan Himalayas is gentler and more of a Jura-type. The Mesozoic section of the Tibetan Himalayas is characterized by differentiation into plastic shales, thick limestones and dolomites. The disharmonic folding is observed in these areas.

The author assumed that the Tibetan Plateau was mainly formed during the first two evolution periods namely when the speed of Indian plate with respect to the Eurasian plate was less than 8 $cmyr^{-1}$, i.e. the last 54 Ma.

3 Seismic Activity

The Himalayan region is one of the most seismically active region in the world (Fig.2). The seismotectonic activity in the Himalayan region is primarily confined to the region between the MBT and MCT. A notable feature of the seismicity of the Himalayan arc is that earthquakes of large magnitude (~ 8.0) have occurred close to the MBT of the Himalayas. These include the Kangra earthquake of 1905 (of magnitude ~ 8.4), Bihar earthquake of 1834 (of magnitude ~ 8.3), Assam earthquake of 1950 (magnitude 8.6), probably the Shillong Plateau earthquake of 1897 (magnitude ~ 8.7), as well as earthquakes approximately of the magnitude ~ 7.0, like Kashmir earthquake of 1985, Kuman Himalaya earthquake of 1803, Nepal Himalayan earthquake of 1833 as well as two earthquakes, that occurred between the 1897 Shillong earthquake and the 1950 Assam earthquake, of magnitude 7.2 (1943) and 7.7, respectively, are also located in this region. The seismic activity is quite intense and it is situated in shallow depths in the eastern and western Himalayas. The seismicity in the central Himalayas is proved to be rather low and irregular in time and space. The linear alignments of the earthquakes, situated transverse to the Himalayan arc system, are also evident. The earthquakes in the neighbourhood of these transverse structures are of a strike-slip character, with one of the planes transverse to the Himalayan arc system, as it is documented by the earthquakes of the Bihar-Nepal (1934, of the magnitude ~ 8.4), Bihar-Nepal (1988, of the magnitude ~ 6.6), Uttarkashi (1991, of the magnitude ~ 6.6). North of the Indus Tsangpo Thrust, i.e. in the region of the Tibetan Plateau, the seismic activity, with shallow earthquake depths, is irregular in time and space. In this Tibetan region, however, intermediate-depth earthquakes were also observed. However, the source mechanisms of most of the earthquakes in the Himalayan arc system are predominantly of the thrust type. During the last century the location of four great earthquakes were situated to the western and eastern Himalayas. The central Himalayas have been relatively quiet.

4 Geodynamic Model of the Himalayas, Profile-Nanda Devi

The Himalayas, the highest mountains in the world, form a well defined arc to the north of the Indian Peninsula and extend over a distance of about 2500 km from the west-northwest (Nanga Parbat 8125 m/Indus River) to the east-northeast (Namcha Barwa 7755 m/Brahmaputra River). They comprise more than 30 peaks, rising to the height of 7300 m above the mean sea level. The formation of the Himalayas is the result of the collision between the Indian and Euroasian plates. The Himalayas can be divided physiographically into four parallel, longitudinal belts: the Outer or Sub-Himalayas, the Lesser or Lower Himalayas, the Higher or Great Himalayas and the Tibetan or Tethys Himalayas. Further north, as a part of the Himalayan orogenic belt, the Trans Himalayas are situated.

The investigated area comprises five main lithotectonic zones of the Himalayas with its foreground: the Indus-Ganga-Brahmaputra plain, the Siwalik belt of the Outer or Sub-Himalayas, the sedimentary belt of the Lesser Himalayas with subordinate volcanics, Central Crystalline zone of the Higher Himalayas and the Tibetan Himalayas.

The Ganga basin appears to have had a very long and complex history of evolution dating back to Precambrian times. At present, the Ganga basin essentially consists of the Upper Tertiary sediments, mainly of Siwalik formation, overlying the Vindhyan formation and Aravalli-Delhi metasediments. Analyses of geophysical data show that the Vindhyan basin extends northward, far beyond its surface exposures in the northern shield, and that an considerable thickness of these sediments underlies the Ganga basin, the deepest part being close to the Himalayan foothills. The Ganga basin is characterized by a small thickness of the lithosphere of about 60km and therefore by an anomalous geothermal field. The thickness of the lithosphere was determined from the author's map of the lower boundary of the lithosphere (Fig.3), based on the analysis of geothermal field made by Negi et al. (1986) and on the author's geodynamical analysis of the Himalayan arc system. This analysis shows that the Indian plate, which is in a collision with the Eurasian plate, starts to thicken in the area of the Central Peninsular India, wheras in the area of the Ganga plain, the bottom of the lithosphere bulged out and the upper parts of the lithosphere subsided, which resulted in weakening of the lithosphere (Fig.4). Moreover, from the thermodynamics it is known that the temperature of melting depends also on the pressure. If the pressure increases then the temperature of melting also increases, if the pressure decreases then the temperature of melting also decreases. Since the bottom of the lithosphere bulged out, the pressure in this area dropped and therefore the temperature of melting rapidly decreased. Then the rocks of the lower lithosphere and asthenosphere are melted or partially melted, which together with the weakened lithosphere produces anomalous geothermal field.

Analyses of geodynamic processes, geological and geophysical data of the area of the Himalayas as well as results of previous similar analyses from Central Aleutian arc system indicate that the major contact between the Indian lithospheric plate and Eurasian lithospheric plate is not the Indus-Tsangpo Thrust (ITT) but the Main Boundary Thrust (MBT). The evolution of the Himalayas has three major periods.

The first and second stages are characterized by the position of the India with respect to the rest of Eurasia in the period of 110 Ma to 40 Ma, i.e. by the period where the Indian continental lithosphere was situated between the Africa and Eurasian plates. These periods are represented by the subduction of the Tethyan lithosphere of oceanic type below the Eurasian plate and by the development of the island arc system (- the Tibet). This configuration is similar to that of the present subduction of the Pacific plate below the American plate. By analogy with the Grow's model of Central Aleutian arc system we can deduce, that the maximal depth of the subducted Tethyan lithosphere is about 300 km. Similar estimate follows from our observation of the thickness of the lithosphere in the area of the collision of the Adriatic microplate with the Eurasian plate. From the analysis of the thickness of the European lithosphere it follows that the ratio between its thickness below Alps and Dinarides is $\sim 3/2$. Since there is a certain similarity between mechanisms of present evolutionary stages in Europe and in India, we can assume that the ratio between the thickness below the Himalayas and below the Burma is also $\sim 3/2$. The present thickness of the lithosphere below the Burmese plate is of about 160-220 km. We can assume that the mechanism of subduction below the Eurasian plate in the direction S-N and below the Burmese plate during the first period of evolution was a bit at least similar as at present in the Aleutian arc system. Then for the maximal depth of the subduction of the Tethyan plate of oceanic type below the Eurasian plate we can find the value of about 240 km to 330 km. Since this archaic Tethyan lithosphere is partially melted, then at present it cannot be manifested by the earthquakes as the deformation energy accumulated in this Tethyan lithosphere is released in other ways, like the creep, chemical processes, heat, solidification and recrystallization of rocks, the second phase change, etc. During this period the deep faults probably originated in the similar positions like in the case of the subduction of the Pacific plate below the American plate as well as in the arc system in the Tethyan Sea. The fault near the archaic trench is the Main Central Thrust. In this geological period, the MCT was directed oblique across the Eurasian plate. The second deep fault situated north of the MCT could or did not have to reach the Earth surface. During this first geological periods the Tibet originated as an island arc system and in the next phase the Tibetan Himalayas and the Trans Himalayas were folded; according to our estimates, the uplifts of the Tibetan and Trans Himalayas during the first period of the evolution, when the speed of the subducted Tethyan lithospheric plate below the Eurasian plate was more than 8 cm/yr, was very small (only a few mm/yr), while during the second period of evolution, when the speed of the subducted lithosphere was less than 8 cm/yr, the uplift was more than 10 mm/yr. During the last 10 Ma the Himalayan arc system above all was developed.

The third geological period of the evolution starts at the moment of the first contact of the Indian continental lithosphere with the Eurasian plate 40 Myrs ago. This contact resulted in the folding in Hindu Kush and Pamir. During the last 10 Ma the Indian plate has been rotating with respect to Eurasian plate, which leads to the concave shape form of the Himalayan arc system. At every time step of evolution, two different deep faults originated. One of them is the Indus-Tsangpo Thrust, the second one is situated north of the ITT in the Tibetan Plateau. The rock formation of the Lesser and High Himalayas are folded. It is evident that the MCT during this second and namely third period of the evolution process changes its position, from the initial oblique direction across the Eurasian plate with an angle of about $115°$ to the present oblique direction with an angle of about $30° - 45°$. The rocks between the MBT and MCT as well as between the MCT and ITT are deformed, which reflects the uplifts of the Lesser and High Himalayas. However, since the rocks between the MBT and MCT are strained by the colliding Indian plate then, due to a small angle between the MBT and MCT direction and also due to the acting Coulombian friction, the uplift of the Lesser Himalayas is less than in the case of the High Himalayas. In the latter region, the angle between the MCT and ITT direction is greater than $90°$. These uplifts are of a different origin than earlier thought based on the isostasy theory. Our numerical investigations indicate two types of volcanism-the calc-alkaline type in the ITT area and the basaltic type to the north.

The Sub Himalayas consist of the Siwalik Hills, with height varying from 250 m to 800 m and with width of about 25 km to 100 km, situated to the south of Main Boundary Thrust (MBT). The Siwaliks comprise molassic sediments of Neogene and younger rocks, like red shale, red sandstones, grey clay in Lower Siwaliks, micaceous, friable grey sandstone in Middle Siwaliks and boulder conglomerate, pebble horizons and sands in Upper Siwaliks its thickness vary from 4500 m to 6000 m. The Siwalik Hills might represent a relic of an arc system from the first two periods of evolution discussed above. Since the Siwalik sediments are of the Neogene and younger age and since the island arc system can be of the same age, it is probable that Siwaliks are the relic of this arc system. Similarly, the Main Frontal Thrust (MFT), situated to the south of the MBT, is the relic of deep fault from the first and namely second period of evolution of Himalayas. Folding and faulting in the Siwaliks are of the early Middle Pleistocene age, and must have preceded the overthrusting of the Lesser Himalayas along the Main Boundary Thrust (Gansser, 1964), which is a result of the following geological period.

The Lesser Himalayas are bounded by the Main Boundary Thrust (MBT) in the south and Main Central Thrust (MCT) in the north and it is built by south-directed nappes involving rocks of Precambrian to Mesozoic age, mainly devoid of fossils. The Lesser Himalayas consist of thick sedimentary sequences, low grade metamorphics and volcanics which are mainly of the Early Proterozoic to Lower Paleozoic ages. All along the Lesser Himalayas, the thick argillaceous formations with some quartzites and limestones are observed.

In the Lesser Himalayas of Garhwal and Kumaun sections, the four principal lithotectonic units are distinguished: Autochthonous-Paraautochthonous sedimentary belt, Berinag Unit, Ramgarth Unit and Almora Nappe Unit.

The autochthonous unit of Precambrian sedimentary rocks are observed in the northern (inner) belt of the Lesser Himalayas. The paraauchthonous nappe of the southern (outer Krol) belt of the Lesser Himalayas comprises the Jaunsar and Mussoorie groups of sediments of the Late Precambrian to Lower Cambrian ages. The Berinag Formation is formed of huge succession of massive coarse-grained to pebbly and often sericitic quartz arenite with metamorphosed basalts and tuffites. The Berinag Quartzite and Chamoli Quarzite and volcanics of Karanprayag-Chamoli area are observed in the Alaknanda valley and Uttarkashi area. The Ramgarh Nappe is built by the crystalline rocks, and it consists of two units: the Ramgarh porphyroids with sericite quartzite and phyllite, and the Amritpur granite. These units are separated by an intervening unit of the Bhowali-Bhimtal volcanics and the Nagthat quartzite. The porphyric and mylonitised granitic gneiss of the Ramgarh Nappe is about 1765 ± 60 Ma old (Trivedi et al., 1984). The Almora Nappe is built by a variety of schists, micaceous quartzite and amphibolite-facies gneisses together with intrusions of granite and granitodiorites (Valdiya, 1980). The Almora granite is of 560 ± 20 Ma in age.

The High Himalayas consisting of 10-15 km thick sequence of metamorphic rocks and granite represent the region of the greatest uplift; our numerical results indicate the uplift of about 19.58-20.73 mm/yr at maximum (obtained from numerical data computed for times $\sim$ 1.50E+5 and/or 1.E+5yrs, see e.g. Fig.5, the used physical data see at the Table 1). The High Himalayas are bounded by the Main Central Thrust (MCT) in the south and by the Indus-Tsangpo Thrust (ITT) in the north. As indicate the author's model discussed this boundary (MCT) between rocks of the Lesser and Higher Himalayas is very sharp and has a dip of about 30° to 45° to the north. The contact between the crystalline rocks and the underlying sediments is exposed. The striking difference between the sediments north and south of this main thrust is the most important observation in the Himalayan geology and, as we shall see from obtained numerical results, this fact confirms the author's idea that the MCT is a deep fault formed during the first and second evolution periods, i.e. the period of subduction of the Tethys' oceanic lithosphere below the Eurasian plate with relative speeds of the Indian plate to Eurasian plate being more than 8 cmyr^{-1} in the first case and less than 8 cmyr^{-1} in the second one. The part of the MCT is formed by gneisses, migmatites, crystalline schists, thick quartzites and some tectonized granite intrusions. In the upper part of the crystalline thrust mass follow distinct marble and lime silicate horizons with amphibolites and psammite gneisses. The High Himalayan crystalline rocks show progressive regional metamorphism ranging from greenschist to upper amphibolite facies. The grade of metamorphism appears to increase northward from chlorite to garnet grade. In the area near the Nanda Devi, the coarse grained kyanite, garnet and biotite-bearing gneiss of Vaikrita Formation dipping to the north are observed.

North of the Indus-Tsangpo Thrust lies a vast stretch of the Tibetan or Tethys Himalayas. The Tibetan Himalayas are made up of a pile of richly fossiliferous sediments in the basin of the Tethys Sea of the Late Precambrian to Upper Cretaceous ages, folded into a synclinorium during the first two evolution periods of the Himalayas, i.e during the period when the Tethyan litosphere of oceanic type subducted below the Eurasian plate with two different types of speed of subducting Tethyan lithosphere discussed above. The Tethyan sediments occur in several basins that lie along almost the entire length of the northern margin of the Himalayas, and that are abruptly cut by the Indus-Tsangpo Thrust. According to our numerical modelling, the ITT represents a deep fault which originated during the third period of evolution of the Himalayas. It brought up ophiolites and melanges. The ITT is accompanied by the calc-alkaline volcanism from the boundary between colliding Indian and Eurasian plates, represented by pillow lavas, volcanics, and basic and ultrabasic rocks, containing also radiolarite.

Table 1. *Data-profile Nanda Devi (after Bhattacharya, 1981, Bhattacharya, 1992, Chun and Yoshii, 1977, Ni and Barazangi, 1984, Sato et al., 1996, Singh et al., 1990, Verma, 1991 and Wang et al., 1982).*

	1	2	3	4	5	6	7	8	9	10
ρ [kg/m^3]	1300	2280	1400	2100	2000	1780	2500	1770	2100	1680
c_p [m/sec]	3400	7000	3500	6900	6800	5500	7100	5500	6000	5000
c_s [m/sec]	2050	4120	2100	4100	4050	3280	4150	3250	3520	2950

	11	12	13	14	15	16	17	18	19	20
ρ [kg/m^3]	2850	3400	1690	2800	3380	3400	3320	3420	3450	3550
c_p [m/sec]	7500	8500	5000	7300	8450	8200	7500	8400	8500	8800
c_s [m/sec]	4350	5100	2950	4280	5080	5020	4340	5050	5100	5160

	21	22	23	24	25	26	27	28	29	30
ρ [kg/m^3]	2350	3420	2250							
c_p [km/sec]	7200	8300	6300							
c_s [km/sec]	4300	5100	3700							

Geology of the Central Tibet is not so complicated. It is characterized by elongated ranges striking mainly E-W. The young volcanic rocks are observed in the Tibetan Plateau and along the Kun-Lun Range. These very young basaltic lavas of the third period of the evolution of the Himalayas are product of basaltic volcanism with a source at the lithosphere/asthenosphere boundary. With these volcanics are associated older volcanic structures, which represent a relic of basaltic volcanism of the previous evolutionary period of the Himalayas.

The analyses of numerical results show that the seismicity in the investigated Himalayan region will be in the upper parts of the lithosphere (Figs 6a-d), where the lithospheric rocks behave elastically while in deeper parts the lithospheric rocks are partially melted due to change of the deformation energy into a heat, due to change of the effect of Coulombian friction into a heat as well as due to chemical processes taking part in this part of the investigated region. Since the deformation energy accumulated in this region is of very high value, the dissipative heat will be also of very high value and similarly for the effect of Coulombian friction and therefore the rocks of lower parts of the lithosphere are partially melted. Hence the deformation energy accumulated in the lower part of the lithosphere will be released by the creep, by visco-plastic effects, by chemical processes, etc.

The geodynamic model of the Himalayas in its cross-section, profile Nanda Devi, is presented in Fig.4. Fig.5 to Figs.6a-d present vertical component of displacement vector in the detailed area of the High Himalayas, the distribution of principle stresses as well as horizontal, vertical and shear components of stresses in collision zone of the Himalayas for the time period $\sim$ 1.5E+5 yrs. The above-given analyses of our numerical results show that the previous models of the Himalayas (see Figs 7a,b) must be revised. We suggest that the main contact between the Indian plate and Eurasian plate is not the Indus-Tsangpo Thrust (ITT) but the Main Boundary Thrust (MBT). In comparison with the previous models of the Himalayas, presented geodynamic model elucidates the evolution process from the initial state when the Indian subcontinent was a part of the African plate up to the present state. It is evident, that geodynamic processes taking place during the evolution of the Himalayas are from the mathematical point of view similar, or practically the same, as in the case of the evolution of the Central Aleutian arc system, Mediterranean region, Japanese arc system as well as the other collision and subduction regions on the Earth's surface. We see that the mechanism is the same, only the physical parameters and the geometry of regions are different. We also see that our numerical results are in a good agreement with geological observations and/or enable to interpret correctly some geological phenomena that were misinterpreted or unclear. From the presented geodynamic model of the Himalayas we see that all geological phenomena observed on the Earth's surface, like MBT, MCT, ITT, volcanisms, etc., are situated in places where they are observed in situ.

The geographical situation and the major tectonic features are presented at Fig.1. Topographic cross-section along the Nanda Devi profile is given in Gansser (1964). A schematic cross section of the present geometries of converging plates in the Himalaya-Tibetan region (after Ni and Barazangi, 1984), and the numerical model of the same region (after Sato et al., 1996) are presented at Figs 7a,b. Seismicity map for the region investigated is presented at Fig.2 (after Verma, 1991).

5 The Eastern Himalayan Area and Surrounding Areas Tectonic Units

The northeastern part of the Indian plate and northern part of the Burma plate is limited by the area between the latidude 22° to 30°N and longitude 88° to 100°E. The large convergent movements directed from northwest and southeast resulted in information of the Eastern Himalayas with the highest peak Namche Barwa (7755 m) in the north, Naga Hills in the south and Assam-Brahmaputra valley in the middle. Large-scale vertical movements result in the uplift of the Shillong Plateau and Mikir Hills in the central part, deposition of great thickness of sediments, approximately of 12-13 km, in Surma Valley, of 12-15 km in Bengal basin, and of 4-6 km in Assam-Brahmaputra valley. This area is built by the following tectonic units (see Verma, 1991):

(1) Eastern Himalayas,
(2) Mishmi Block,
(3) Naga Hills,
(4) Assam-Brahmaputra Valley,
(5) Shillong Plateau,
(6) Mikir Hills,
(7) Arakan-Yoma fold belt,
(8) Bengal basin,
(9) Surma Valley.

The area of Eastern Himalayas lies approximately $27° - 19°$N and $88° - 96°$E and is characterized by Paleozoic, Mezozoic and Tertiary rocks. The tectonics of the region is dominated by an extensive thrust sheet, which have moved from north towards south-east or south. The outer ranges of the Eastern Himalayas consist of several overthrust units. Major part of the ranges is formed by Tertiary rocks, which are overlaid by a thrust sheet composed of Gondwana strata. This sheet has been overthrust by a sequence of older crystalline rocks.

The Himalayan fold belt lies between 94° to 98°E and has a strike of NE-SW with a sharp turn near 96^0E, where Abor-Mishmi Hills were formed. Near the southern part of Mishmi Hills, the Himalayas bifurcate into two parts-Naga-Luskai Hills tranding NE-SW and Arakan-Yoma fold belt trending NNE-SSW in the northern part and N-S in the southern part. The Naga Hills are formed by a complete Tertiary succession of Eocene and Paleocene rocks.

The Assam-Brahmaputra Valley lies between the Eastern Himalayas and Naga Hills in the eastern part and between the Himalayas and Shillong Plateau in its western part.

This valley has been formed as a result of the transfer of sediments that have their source area in Himalaya, Naga Hills and Shillong Plateau. The main direction of this valley is NE-SW in its eastern part.

The Shillong Plateau and Mikir Hills, the so-called Meghalaya Plateau, lying between 25° to 26°N and 90° to 93°E form a part of northern Indian plate. The height of this plateau is about 1000 m and if is built by an Archean gneiss complex with Proterozoic intracratonic Shillong Series.

The Arakan-Yoma fold belt extends from 16°N to 27°N and consists of a series of thrusts and thrust slices. Rocks in this region represent Paleogene flysch in its western part, whereas Central Burma basin is formed by molasse of the Quaternary age. In Arakan-Yoma suture zone, highly deformed sediments, fossiliferous limestone from Cretacerous-Eocene, deeply exposed Paleozonic metamorphics with ophiolites and also Paleozoic sediments occur.

The Bengal basin is situated south of Shillong Plateau and it is bounded by Tripera fold belt on the east and by Indian Peninsular Shield on the west. It was formed during Eocene to Pleistocene period and comprises sediments of 12-15 km thickness. In the western part, Tertiary rocks are found to rest over older Gondwana formations, whereas in the eastern part, the rocks are folded due to thrust movements directed from east towards west. This part is known as Tripura Fold Belt.

6 Seismotectonic Activity of the NE Indian Plate

The Indian oceanic plate is of an interest for geophysicists and geologists as it can contribute to our understanding of the history of the last period of existence of Gondwana land and the evolution of India, Australia, Antarctica as well as Africa with the island of Madagaskar and small islands in Indian Ocean. From the topographic point of view the Indian plate can be divided into the following topographic elements:

(1) Central Indian Ocean Ridge,
(2) South Eastern Indian Ocean Ridge,
(3) South-Western Indian Ocean Ridge,
(4) Ninetyeast Ridge,
(5) Chagos-Laccadive Ridge,
(6) Broken Ridge,
(7) Kerguelen Plateau in the southern part,
(8) Crozet Plateau in the southwestern part,
(9) Mascarene Plateau with Seychelles Islands in central part,
(10) Chagos-Laccadive Islands on the western side of the Indian Peninsula,
(11) Andaman and Nicobar Islands in the eastern side of the Indian peninsula.

Transform faults represent strike-slip faults along which motion takes place. In a transcurrent fault, shear motion takes place on both sides of the fault indefinitely. However, in a transform fault, the lateral expansion of a ridge ends abruptly at one end of the transform fault and the same continues at the other end. A fault which terminates abruptly with mid-oceanic ridge at both ends we call as a ridge-ridge type of the transform fault. Transform faults have played a major role in time of the evolution of the Indian Ocean. The Chagos-Laccadive-Maldives ridge and Mascarene plateau formed a part of the transform fault connecting the southeast Indian Ridge with the Carlsberg Ridge. Several fracture zones in the Indian Ocean are related to transform faults. These include the Ninetyeast, the Amsterdam fracture zone in the south, the Rodriguez fracture zone in central part and the Owens fracture zone with its continuation known as the Chaman fault in the north. All transform faults are seismically active.

The Ninetyeast Ridge is a transform fault and represents a major topographic feature situated in the eastern part of the Indian Ocean. It extends from 30°S to 9°N and is of several hundred kilometers long. It stopped to be active during the Paleocene-Eocene period. Large magnitude earthquakes of 1928 (magnitude 7.7) and 1939 (magnitude 7.2) have occurred along the northern part of the Ninetyeast Ridge. The analyses show strike-slip faulting. The sense of motion is left lateral.

The Chagos-Laccadive Ridge in its southern part, known as the Chagos Bank, is a region with a high seismicity (magnitude of 5.1 to 6.2). Seismicity along the Chagos-Laccadive area indicates a complex pattern of deformation, as focal mechanism show thrust, normal as well as strike-slip faulting.

The Andaman-Nicobar arc in the northeastern part of the Indian plate, together with the Burmese arc, define nearly a 2100 km long margin of the subducting Indian plate below the Burmese plate. The two arc systems represent an important link between the Himalayan collision zone in the north and a major island arc trench system of southern Asia, the Indonesian arc. Major tectonic elements of the Andaman-Nicobar arc system include the Andaman Trench, Andaman-Nicobar Ridge, Nicobar Deep, West Andaman fault, Andaman Sea and the Mergui terrace. The volcanic arc forms a part of the Andaman Sea. The Andaman-Nicobar islands are built by sediments of the Bay of Bengal.

The seismicity of this region was presented by Mukhopadhyay (1984), who showed that it is seismically very active. This active region extends from 4°N to 20°N and 93°E to 97°E. Seismicity is particularly associated with the Andaman-Nicobar Ridge and the Andaman Spreading Ridge. The Andaman-Nicobar Ridge seismic zone is about 90 km to 160 km wide and it extends eastwards up to the volcanic islands, containing shocks up to 150 km depth. The gravity field is highly anomalous, the free-air values reach values of as much as -190 mgal. The back-arc seismic zone related to the Andaman-Nicobar Ridge is relatively narrow and it shows less intense seismic activity. The activity is well defined along the ridge axis.

6.1 Seismotectonic activity of the eastern Himalayan area and surrounding areas

The Eastern Himalayan Syntaxis joins the Main Boundary Thrust (MBT) in the Himalaya with the Burma Arc. In the Himalayan region, the Indian plate is underthrust from the beginning at small angles, while at the Burma Arc system the Indian subducted plate dips at angle $30° - 60°$ and is seismically active up to the depths of 150-200 km (Holt et al., 1991, Ni et al., 1989). The seismicity of this investigated region is a consequence of collision tectonics in the Himalaya and the subduction tectonics below the Burmese plate.

Most of seismic activity is associated with a few major geological units in this region, such as the Ganges foredeep, the Shillong-Mikir Precambrian plateau, the Kopili lineament, the Bomdila lineament, the Naga Hills, the Assam Valley, the Tripura fold belt. Seismic activity in the Ganga foredeep corresponds to the Tista and to parallel lineaments. The seismic activity then extends northwest to the Sikkim Himalayas, where the Gangtok lineament represents the eastern activity region. The Shillong-Mikir Precambrian plateau is seismically active as a whole. During 86 years, three big earthquakes, the first at 1897 ($m = 8.7$), the second at 1930 ($m = 7.1$) and the last at 1943 ($m = 7.2$), occurred here. In addition there occurred earthquakes of smaller magnitudes. The activity of the plateau is attributed also to its northern lineaments, the most extensive of which is the Brahmaputra lineament. The Brahmaputra lineament crosses the Himalayan foredeep. However, no earthquakes are observed beyond the east edge of the Mikir massif. The Dauki fault and Yamuna lineament, representing the south and west margins of the plateau, are seismically inactive. The Kopili lineament of the Kopili Valley is located between the Shillong and Mikir massifs. It is vary active and the focal depths of earthquakes are of about 70 km. Its activity extends from the Himalayas to the Naga Hills of the Burmese arc system in NNW-SSE direction. The Bombila lineament trends across the eastern Himalayas to the east of the Kopili lineament. It is possible that the lineament extends to the Naga Hills across the Assam Valley through the east edge of the Mikir massif (Nandy, 1980, Rastogi et al., 1973). The lineament is active within the Himalayan domain, and on the last part is inactive. East of the Bombila lineament several thrusts of the eastern Himalayas are highly active. Similarly, the eastern syntaxis of the Himalayas is highly seismically active. The north-western fault system of the eastern syntaxis produced the big Assam earthquake at 1950 ($m = 8.7$) with several of its major aftershocks (Ben Menahem et al., 1974), (Chandra, 1978), (Chhibber, 1934).

The Mishmi thrust, which borders the Assam Valley, is also active, but with epicentral distribution different from those of the neighbouring Himalayan and Burmese arcs. Seismic activity in the Naga Hills is located in the area of their southern thrusts. The Assam Valley, which is situated between the Himalayas and the Burmese arc, is aseismic. The western part of the Burnese arc, the Tripura fold belt, is seismically active and correlates with north-striking structures of the arc, reflecting its westward convexity of the arc. Seismicity of the Bengal basin is located to the Padma, Madhupur, Sylhet and their parallel lineaments. The Madhupur lineament produced 1918 Srimangal earthquake ($m = 7.6$).

6.2 Seismotectonic activity in Burma

East of the Eastern Himalayas and their foredeep is situated Burma and north-east the south-west-south China. This area is located between 19°N and 35°N and between 91°E and 108°E, and it includes several geotectonic units defined by several prominent thrusts and faults.

It was suggested above that the Indian plate actively subducts below the Burnese plate, as shown by an east-dipping Wadati-Benioff zone that extends to about 180 km depth below the central lowlands east of the Arakan-Yoma mountain. Burma is divided into three major parts trending north-south: the Shan plateau in the east, the Central Belt including the basins of the Irrawaddy, Chindwin and Sittang rivers, and the Burmese Fold Mountain Belt in the west. The Fold Mountain Belt includes the Arakan-Yoma mountains and the Chin, Naga, Manipur, Luskai and Patkai Hills. Westwards is situated the Bengal Basin and the Bengal Fan. The Shan plateau, as the easternmost tectonic unit, forms the eastern highlands, which are elevated $\sim$ 700 m above the Burmese plains built up of Tertiary deposits. West of the Shan plateau, Burma is probably formed by the old Gondwanaland. The Mogok metamorphic belt, which is built up of metamorphic rocks with acid intrusives, runs more than 1000 km north-south along the western edge of the Shan plateau. Its width is about 24-40 km. The Mogok Series represent migmatites, garnet-biotite gneiss, biotite gneiss, garnet-graphite-sillimanite gneiss, calcareous rocks such as marble, mafic rocks of the Late Eocene-Early Oligocene age, pegmatites of Middle Miocene age, the Kabaing granites (largest and youngest intrusive) of the Late Miocene-Pliocene age.

The Central Belt and the Burmese Fold (Mountain) Belt form a 1000 km long and 250-700 km wide arc convex towards the foreland. This belt is a product of the Tertiary orogeny, and it can be divided into seven zones, from east to west: the Pegu-Sagaing Rise (a molasse basin resting on a Paleozoic basement), the Central Belt molasse basin (post-Eocene molasse rests on flysch and folded older metamorphics), the Inner Thrust zone (sediments with pre-Alpine metamorphics including Cretaceous-Miocene flysch, resting on abyssal Cretaceous-Mesozoic strata with ophiolites), the coastal Ramri-Andaman Ridge (Cretaceous-Eocene strata, strongly folded), an outer molasse basin (Tertiary sediments about 20 km thick - Tripura fold belt), the Bengal-Surma foredeep (beyond the Indian border) and the Bengal-Assam foreland (Pleistocene deposits). In the Central Belt, young andesitic and basaltic volcanics occur along the active Mt.Popa-Chindwin-Wuntho volcanic system located in south Burma. It continues through the Jade area in north Burma with Cretaceous ophiolites, basic and ultrabasic submarine flows (Chhibber, 1934). The Upper Cretaceous-Eocene flysch of the Burnese Fold (Mountain) Belt is bounded by faults and thrusts on the east and partly on the west side. The Eastern Boundary Thrust is a major tectonic element of the area of ophiolitic rocks and it defines the contact between the Burnese Fold Mountain Belt and the Central Belt throughout Burma.

The Bengal basin is formed by $\geq$ 13 km thick of Cretaceous-Tertiary deposits. The Hinge zone represents the Eocene limestone reflector. The Hinge zone is $\sim$ 500 km long area between Calcutta on the south and the Dauki fault on the north, and it continues into the Bay of Bengal. Its width is between 25 km to 110 km.

The seismicity of Burma

The Naga Hills area, the Chin Hills area, the central portion of the Burmese arc are seismically very active, whereas coastal Burma, where the arc orientation changes to NW-SE, is far less seismic. The earthquake foci range in depth from near surface to arround 180km. The foci distribution in the inclined seismic zone is used here to define the boundaries of the subducting Indian lithospheric plate below the Burmese plate. The lower boundary of the lithosphere is below the lowest foci at a depth of 80 km. Eastward in, this boundary dips to the east in region below the Burmese Fold (Mountain) Belt and reaches a depth of about 180 km the area of the Naga Hills, the Chin Hills and in the central portion of the Burmese arc a depth of about 160 km, while in the coastal Burma only about 100 km. The thickness of the Wadati-Benioff zone below the Central Belt is approximately of 60 km and it has an average dip of about 45° according to several authors, the Burmese volcanic arc is located above the deepest part of the Wadati-Benioff zone. East of the Wadati-Benioff zone within the overriding Burma plate in the Naga Hills, the Chin Hills areas and in the central portion of Burmese arc, two shalower seismic zones are situated. The relatively shallow seismic zone occurs below the Chinswin forearc basin. We speak therefore about the forearc seismicity, whereas in the case of the crustal seismic zone about backarc activity. The backarc seismicity is most intense in the Jade Mines area in north Burma. The forearc and backarc seismic zones are separated by an aseismic zone between them. The island seismic slab underlying the Central Belt is about 50-80 km thick.

The distribution of density in the lower lithosphere and asthenosphere is assumed to be primarily due to temperature perturbations in the upper mantle. Similarly as in the Grow (1973) it is assumed that subcrustal lithosphere has a higher density (~ 3.40 gcm^{-3}) than the upper parts of asthenosphere (~ 3.35 gcm^{-3}). At destructive plate margins at depths below 150 km, the asthenosphere is assumed to be denser than 3.35 gcm^{-3}. In the case of the descending lithosphere, the density is probably higher than 3.40 gcm^{-3} (similarly as in Grow (1973)). These considerations concerning densities in these regions follow from the gravity, as the positive gravity contributions must be compensated by a lower density zone above the descending slab, in order to fit the observed gravity anomalies over the Central Belt and the andensitic volcanic zones. The density in the upper and lower part of the andesitic volcanic zone is from 2.60 gcm^{-3} to 2.80 gcm^{-3}. The density of 2.60 gcm^{-3} for the upper portion of the volcanic arc corresponds to the rock density of the volcanic rocks of the andesitic volcanic zone being measured.

The Indian shield crust underlying the western part of the Bengal basin is almost 30km thick, and it is built of two layers of densities of 2.70 gcm^{-3} and 2.90 gcm^{-3} in its upper and lower parts, respectively. The observed anomalies of density that change in the oceanic crust from 2.90 to 3.40 gcm^{-3} at about 30 km depth can be according to Grow (1973) explained by the phase change in the oceanic crust (i.e. basalt to eclogite) at pressures between 10 to 20 kb (corresponding to 30-60 km depth) in a Wadati-Benioff zone environment.

6.3 Seismotectonic activity in the area of south China

The area is characterized by several geotectonic units defined by the prominent thrusts and faults, shown in Fig.3 of Verma et al. (1980). Major faults are as follows: the Athyn-Tagh, Kansu, Kun Lun, Kang Ting and Red River faults, which run for a few hundered kilometers. Many of these deep faults run in an east-west direction over the central part of Asia and Tibet. Near eastern Tibet, they change their trend to the south-east and to the south in south China. The mechanism of movement is predominantly strike-slip faulting.

A major change in structural trend is observed at the Assam wedge, where the NNE-SSW trending Arakan-Yoma fold belt of west Burma joints the ENE-WSW trending Himalayan ranges. To the east of the Arakan-Yoma ranges is situated the Burmese plain (the so-called Irawaddy basin), with thick sequence of sediments, and the Shan plateau of Burma. To the north-east and east of the Shan plateau, the Hengtnan mountain arc is located, characterized by a north-south strike in regions of Yunnan, Sikang and south-west Szechwan. The tectonics and seismicity of south China are more complex in regions where the north-south structures are cut at right angles by east-west trending mountain ranges. The seismicity is widespread in the whole discussed region. Intense seismicity is observed along the whole Burmese arc, including the Arakan-Yoma mountains and the Burmese plain, the western part of Shan plateau, the Assam wedge, Hengtnan ranges, Kang Ting fault, the Red River fault, Kang Tien ranges and western Szechwan region.

In the Hengtnan ranges and in the east Yunnan-Sikang belt, a combination of thrusting, normal and strike-slip faulting were observed. Over the Yunnan-Sikang belt and the Shan plateau, normal, strike-slip, and thrusting are observed. The majority of the earthquakes in the Yunnan-Sikang region is characterized by shallow foci and therefore that deformation mostly involves the upper part of the lithosphere. The NW-SE trending Red River fault and other parallel faults in south China are seismically active. Ben Menahem et al. (1974) presented the analysis of the Assam earthquake of 1950 and they showed, that it was of strike-slip character. They have chosen the NNW-SSE trending model plane as the fault plane on the basis of aftershock distribution, and moreover, they have suggested that this earthquake was caused by a motion of the Asian plate relative to the eastern flank of the Indian plate, where the northeast Assam geological block has a tendency to rotate.

7 The Geodynamic Model of the Andaman Arc System, Profile Across North Andaman Island, Barren-Narcondam Volcanic Arc System and Burmese Plate

The Andaman-Nicobar island arc system is situated in the northeastern Indian Ocean. This island arc system is 2200 km long and is formed by sedimentary rocks of very young age. Its origin is related to the subduction of the Indian plate below the Burmese plate. The active subduction is documented by the active Barren-Narcondam volcanism (the last eruption of the Barren volcano was in June 1991), by the Wadati-Benioff zone with earthquakes with focal depth upto 200 km, by anomalous geothermal and gravity fields (Dasgupta and Mukhopadhyay, 1993, Mukhopadhyay, 1988, Verma, 1991). The geology and tectonics of this region is characterized by the active faults such as the Semangko fault near Sumatra, the West Andaman fault in the Andaman arc system, the Sagaing transform in the Burmese plate and the Neogene Andaman-Nicobar backarc spreading ridge, where the convergence of Indian plate against Eurasian plate is realized. The greater part of the Andaman Sea is occupied by the Andaman basin.

In this section we present a geodynamic model in a cross-section along regional traverse extending from the Bengal Fan across North Andaman Island and Barren-Narcondam volcanic arc system to the Burmese plate (see Fig.8. Geodynamic model of the North Andaman Island arc system). We use the available geological and seismic data of this region (Curray et al., 1979, Dasgupta and Mukhopadhyay, 1993, Eguchi et al., 1979, Mukhopadhyay and Gupta, 1988, Verma, 1991). The model corresponds to the NNE-trending Andaman arc system where the Barren and Nancondam volcanic islands represent an intraoceanic magmatic arc and where the Alcock Seamount divides the Andaman Sea into fore- and back- arc basins. The thickness of the Indian subducting plate is assumed to be about 80 km, the Burmese plate is then 85 km thick. The Indian plate is subducted below the Burmese plate to the depth of about 160 km in the studied region, of about 180 km in the region of Nicobar island arc system and to 200 km in the region of Sumatra. The dip of Wadati-Benioff zone is of about $40°$ in the studied region and change to $40° - 45°$ in the Nicobar island and Sumatra region. The seismic zone in studied region which displays the back-arc seismicity related to spreading of the Burma plate (Dasgupta and Mukhopadhyay, 1993) is situated relatively shallow below the Andaman Sea.

The density distribution in the lower lithosphere and asthenosphere is assumed to be primarily due to temperature perturbations in an upper mantle, and similarly as in Grow (1973), we assume for the lower lithosphere higher density (3.40 gcm^{-3}) than for upper asthenosphere (3.35 gcm^{-3}) which results from the anomalous gravity field. At depth greater than 150 km, the density is probably higher that 3.40 gcm^{-3}. Even though the density contrast of about 0.05 gcm^{-3} between the dipping lithosphere below the Andaman Sea and its surroundings is relatively small, the positive gravity contribution of the descending lithosphere reaches a maximum of about 20 mgal at the trench axis and 35-40 mgal over the volcanic arc. The observed anomalous gravity field is supposedly a combination of this deep gravity effect and the effect of the overlying layers. According to Grow (1973), this positive gravity anomaly must be compensated by a lower density zone above descending slab (below 75 km) in order to fit the observed anomalous gravity field. For the Bengal basin we assume the crustal layer 30 km thick with densities of about 2.70 gcm^{-3} to 2.90 gcm^{-3} in its upper and lower parts, respectively. Moreover, it has been argued that heating due to friction between the subducting and obducting plates may produce a high-temperature, low-density zone generally below the volcanic arc. For oceanic crust we assume densities between 2.50 gcm^{-3} to 2.90 gcm^{-3}. These densities were estimated on the basis of observed seismic velocities under the Ganges Cone and Andaman arc-trench areas (Curray et al., 1979). The sedimentary layer of the Campanian to Holocene age which underlies this part of the Bengal Fan is assumed to be 4.5-5 km thick; the thickness is quite variable. The velocity of P waves is assumed to be 6.22 kms^{-1}. The thickness of the Cretaceous-Tertiary sediments below the Andaman-Nicobar sedimentary islands (also called as Andaman-Nicobar Ridge) is of about 5-6 km to 13 km. Similarly to Grow (1973), the effect of gravity anomalous field on density gives the density of oceanic lithosphere between 2.90 gcm^{-3} to 3.4 gcm^{-3}. For the northern part of

Indian oceanic crust the seismic velocities of compressional waves were estimated as 5.04 ± 0.69 kms^{-1} to 6.73 ± 0.19 kms^{-1} on the top and bottom of the crust, respectively.

Fig.8 represents the geodynamic model of the subduction of the Indian oceanic plate of Bengal Bay below the Burmese plate, with a presumption (for numerical computation only) that the lithosphere and its support behave elastically. We assume that the time of our investigations is less than a chacterized time $t_c = \nu/\mu$, where ν is the effective viscosity at a given stress level and μ is the elastic rigidity. For mathematical model the author's theory was used (see Appendix). The main results are illustrated in Figs 10-12 for time period 1.5E+5 yrs. Physical parameters were derived from results of authors refered to above and are given in the Table 2. Geometry of the region investigated is based on the gravity and seismic data of this region and use knowledge derived from studied geodynamic problems. Numerical results (the SI system was used) indicate that the horizontal component of displacement has value of about 20.506 mm/yr to 17.946 mm/yr in the Bengal Bay and vertical component has maximum value 19.106 mm/yr to 14.047 mm/yr in the region of Andaman islands, what corresponds to the observed values. Similar results were obtained for time period 1.E+4, where the value of horizontal and vertical components of displacement was about several milimeters per year smaller. The resulting shift in the area of the Bengal Bay is of about 18 mm/yr and the resulting uplift of the Andaman island is of about 20 mm/yr. At Fig.9 the vertical component of displacement vector (time period 1.5E+5 yrs) in detailed area of the Andaman island arc is presented. Fig.10a to Fig.10c demonstrate horizontal, vertical and shear components of stress tensor, while Fig.11 demonstrates the distribution of the principal stresses in the investigated region. The distribution of principal stresses indicate the bulged out regions where the materials are quickly melted. These regions indicate the volcano chambers (at Fig.11 the black areas), namely of the Barren-Narcondam active volcanic arc (the Barren last erupted in June 1991). The places of great compression are indicated below the Andaman island arc system at the depth ~ 25 km and below the volcanic arc system at the depth $\sim 150 - 160$ km. The seismicity map for the region of Burma plate is given at Fig.12. At the end we can remark that in the aseismic zones the volcanic deposits are observed in depth of several kilometers. Since these volcanic deposits are of basaltic type then the magma is originated from the bigger deposit from the contact boundary of the lower lithosphere and the asthenosphere (personal communication of one colleague from the Dept.of Geology, Indian School of Mines, Dhanbad, during the author's lecture at the Indian School of Mines, Dhanbad).

Table 2

	1	2	3	4	5	6	7	8	9
ρ [kgm^{-3}]	2500	2700	2900	2900	3400	3350	3400	2600	2700
c_p [ms^{-1}]	3900	6200	6410	6400	7900	7800	7900	6200	6300
c_s [ms^{-1}]	2140	3600	3700	3700	4560	4520	4560	3600	3700

	10	11	12	13	14	15	16
ρ [kgm^{-3}]	2900	2900	2600	2800	3400	3300	3350
c_p [ms^{-1}]	6500	6450	6200	5950	7900	7780	7850
c_s [ms^{-1}]	3720	3710	3600	3400	4560	4500	4600

Appendix

Mathematically the problem studied is based on the author's theory about simulation of plate tectonics, which mathematically describes the collision and subduction zones in the sense of the new global tectonics (Nedoma, 1990, Nedoma, 1997a). The stresses in the lithospheric plate arise from tectonic stresses, thermally-induced stresses, stresses resulting from the action of gravitational and magnetic body forces, as well as of tidal forces, acting on the density distribution on the lithospheric plate. In the present study the influence of tectonic stresses are assumed only. It is well-known Nedoma (1997a) that for time periods shorter than a characteristic time $t_c = \nu/\mu$ the upper parts of the Earth behaves elastically, while for longer time it behaves thermo-visco-elastically up thermo-visco-plastically. Here ν is the effective viscosity at a given stress level defined by

$$\nu = \left(\tau_2 + \left[(2/3D_{ij}D_{ij})^{\frac{1}{2}}\gamma^{-1}\right]^{1/n}\right) / \left(2^{\frac{1}{2}}3D_{ij}D_{ij}\right)^{\frac{1}{2}},$$

where D_{ij} is the strain rate tensor, τ_2 is the yield stress and $\gamma = 1/L\int_L \gamma_0 \exp(-Q/RT)dx_3$, where Q is an activation energy, R the universal gas constant, L the thickness of the lithosphere and μ is the elastic rigidity. For γ tends to infinity the lithosphere behaves like the rigid perfectly plastic flow. Practically the characteristic time is less than $1.5 - 2.10^5$ years.

It is well-known that the stress field in the lithospheric plate satisfies the equilibrium equations

$$\partial\tau_{ij}/\partial x_j + F_i = 0, \quad i = 1,3, \tag{7.1}$$

where F denotes the body forces vector and τ_{ij} is the elastic stress tensor and where the summation convention is used.

The traction vector on the boundary

$$\tau_i = \tau_{ij}n_j \tag{7.2}$$

can be decomposed into the normal and tangential components

$$\tau_n = \tau_i n_i = \tau_{ij}n_j n_i, \quad \vec{\tau}_t = \tau_i t_i = \tau_{ij}t_i n_j, \tag{7.3}$$

where $\mathbf{n} = (n_1, n_3)$ and $\mathbf{t} = (t_1, t_3) = (-n_3, n_1)$ are the outward normal and tangential unit vector.

The displacement vector can be decomposed into the normal and tangential components

$$u_n = u_i n_i, \quad \mathbf{u}_t = u_i t_i. \tag{7.4}$$

Under our above assumption, the lithosphere and upper mantle behaves elastically, so that our problem can be solved in linear elasticity. Therefore, the stress-strain relation is defined by the Hooke's law, i.e.

$$\tau_{ij} = 2\mu e_{ij} + \lambda e_{kk}\delta_{ij}, \quad e_{ij} = e_{ji} = \frac{1}{2}(\partial u_i/\partial x_j + \partial u_j/\partial x_i), \tag{7.5}$$

where λ and μ are Lame's coefficients, δ_{ij} is Kronecker's symbol and e_{ij} is the small strain tensor. From (7.1) and (7.4) we obtain the well known equilibrium equation for the elastic rheology. The elastic coefficients λ and μ are positive, $\lambda(\mathbf{x}) > 0$ and $\mu(\mathbf{x}) \geq \mu_0 > 0$, almost everywhere in the plate and the upper mantle and therefore fulfil the condition

$$\frac{1}{2}(\lambda e_{kk}^2 + 2\mu e_{ij}e_{ij}) \geq \mu_0 e_{ij}e_{ij}, \tag{7.6}$$

almost everywhere in the investigated region, which is denoted as $\Omega \subset R^2$.

We shall assume that $\Omega = \bigcup_{\iota=1}^{s} \Omega^\iota and$ that the boundary $\partial\Omega$ is sufficiently smooth and consists of three parts Γ_τ, Γ_u and Γ_c, i.e. $\partial\Omega = \Gamma_\tau \bigcup \Gamma_u \bigcup \Gamma_c$. We consider the following boundary conditions:

On the Earth's surface Γ_τ the loading is prescribed, i.e.

$$\tau_{ij}n_j = P_{0i}. \tag{7.7}$$

On the boundary Γ_u, which limited the investigated domain Ω, the displacement vector $\mathbf{u}$ is prescribed. To model the shifting of the moving lithospheric plate we will determine the value of the displacement vector on the part of boundary Γ_u, which corresponds to a motion of the invasing plate, at a distance sufficiently far from the collision or subduction zones. The values of the displacement vector components can be determined from our knowledges of the spead of the invasion plate per year and/or of the recent (horizontal and vertical) movements. Thus

$$\mathbf{u} = \mathbf{u}_0 \quad \text{on } \Gamma_u, \tag{7.8}$$

where $\mathbf{u}_0 \neq 0$ simulates the shift of the invasion lithospheric plate in time.

On the contact boundary Γ_c between colliding lithospheric plates and blocks we have the following conditions:

$$u_n^k - u_n^l \leq 0, \tag{7.9}$$

representing the condition of nonpenetration of colliding lithospheric plates and blocks,

$$\tau_n^k = -\tau_n^l \leq 0, \tag{7.10}$$

representing the condition for contact forces (it follows from the principle of action and reaction and from the fact that the normal contact component cannot be tensile), and the condition

$$(u_n^k - u_n^l)\tau_n^k = 0. \tag{7.11}$$

If the colliding lithospheric plates and blocks are in a contact then $(u_n^k - u_n^l) = 0$ and $\tau_n^k < 0$, i.e. there exists such nonzero contact forces τ_n^k that both colliding plates are in a contact. If the colliding plates are not in contact then $(u_n^k - u_n^l) < 0$ and then the contact forces $\tau_n^k = 0$. From the thermodynamics it is known that the temperature of melting depends on the pressure. If the pressure increases the temperature of melting also increases, if the pressure decreases the temperature of melting also decreases. These facts represent the mechanism of the origins of volcano chambers in collision or subduction regions. Moreover, if both colliding plates are in a contact, then a Coulombian friction acts on the contact bounadary Γ_c. The frictional force g_c^{kl} acting on the contact boundary Γ_c is proportional, in its absolute value, to the acting normal forces i.e. to $|\tau_n^k|$ and the coefficient proporcionality is the coefficient of Coulombian friction F_c^{kl}. Then

$$\text{if} \quad u_n^k - u_n^l = 0, \quad \text{then} \quad |\vec{\tau}_t^{kl}| \le F_c^{kl} |\tau_n^k| = g_c^{kl}, \quad \text{on} \quad \Gamma_c^{kl}, \tag{7.12}$$

and

$$\text{if} \quad |\vec{\tau}_t^{kl}| < F_c^{kl} |\tau_n^k| = g_c^{kl} \quad \text{then} \quad \mathbf{u}_t^k - \mathbf{u}_t^l = 0, \tag{7.13}$$

$$\begin{aligned} &\text{if} \quad |\vec{\tau}_t^{kl}| = F_c^{kl} |\tau_n^k| = g_c^{kl} \\ &\text{then there exists nonnegative } \theta \ge 0 \text{ such that } \mathbf{u}_t^k - \mathbf{u}_t^l = -\theta \vec{\tau}_t^{kl}, \end{aligned} \tag{7.14}$$

i.e. the mutual schift of both colliding plates or blocks, resp., is proportional to the acting tangential forces and acts in the opposite direction as these acting tangential forces. We see that if the absolute value of the tangential forces is less than the friction forces g_c^{kl}, then the tangential forces preclude the mutual motion of both colliding plates (see (7.13)) and in the plates the deformation energy cumulates. If the absolute value of the tangential forces reach the value of friction forces g_c^{kl} (see eq. (7.14)), then there are no forces which can preclude the mutual (bilateral) motion of both colliding plates. If $F_c^{kl} = 0$ then we have the case without Coulombian friction. The Coulombian coefficient of friction can be estimated by

$$\| F_c^{kl} \| < \min_{\Omega} (2c_s/(c_p + c_s))^{\frac{1}{2}} \;\text{ or }\; \| F_c^{kl} \| < \min_{\Omega} (c_s/c_p)^{\frac{1}{2}}, \;\text{ respectively.} \tag{7.15}$$

To solve the problem numerically we minimize the functional of potential energy $L(\mathbf{u})$ over the FEM approximation of the set of addmissible displacement $K_h = \{\mathbf{v} | \mathbf{v} \in V_h, v_n^k - v_n^l \le 0\}$, where V_h denotes the FEM approximation of the space of virtual displacements. The problem leads to solve numerically the variational inequality. For more details see Nedoma (1987), Nedoma (1994) or Nedoma (1997a, Chap.VII).

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Fig.1 *Map of Himalayan Arc, Tibetan Plateau, and surrounding areas (after Ganser, 1977, Molnar and Tapponier, 1975 and Ni and Barazangi, 1984).*

Fig. 2 *Seismicity map for the area covering the Himalayan Arc System in its western part.*

Fig.3 *Map of the lower boundary of the lithosphere.*

Fig.4 *Geodynamic model of Himalayas, profile Nanda Devi.*

Fig.5 *Vertical component of the displacement vector in the area of the High Himalayas-profile Nanda Devi.*

Figs 6a-d *Principle stresses and horizontal, shear and vertical components of stresses-profile Nanda Devi.*

Figs 7a,b a) *A schematic cross section of the present geometries of converging plates in the Himalaya-Tibetan region (after Ni and Barazangi, 1984),* **b)** *Numerical model of the same region (after Sato et al., 1996).*

Fig.8 *Geodynamic model of the North Andaman Island Arc System.*

Fig.9 *Geodynamic model of the North Andaman Island Arc System; vertical component of displacement u in the detail area of the Andaman island arc system.*

Fig.10a-c *Geodynamic model of the North Andaman Island Arc System; horizontal, vertical and shear components of stresses.*

Fig.11 *Geodynamic Model of the North Andaman Island Arc System; distribution of principal stresses.*

Fig.12 *Seismicity map of the Burma Plate and Surrounding Areas (after Verma, 1991).*