Chapter 7

Tectonics Of Manipur

It is useful to be assured that the heavings of the earth are not the work of angry deities. These phenomena have causes of their own

Seneca (4 B. C.- 65 A. D.)
Manipur Hills lie between the Naga-Patkoii Hills on the north and north-east and the Chin Hills on the south forming an integral part of the Indo-Myanmar Range (IMR). The structural and tectonic pattern of the region is transitional between NE-SW trending pattern of the Naga-Patkoii Hills and the N-S trending pattern of the Mizoram and Chin Hills. The rocks of the region is principally made up of Tertiary sediments with minor igneous and metamorphic rocks associated with deep sea sediments such as cherts, red and black clays/shales, limestones and flysch sediments. The general strike of the lithounits, structures and tectonic lineaments in the region is NNE-SSW although, it varies between N-S and NE-SW. The geological and tectonic setting of the state is similar to that of the IMR. So, we will discuss about the tectonic setting of this range briefly in detail. But, before not going into the details of tectonic accounts of the state and that of the IMR, the tectonic setting of Northeast India, its gravity and seismic activity will be described very briefly, for these have a direct bearing in the evolutionary history of the region also.

7.1 TEUTONIC SETUP OF NORTHEAST INDIA

The tectonic setting of Northeast India in general and that of the IMR in particular has a distinct entity in the tectonic framework of Southeast Asia, for it represents a collision mechanism between the Indian plate and the Eurasian (Myanmar) plate having an island arc collision characteristics in the IMR. Almost all along the northern margin of the Himalayas and the eastern margin of the IMR, ophiolites of late Mesozoic do occur along tectonised belts known as Indus-Tsangpo Suture Zone and Naga-Manipur-Chin-Arakan-Yoma Suture Zone. These narrow ophiolite belts are obducted remnants of the Tethyan oceanic crust resulted from the collision between the Indian-Australian and Eurasian plates. It is believed that the
IMR is the product of an island arc type of collision between the Indian and Myanmar plates while the eastern Himalayas evolved as a result of continent-continent collision between the Indian-Australian and Eurasian plates.

7.1.1 Tectonic Provinces of Northeast India and Adjoining Region

Geologically and tectonically Northeast India and the adjoining region can be divided into a number of belts or provinces. Various tectonic subdivisions have been made by Evans (1964), Raju (1968), Verma et al (1976), Nandy (1980), etc. The following subdivisions are adopted from Acharyya et al (1986) and shown in figure 7.1.

(i) Eastern Himalayas
(ii) Shillong-Mikir Massif
(iii) Tsangpo Ophiolite Zone and Trans Himalayan Magmatic Arc
(iv) Mishmi Hills
(v) Assam Shelf
(vi) Outer Molasse Basin (Naga Foothills and Surma Valley)
(vii) Indo-Burman Range (Indo-Myanmar Range)
(viii) Central Burma Molasse Basin
(ix) Highlands of Eastern Burma

The Eastern Himalayas lying mainly between the Brahmaputra on the south and Tsangpo on the north constitutes the Arunachal Himalayas. Like the Western Himalayas, it can also be divided into Sub, Lesser, Great/Central and Tethyan/Trans-Axial Himalayas. A number of important mega lineaments or structures such as Main Boundary Thrust (MBT), Main Central Thrust (MCT), etc. characterise this tectonic province. Shillong and Mikir Massif is thought to be the north-eastern prolongation of the Indian shield and the eastern margin of the continental mass of the shield. The massif is principally made up of high grade metamorphic basement gneissic rocks underthrusting the “belt of schuppen” in the east and limited in the south by the Dauki fault. Tsangpo Ophiolite Zone and Trans Himalayan Magmatic Arc belt extends...
Fig. 7.1. Structural and tectonic map of Northeast India showing various tectonic provinces (after Acharyya et al., 1986).

LEGEND

- Thrust and Thrust Sheet
- Fault and Fracture
- Fold Axis
- Ophiolite Mélange
- Plutons (B Basic, A Acid)
- Volcanics (B Basic, A Acid)
- Archaean
- Proterozoic
- Himalayan Central Crystalline
- Devonian
- Carboniferous
- Permian
- Carboniferous-Permian with Eocene
- Mesozoic
- Triassic
- Jurassic
- Cretaceous
- Palaeogene
- Neogene
- Quaternary
- Neogene Quaternary
between the Tethyan Himalayas and the Tibetan plateau. This unit represents the eastern part of the Indus Tsangpo (Zangbo) Suture Zone which is thought to be characterised by the Main Mantle Thrust (MMT) (see Coward, 1983). Magmatic Arc mainly comprising volcanic/plutonic acid rocks lies on the northern side of the ophiolite belt. *Mishmi Hills* occur at the junction of the Eastern Himalayas and the IMR. There are conflicting views on the evolution of the NW-SE trending Tidding Suture and its serpentinites, metavolcanics, metasediments and the huge Mishmi diorite-granodiorite batholithic complex (Acharyya, 1982; Chattapadhyay and Chakraborty, 1984; Nandy, 1980, 1982, 1983). The rocks of the Lesser Himalayas and the Central Crystallines appear to be greatly attenuated and truncated in the Mishmi Hills by Lohit thrust, the Tidding Suture and the Mishmi frontal thrust zone. The Mishmi frontal thrust zone overrides the Quaternary sediments in its foothills and ultimately truncates the fold and thrust systems of the IMR.

*Assam Shelf* lies on the flanks of Shillong and Mikir massif sloping towards south-east and north-east. It is believed to have a widespread Gondwana sediments as evident from the abundant presence of Lower Gondwana miospores in the younger Tertiary sediments. Cenozoic sedimentation on the shelf began with deposition of Jaintia Group of Paleocene-Eocene age equivalent to the Disangs. *Outer Molasse Basin* occupies the outer Naga foothills, Surma Valley (Basin) and Tripura, and better developed in the two latter regions. It comprises the Surma Group, the oldest unit of the molasse basin, resting unconformably upon the Barails, and the younger Tipam Group of rocks. The molasse sediments of the Surma Valley is characterised by N-S trending folds known as frontal folds. *Central Burma (Myanmar) Molasse Basin* lies between the IMR on the west and the Shan Scarp (Sagaing Fault), forming the western margin of the Eastern Burma (Myanmar) Highland, on the east. It is divided into western trough (Chindwin Basin) and eastern trough (Irrawaddy Basin) by a lineament known as Central Burma Volcanic Line (Myanmar Volcanic Arc). It consists of Mio-Pliocene molasse sediments overlying unconformably the Naga Metamorphics, ophiolites and Cretaceous-Eocene sediments. Along the Central Volcanic Line, Middle Cretaceous tonalites intrude basaltic andesites. Lower Eocene quartz-diorite stocks and andesitic dykes are also found (see Mitchell and Mckerrow,
1975). Highlands of Eastern Burma (Myanmar) comprises the east Burma Highlands, Shan plateau and peninsular Myanmar-Thai-Malaysia group and lies on the eastern side of the Shan Scarp. The Myitkyina and Mandalay ophiolites occur along the northern part of the tectonised margin between the Central Burma Molasse Basin and the Eastern Burma Highlands. Serpentinites, gabbro, amphibolite and basaltic andesite are also found. Sedimentary rocks include limestone, calcareous grit, sandstone, shale, and all are associated as melange.

Detailed descriptions and accounts of the tectonic provinces mentioned above are beyond the scope of the present work. So, the reader(s) can refer for detailed description about the tectonic setting of Northeast India from the literature cited above as well as from the followings e.g. Mathur and Evans (1964), Bhattacharjee (1987), Mitchell (1993), etc.

7.1.2 Gravity Anomaly in Northeast India.

In Northeast India seismic profiles of oil exploration or of any other investigation purpose is practically negligible or inaccessible. So, subsurface interpretation of geological and tectonic structures in the region is still primarily dependant on geophysical data such as gravity anomaly and seismic (earthquake) activity only. Evans and Crompton (1946) probably were the first workers to make gravity interpretation in relation to structural and tectonic framework of the region. Later, Verma et al (1976), Verma and Mukhopadhyay (1977), Nandy (1986), Mukhopadhyay (1988), Mukhopadhyay and Dasgupta (1988), Kumar (1981, 1990) have also made similar interpretations about the structural and tectonic framework of the Northeast India and that of the Andaman region using either gravity or seismic or both gravity and seismic data. A gravity anomaly map of the region is shown in figure 7.2. Various types of gravity anomaly can also be obtained from Gulatee (1956). Detailed study and interpretation of the different sectors of the IMR and that of Northeast India can be referred to the literature mentioned above. Interpretation made here is a very simple one only to show the relationship between the tectonic analysis
Fig. 7.2. Bouger gravity anomaly map of Northeast India and adjoining region (after Nandy, 1986). Note the three broad patterns of gravity anomaly - an E-W trend corresponding to the Eastern Himalayas, a N-S trend corresponding to the Indo-Myanmar Range while the central close pattern corresponds to Shillong-Mikir Massif (see text also).
conducted and geophysical data of the region. Based on the map shown (Fig. 7.2), the following three gravity patterns can be described in the area.

(i) An east-west trending pattern, (ii) A north-south trending pattern, and (iii) A central gravity high pattern.

(i) An East-West Trending Pattern

This gravity pattern follows the structural and tectonic lineament trends of the Eastern Himalayas. From south it gradually increases towards north to a maximum of about -550 mGals on the northern side of Tsangpo in the Trans Himalayan region. The excessive negative value is probably resulted from the thickening of the crust due to underthrusting of Indian landmass below the Tibetan landmass. There is no other abnormal behaviour in the pattern indicating any tectonic significance.

(ii) A North-South Trending Pattern

The pattern, like the one described above, follows the structural and tectonic lineaments of the Indo-Myanmar Range and the Myanmar Volcanic Arc. The value gradually decreases towards east to a minimum of -175 mGals along the Chindwin river (Chindwin Basin). This probably coincides with the residual forearc basin where crustal thickening is resulted from the molasse sediments and the continental crust below. The increase in gravity further east is due to the volcanic arc where the value approaches -25 to 0 mGals. Then in the east, it further decreases to about -100 mGals corresponding to the Irrawaddy Basin covered by the Neogene molasse sediments. Although, local gravity anomaly variation occurs, in the regional scale there is no fluctuation in the values of anomaly around the Ophiolite Melange Belt indicating that these are basically rootless bodies emplaced and associated with the lighter sediments of the IMR.
A Central Gravity High Pattern

This coincides with the Shillong and Mikir massif. It has a maximum value of +40 mGals on the south-western part of the Shillong massif. On the northern side the value gradually decreases showing sloping continuation of the massif. On the eastern side it causes some contortions possibly revealing the underthrusting nature of the massif below the IMR sediments. On the southern side, the close spacing of the gravity contours suggests abrupt ending of the massif probably along the Dauki fault. Models for the crust and mantle configuration of the Shillong plateau, Assam valley and Bengal basin have been discussed in detail by Verma and Mukhopadhyay (1977).

7.1.3 Seismic Activity in Northeast India

Seismic study is another geophysical tool that can be used for understanding the tectonic mechanism undergoing between the Indian-Australian and Eurasian plates rather Indian-Myanmar plates. A number of research papers have been published dealing with seismic activity of Northeast India, some of which have already been cited in chapter 4. Barazangi and Dorman (1969) showed that Northeast India in general and the IMR in particular is mainly characterised by earthquake hypocenters of intermediate depth. So, it is suggestive of an island arc type of convergence mechanism between the Indian and Myanmar plates (Verma et al., 1976; Saikia et al., 1987; etc.). Subduction of Indian plate below the Myanmar plate at about 30°-60° along the Myanmar-Andaman Arc is also discussed and interpreted by a number of other workers e.g. Kumar (1981, 1990), Nandy (1986), Verma and Krishnakumar (1987), Mukhopadhyay and Dasgupta (1988), Dasgupta and Mukhopadhyay (1993), Kayal (1995), etc. Relationship between these seismic studies and gravity constraints of the region has also been discussed by them. The underthrusting nature has been studied earlier by Füch (1970) also. From all these observations it is evident that Indian plate is underthrusting the Myanmar plate and the nature of convergence is similar to that of an island arc type. So, the IMR was probably evolved as an accretionary prism (James et al., 1989). However, there are contrasting opinions in regard to the maximum compressive stress orientation in the
Fig. 7.3. Seismicity map of Northeast India showing distribution of earthquake epicentres. Note the maximum concentration of epicentres along the Indo-Myanmar border below the 25°N latitude (courtesy Earthquake Monitoring Project, Dept. of Earth Sciences, Manipur University).
Fig. 7.4. Seismicity map of Northeast India showing distribution of hypocenters. Note the maximum abundance of shallow to intermediate depth earthquakes with a few deep ones (courtesy Earthquake Monitoring Project, Dept. of Earth Sciences, Manipur University).

**SEISMICITY IN NE-INDIA**

*(DEPTH WISE DISTRIBUTION)*

**CHINA**

**BANGLADESH**

**MYANMAR**

**Scale:** 1 cm = 121.7 km

**RANGE OF DEPTH**
- ○ 187 to 232 (9)
- ○ 141 to 187 (13)
- ○ 96 to 141 (17)
- ○ 51 to 95 (54)
- ● 5 to 50 (450)
- ● all others (24)

**SOURCE:** SEISMICITY DATA FROM RRL, JORHAT
(From 1622 to 1987)

**PROCESSED AT DEPARTMENT OF EARTH SCIENCES, MANIPUR UNIVERSITY**
subducting plate (James et al., op. cit.; Verma and Krishnakumar, op. cit.). E-W compression stress in Northeast India is mentioned by Kayal (op. cit.), but NE-SW to N-S maximum compression stress is proposed by Rajendran and Gupta (1989) in compatibility with the Indian plate motion. However, they also point out the possibility of having variable stress orientation from place to place due to microtectonics along the Myanmar-Andaman Arc. The present work, as already pointed out in the plate kinematics, suggests an E-W compression and N-S extension which in turn can explain the structural and tectonic styles of the Indo-Myanmar Range.

Seismicity maps of Northeast India and adjoining regions have been shown in figures 7.3 and 7.4 depicting the epicentral and hypocentral distribution in different parts of the region. Maximum concentration of earthquake epicentres is found along and slightly on the eastern part of the IMR and the hypocenters are basically moderate in depth i.e. principally shallow to intermediate depth with a few deep earthquakes extending below 150Km. These features clearly support the concept of subduction of the Indian plate below the Myanmar plate similar to a modern island arc subduction type (see also section 7.4) where the earthquake foci are principally distributed in the subducting Indian plate. On the other, earthquake hypocenters are chiefly of shallow depth in the Arunachal Himalayas probably reflecting continent to continent collision in this sector of Northeast India. So, the ideal subduction type convergence doesn’t extend beyond the north of 26°N latitude (Kayal, op. cit.). Therefore, ideal structural and tectonic features associated with island arc type subduction may be found only below i.e. towards south of this latitude.

Figure 7.5 represents a preliminary outcome of Earthquake Monitoring Project undertaken by the Dept. of Earth Sciences, Manipur University (see also Kumar and Sanoujam, 1998) showing the relations between the earthquakes and the lineaments of Manipur state in order to decipher the active fault zones. It shows distribution of epicentres around and along the major lineaments (compare also with Fig. 3.1) revealing almost all the lineaments oriented in the directions comparable to those discussed in chapter 3 are still likely active. Even the Imphal Valley, covered by
Fig. 7.5. Seismicity map of Manipur and its adjoining region showing the relationship between the structural and tectonic lineaments, and distribution earthquake epicentres. From the figure, it appears that still a number of the lineaments are seismically active. Low frequency of the epicentres in the western part of the state is due to lack of seismic stations (courtesy Earthquake Monitoring Project, Dept. of Earth Sciences, Manipur University).
alluviums, shows good distribution of epicentres. The low frequency of epicentral points, on the western side of the state, does not necessarily indicate low activity but lack of monitoring stations. From the above observations, it can be evidently inferred that almost all types of faults i.e. normal, strike-slip and thrust faults can still occur in the region. Such findings, along the IMR, have also been provided by Kayal (op.cit.), Rajendran and Gupta (op.cit.), Kumar (op.cit.) as mentioned above.

7.2 TECTONIC SETUP OF INDO-MYANMAR RANGE

Indo-Myanmar Range (IMR) is thought to be the northern prolongation of the Indonesian island arc. It links up with the eastern end of the Himalayas probably along the Tidding Suture zone. It is an arcuate shaped tectonic belt having a convexity towards west. Longitudinally it is divided into three segments (Brunnschweiler, 1974; see also Acharya, 1991). But the basis of dividing the range into three segments is not clearly described although, it appears to be based on structural and tectonic patterns of the segments. Each of the segments is 400Km long approximately and from north to south represented as

(i) Naga Hills, (ii) Chin Hills and (iii) Arakan-Yoma Segment

A comprehensive account about the segments in terms of their geomorphic features, time relationship, structural and tectonic evolutionary trends can be had from Brunnschweiler (op.cit). A brief description on the geological and tectonic setting of each segment is provided below.

7.2.1 Naga Hills

This unit, trending nearly NE-SW, is the northernmost segment of the range. It terminates against the continental mass of the Mishmi Block/Hills. The segment principally comprises of Naga-Patko Hills of Nagaland and northern part of the Manipur Hills. The region has been described in terms of three major lithostratigraphic units viz. Naga Metamorphic Complex, Naga Hills Flysch and
Upper Chindwin Molasse Basin (Chindwin Basin) by Brunnschweiler (1966). A comprehensive account about the Naga Hills orogenic belt is also given by Roy and Kacker (1982), Acharyya et al (1986). The first two units mainly constitute the Naga Hills while the third one is mainly northward prolongation of the Central Burma Molasse Basin. Acharyya et al (op. cit.) has described the geological and tectonic setting of this segment of the IMR in terms of two distinct longitudinal belts. They are:

(a) Central Naga Hills Palaeogene Flysch Sediments and
(b) Naga-Chin Hills Ophiolite Belt

(a) **Central Naga Hills Palaeogene Flysch Sediments**

This unit occupies the western part of the belt and mainly composed of folded and faulted thick pile of monotonous flysch sediments of the Disangs succeeded by the sandy flysch to molasse Barails. The outer (western) contact of this tectonic unit with the sediments of the Outer Molasse Basin is marked by a thrust known as the Disang Thrust. In Manipur it is marked by the Nungba Thrust because, beyond this thrust (i.e. towards west) it is practically composed of the sediments of the Surma Basin. The thick monotonous argillaceous sequence of the Disang Group has turbidite character specially at the upper part, near the contact with the Barails. There is lithological and faunal continuity between the two groups of rocks. Fragmentary plant and faunal remains from the Upper Disangs and Lower Barails suggest a shallow, near shore, marine, intertidal, estuarine or deltaic environment of deposition. In the distal part, characterised by dark grey shales, relatively deeper, marine depositional environment is evident. In eastern Manipur, beds, lenses and blocks of Upper Cretaceous limestone occur in association with shales of Eocene age.

The Barails characterised by alternate beddings of arenaceous and argillaceous sediments have turbidite characters. These are associated with carbonaceous matters occasionally, and probably deposited in a relatively shallow water environment. Because, they usually show well developed primary sedimentary structures. In the Kohima synclinorium the sediments are folded and faulted in a complex manner.
(Acharyya et al, 1986) while the unconformably overlying younger sediments are open folded. The latter is part of the Outer Molasse Basin as already described in the previous section.

(b) Naga-Chin Hills Ophiolite Belt

The Naga-Chin Hills Ophiolite Belt (NACHO) which can also be referred as the Naga-Manipur-Chin Hills Ophiolite Belt (NAMCHO), is principally composed of tectonised and dismembered ophiolite complex, associated with deep sea sediments such as cherts, limestones, shales and sandstones. It thrusts over the younger Disang sediments to the west, while to the east, these ophiolitic rocks and their sedimentary envelope are overthrust at places by the Naga Metamorphics or their equivalents. Further east, the metamorphics are overlain unconformably by the sediments of the Central Burma Molasse Basin (Chindwin Basin). In Manipur and Nagaland states, the ophiolite belt occurs as a NNE-SSW trending, westerly convexed arcuate body of about 200 Km long from north-east of Chokla in Nagaland to Moreh in Manipur, with an average width of about 10-15 Km.

Radiolaria, calcareous nanно fossils and foraminiferal assemblages from cherts and limestone beds, intimately associated and interbedded with the ophiolite complex and occurring as slivers elsewhere, have been dated as Maestrichtian. The long ranging radiolarian assemblages may include older beds. The dismembered ophiolitic suite is nonconformably overlain by the newly christened Phokpur Formation (Acharyya et al, 1986). It is principally made up of ophiolite derived, volcanoclastics, marine to paralic sediments which have been dated as Eocene. The Phokpur Formation is folded, imbricately thrust faulted with the ophiolitic rocks, overridden by the Naga Metamorphics and feebly metamorphosed at places.

7.2.2 Chin Hills

The Chin Hills, trending nearly N-S, lie between the Naga-Patkoi Hills on the north and the Arakan-Yoma Segment on the south. It has a more or less similar
tectonic setting to that of the Naga Hills. The unit is also principally composed of flysch sediments with minor igneous and metamorphic rocks. On the southern part of the segment, in Kanpetlet area, a group of schistosed rocks do occur which is known as Kanpetlet schists. It overthrusts the lower Tertiary unmetamorphosed shales and sandstones with conglomeratic layers to the west. It is usually thought to be equivalent to the Naga Metamorphics and forms the Pre-Mesozoic basement complex, brought up and emplaced tectonically on to the sediments of the IMR (Brunnschweiler, 1966).

Basic and ultrabasic complexes having coarse texture, peridotite to gabbroic in composition have been reported by Brunnschweiler (op.cit.) although, he did not regard them as ophiolites. Probably this is because of the relatively fresh and clear intrusive character of these rock masses. So, the same writer (1974) described this segment of the IMR without ophiolites but with exotics in the flysch sediments and unknown in the Miocene strata. This could be, probably, the basis on which the two segments - Naga Hills and Chin Hills were separated. But, these basic and ultrabasic complexes do occur in the same strike continuity of the ophiolite belt of Manipur and Nagaland, and the limestone exotics also occur along the same belt. Moreover, if the basic and ultrabasic complexes form the basement rocks, there is a likely chance of widening with depth, that in turn will create a gravity high pattern, which is not so, as evident from figure 7.2. So, probably these are ophiolitic mass lying as a tectonic slice overthrusting the younger sediments as in the case of the previous segment i.e. the Naga Hills.

In the central Chin Hills, on the eastern side of the Kennedy peak, large exotics of flysch containing Globotruncana bearing limestones are found which are similar to the olistostromal limestones of Ukhrul and Chandel districts of Manipur. Similar exotics are found again in a similar strike continuity. So, the two segments have a similar tectonic setting.
7.2.3 Arakan-Yoma Segment

This segment lies on the southern side of the Chin Hills comprising relatively low hills and coastal areas of Myanmar. The tectonic setting of this segment of the IMR is also more or less the same with the other two although, the general strike of the tectonic lineaments is NNW-SSE. Ophiolitic rocks are relatively less and usually form small outcrops mainly on the eastern side of the IMR and western part of the Chindwin Basin.

In the coastal areas near Sandoway the oldest stratigraphic record is represented by strongly folded, thinly bedded alternations of fine-grained sandstones having occasional gritty layers with grey and dark shales, marly shales and mostly light coloured limestones with Globotruncana. These are of Late Cretaceous age and associated with grey and reddish radiolarian cherts and silicified radiolarian limestones at regular intervals. Igneous rocks of ophiolitic nature are also found associated with tuffs and agglomerates. Constituents of the agglomerates are mainly derived from immediately underlying part of the Upper Cretaceous sequence and usually small, upto fist size (Brunnschweiler, 1966). These sediments are unconformably (partly conformable) overlain by sand and gritstone with associated clasts containing basic rocks that are similar to the wild flysch of the Swiss Alps. They are characterised by intraformational slipping and thrusting which is possibly similar to the gravity sliding or slumping of the Disangs as already discussed. Two distinct structural trends have been observed where the older Upper Cretaceous sediments have a N-S structural trend while the younger lithounits show NW-SE trend.

In Cheduba island area, the rocks seem to be younger than those found in any of the Sandoway area. These are chiefly composed of a sequence of grey, clayey, mudstones and siltstones with sandy horizons. The sandy horizons, a few metres thick, have hydrocarbon potentials. The rocks of the island are characterised by anticlinal ridges and synclinal valleys as in the case of the Surma basin. In Ramri island, the geological sequence is more or less the same to that of the Cheduba, but
both older and younger rocks are found in this area. However, the rock types and other tectonic parameters suggest a similarity to those of Bhuban Formation of Northeast India. Like the other parts of Arakan-Yoma, a number of exotics are found within these lithounits which cause many geologists to misinterpret the geology of this region.

From the deliberations made above it is evident that the entire length of the Indo-Myanmar Range has a more less similar geological, structural and tectonic characters. It can simply be described as a tectonic belt that has either sedimentary melange (olistostromal) or tectonic melange or both, embedding within the sediments of the range suggesting a very interesting and common tectonic setting.

7.3 TECTONIC SUBDIVISIONS OF MANIPUR

It has been mentioned that Manipur lies between the Naga-Patkoi Hills on the north and Chin Hills on the south forming an integral part of the IMR. We have also made discussions on the broad tectonic framework and subdivisions of Northeast India and the IMR. Tectonostratigraphic sequence of the state in relation to the tectonic setting of the IMR has also been outlined in chapter 2. In this section, it is, therefore, attempted to make a tectonic subdivision of the state. Because, it will provide a better understanding about the relationship of various lithostratigraphic units in terms of the tectonic setting of the region. The tectonic subdivision is made considering all the geological, structural and tectonic features of each of the units. They are given below and shown in figure 7.6.

(i) Metamorphic Belt
(ii) Ophiolite Melange Belt
(iii) Central Manipur Flysch Belt
(iv) Western Manipur Molasse Belt
(v) Imphal Basin
A brief discussion on each of the tectonic units is presented below with certain justification of their subdivision.
Fig. 7.6. Map showing tectonic subdivisions of Manipur.
7.3.1 Metamorphic Belt

This unit has been described as Naga Metamorphic Complex and its brief geologic setting and description have been provided in Chapter 2. It lies on the easternmost part of the state and occupies a small areal extent. Larger outcrops of the complex occur in the territory of Myanmar and Nagaland state of India on the north-eastern border of the state.

It occurs as klippen overlying the younger Ophiolite Belt and its associated sedimentary rocks with a moderately high angle thrust fault called Metamorphic Thrust dipping towards east. On the eastern and south-eastern sides it is unconformably overlain by the sediments of Chindwin Basin. It has distinct characters from the underlaying as well as overlaying lithounits as characterised by medium grade metamorphic rocks such as phyllite, schists, calcisilicates, marbles and granite-gneiss. None of the associated rocks show such degree of metamorphism at the contact or elsewhere. So, it is treated as a separate tectonic unit of the region.

Brunnschweiler (1966) assigned a Pre-Mesozoic or older age to the rocks of this belt while Acharyya et al (1986) suggested a Proterozoic age. As evident from the lithocharacters and degree of metamorphism, it is likely to be of continental landmass in nature and may be counterpart of the Shillong-Mikir Massif of the Indian plate. Probably, the complex is part of the Myanmar plate and it overrides the sediments of the accretionary prism resulted from the convergence between the Myanmar and Indian plates. It is probably of Pre-Cambrian age and much older than any of the lithounits occurring in the state. However, its pre-accretionary geological, structural and tectonic history is still little known.

7.3.2 Ophiolite Melange Belt

This tectonic subdivision lies on the eastern part of the state forming a narrow belt of about 180Km in length and 10-15Km in width forming an integral part of the Naga-Manipur-Chin Hills Ophiolite Belt (NAMCHO). Although, it is made up of a
distinct ophiolite complex and a sedimentary sequence, there is no clear geologic criterion to separate them. A number of lithostratigraphic (Formation) names have been proposed and/or coined in the adjoining state of Nagaland to describe various types of lithounits (see Mitra et al, 1986a). However, their lithostratigraphic and chronostratigraphic relationships are still not clearly known since these are all intermingled tectonically.

It is quite likely that the Melange Belt extends far into the adjacent Disangs as evident from the presence of olistostromal bodies within these sediments. Presence of similar exotics in the Chin Hills and Arakan-Yoma Segment within much younger sediments has been discussed. In Manipur the belt is possibly wider than shown in figure 7.6 and may extend upto the Thoubal Thrust on the west.

This unit is overlain by the Metamorphic Belt and underlain by the Disang sediments. Although, not defined clearly everywhere, the contact between the Disangs and this belt is marked by a moderately high angle thrust fault called the Tengnoupal-Narum Thrust as in the case of the Metamorphic Belt. As already discussed in chapter 2, the belt is principally made up of ophiolites of basic/ultrabasic extrusives and intrusives with associated sediments of deep sea origin. Among the sediments limestones, cherts, shales, sandstones, grit and conglomerates composed of basic, ultrabasic, chert clasts are common. Similarity between these sediments and modern ocean deposits suggests a deeper part of the depositional basin probably of the open oceanic crust. The belt further requires detailed exploration to know the geological, structural and tectonic relationship of each of the smaller units that are mingled in the form of melange.

7.3.3 Central Manipur Flysch Belt

It is the largest tectonic unit of the region occupying almost 60-65% of the state's total areal extent. It is bounded by the Tengnoupal-Narum Thrust on the east and Nungba Thrust on the west and includes a number of other large regional lineaments such as the Thoubal Thrust, Churachandpur-Mao Thrust, etc. This unit is
made up of two distinct group of lithounits known as Disang Group and Barail Group. The former is predominantly composed of argillaceous sediments represented by a thick monotonous sequence of dark grey to black shales with intermittent silty and sandy interbeddings occasionally. At places, therefore, they look like rhythmite beddings. The latter (Barails), on the other hand, is characterised by its predominance in arenaceous sediments with frequent argillaceous interlayerings. Both the groups of rocks display typical flysch characters. The shales of Disang Group are characterised by sandy nature while the sands of Barail Group are argillaceous. So, these are collectively nothing but turbidites formed by turbidity currents in a flysch basin (see also Bouma, 1962).

The Disangs usually display features of relatively deep water depositional environment because, shallow water characteristics such as primary structures are absent. Fossils are also relatively rare, and gravity sliding or slumping type of deformation is common. On the other hand, the Barails are characterised by primary structures such as ripples, cross beddings, track and trails, etc. Its relatively abundant carbonaceous matters, specially in the middle horizons, also probably suggest shallow water environment of deposition.

Although, there is difference in relative depth of water in the environment of deposition, the other factors seem to remain more or less the same. Because, greywacke to subgreywacke nature of the sandy horizons, their other physical properties, intertonguing nature of argillaceous matters in the sands and arenaceous matters in shales, all suggest a similar basin characteristics.

7.3.4 Western Manipur Molasse Belt

With a widespread surface of erosion on top of the Barails, the sediments of Surma Valley (Basin) were deposited. Probably the source of sediments of the basin was the uprising Indo-Myanmmar Range. The basin occupies western part of the state and is better developed in Mizoram and Tripura states. The rocks of the basin, on the western part of the state, is in fact, the north-eastern extension of the basin from the
two states. So, only the eastern boundary of this belt is marked by the Nungba Thrust in the state which is partly tectonic and partly lithological (Fig. 7.6).

The sediments of Surma and Tipam Groups are predominantly sands although, considerably subordinate shales are associated specially in the lower horizons. Unlike the Disangs and Barails, the sediments of this basin do not show much argillaceous or arenaceous characters; the sands are less argillaceous while the shales are low in arenaceous matter (see Evans, 1932). So, there is a clear indication of change in the environment of deposition. The sedimentary structures and occasional ferruginous sandy horizons suggest relatively shallow water, probably slightly continental, environment of deposition. It is, therefore, likely that the sediments were deposited in a frontal flexural basin evolved as a result of the uprising IMR. As the accretion of the IMR continues from the convergence between the Indian and Myanmar plates, the sediments have been further compressed thereby forming the frontal folds belt found in the western part of the state as well as in Mizoram and Tripura.

7.3.5 Imphal (Manipur) Basin

It is a small basin lying within the Central Manipur (Disang-Barail) Flysch Belt in the middle part of the state. It is also known as the Imphal/Manipur Valley. Brief topographic and geomorphic descriptions of the valley have already been provided in chapter 1. The basin does not contain any lithounits of old age except the Quaternary deposits of fluvi-o-lacustrine origin. However, it has certain distinct geomorphic and geologic entities and thus treated as a separate tectonic unit. Discussion on the evolution of this basin/valley is made in the next chapter and so, detailed account about the basin can be referred to the next chapter.

7.4 TECTONIC EVOLUTION

It has already been discussed that the collision between the Indian and Myanmar plates represents an island arc type of convergence. An evolutionary model of the Indo-Myanmar Range, therefore, must account for all the structural and
tectonic constraints of the range and its associated features. Various evolutionary models have been proposed by different workers (Mitchell and Mckerrow, 1975; Nandy, 1982; Roy and Kacker, 1982; Acharyya, 1986, 1991). An evolutionary model, depicting all the characteristic features of the accretionary prism is shown in figure 7.7. Different stages of evolution of the IMR have been discussed by the workers mentioned above and so no further model on the evolutionary stages is included here. Instead, a brief discussion on the various stages of evolution is provided based on the lithological characters and depositional environment. Finally, compatibility between the model and that of an accretionary island arc prism is discussed.

Pre-rifting geology of the range is still very little known. On the Indian landmass, it is probably characterised by Gondwana sedimentation while on its Myanmar counterpart by Albian limestone deposits. Rifting and stretching of the crustal layer (lithosphere) possibly initiated sometimes towards the close of Mesozoic Era (Upper Cretaceous). It led to the formation of a basin due to lithospheric stretching. Various mechanism of lithospheric stretching can be obtained from Mckenzie (1978), Wernicke (1985), Coward (1986). A good discussion on various types of basin and their tectonic environment can also be obtained from Allen and Allen (1990), Reading (1982). It is likely that the basin was evolved as a result of passive rifting of the continental margins. However, it is not clearly known whether there were an element of basin associated with strike-slip deformation. The stretching was more or less synchronously associated with the formation of new oceanic crust. It seems that the basin remained sedimentary starved for some time characterised only by the deposition of cherts and calcareous rocks and associated submarine erosion giving rise to the patchy and localised conglomerates at the base of flysch sediments (see also section 2.5).

This stage was followed by the deposition of the huge pile of Disang and Barail flysch sediments. Evans (1932) estimated a thickness of about 4500m for the Barails. Assuming a similar thickness of the Disangs, a total column of about 9000 - 10,000m thick sedimentary succession has been deposited. It is questionable whether
Fig. 7.7. Schematic tectonic evolutionary model of the Indo-Myanmar Range in terms of an accretionary prism of an island arc type convergence. Profile runs WNW-ESE passing through the Imphal Valley.
a relatively small size basin can accumulate such a huge thickness of sediments. There are two possible explanations to this question.

(i) The excessive thickness of the sedimentary sequence is due to thrusting of one unit over the other. So, it appears to be much thicker than what is in reality.

(ii) The huge thickness of the sequence is due to sediment load induced flexure coupled with continuous but pulsatic stretching as briefly outlined in chapter 2 (section 2.5). Such type of basin development and sedimentation have been widely studied e.g. Watts and Ryan (1976), Steckler and Watts (1978).

With the gradual filling of the basin, towards the close of Oligocene, convergence of the two plates started gradually accreting the Indo-Myanmar Range. In front of the rising range, a frontal basin was subsequently formed as a flexural basin which is termed as the Surma Forediep (Basin) by many workers. In this basin the molasse sediments were deposited giving rise to the Surma and Tipam Groups of rocks.

During the accretion process, the oceanic crust was sliced off along with its associated deep sea sediments by the overriding Myanmar plate and emplaced onto the flysch sediments. As the convergence continues, not only folding and intense deformation of the turbidites occur, but a number of thrust faults also spread out giving rise to the prominent tectonic lineaments. All these thrust faults join the sole major thrust probably running on top of the Indian plate. These imbricate system of thrust faults possibly control the geometry of the frontal folds of the molasse basin also.

Now comparing the model with that of an accretionary prism of an island arc convergent boundary, the following points can be discussed. For an island arc model, the reader may see Allen and Allen (1990, p 242).
(i) The Surma Basin (Western Manipur Molasse Belt) probably including some parts of Bengal Basin is the frontal flexural basin similar to the trench fills of the island arc.

(ii) Indo-Myanmar Range and western part of the Chindwin Basin (Western Trough) forms the accretionary prism. Chindwin Basin is similar to the residual forearc basin (RFB).

(iii) Myanmar Volcanic Arc corresponds to the active volcanic ridge of the island arcs on the continental side. The above three units form what is known as the forearc.

(iv) On the eastern side of the volcanic arc, the Irrawaddy Basin (Eastern Trough) represents the backarc basin. Further on the eastern side of the basin, the high angle fractures correspond to the Shan Scarps beyond which is the Highlands of Eastern Myanmar. Probably, Irrawaddy Basin is characterised by polyphase tectonic phenomena, so, a clear backarc basin features are absent being intermingled with other deformation structures.

7.5 ESTIMATION OF TECTONIC SHORTENING AND STRETCHING

Orogenic belts deformed by folding and thrust faulting have been formed due to plate movements specially in the boundary areas. Nowadays, one of the most widely used methods of estimating tectonic shortening as well as stretching (or extension) of fold-thrust belt is the principle of balanced cross-section. The concept was first introduced by Dahlstrom (1969) while he was trying to restore the geological profiles of the Eastern Canadian Mountain Fronts. Although, there are certain demerits in this method, because of its geometric simplicity, it is used almost by everyone who studies any fold-thrust mountain belt. The rapid use and application of this method is also due to the sudden increase in the availability of seismic profiles of these belts as well as other parts of the crust through oil exploration. Because, from these profiles a number of subsurface data, which were previously inaccessible, can be obtained which in turn forced to reinterpret the geometric and other kinematic
modelling of these mountain belts. Hundreds of papers, specially in the western countries have been published since the introduction of this concept in the geological literature. More simplified versions about the balanced section have also been discussed in Hossack (1979), Ramsay and Huber (1987), Marshak and Mitra (1988), etc.

In order to restore the geological profiles to its original disposition and to calculate either the tectonic shortening or extension, some assumptions and necessary conditions are adopted. Details of these conditions can be had from the literature cited above. But some of the important conditions are as follows.

1. The first and most important assumption in the study and use of balanced cross-section is the principle of plane strain condition i.e. area is conserved.

2. The geological profile(s) that is to be restored to its original unstrained state should be drawn normal to the tectonic trend for shortening, calculation and parallel to the regional strike for estimating extension (or stretching). In other words, the profile(s) should be parallel and perpendicular to the tectonic transport direction respectively for estimating crustal shortening and extension.

3. The geological profile(s) may be better supported and made near complete by incorporating a lot of subsurface data from seismic profiles and/or gravity anomaly studies. Otherwise, balancing may be practically difficult and estimation of shortening could be highly erroneous.

4. The geological profiles may be used either for area balancing or line length balancing depending on the information available in the profiles.

5. The type of geological profile - whether it could be deformed-state section, admissible section, restorable section, viable section, or balanced section may be examined depending on the profiles drawn (for details see Elliot, 1983).
Geological profiles may require modification when new data become available. Generally profiles that can be balanced are correct whereas section that does not balance is probably wrong. So, profiles that lack sufficient surface and subsurface data should not be used for balancing.

7.5.1 Crustal Shortening

Using the method of balanced section, crustal shortening of Manipur Hills can also be estimated. The concept of balanced section where plane strain condition of area conservation is very simply exemplified by the figure 7.8. Figure 7.8A shows the excess crustal thickening beneath a mountain due to tectonic contraction. Thickening of the crust is proportional to the shortening of crust in the mountain belt (Coward, 1992; see also Gwinn, 1970) and so, under plane strain the relationship are as follows:

\[ \text{Crustal thickening, } E = \frac{T}{H}, \text{ and Crustal shortening, } S = \frac{H}{T}, \]

where \( T \) is the excess crustal thickening due to tectonic contraction, 
\( H \) is the crustal thickness outside the mountain belt, 
\( T+H \) is the thickness of the crust beneath the mountain belt 
\( t \) is the difference in elevation between the mountain belt and its foreland.

In order to compute the original width/length of the mountain belt, the following relation is used, i.e.

\[ \text{Original width} = \sum \text{Final width} \times \delta \frac{1}{s} \quad (7.1) \]

or,

\[ \text{Original width} = \text{Area beneath the curve and datum} \times \delta \frac{1}{s} \]

That is, find the area beneath the curve defined by \( 1/s \) against distance across the belt.

Figure 7.8B is another method of computing tectonic shortening where conservation of area under plane strain condition is assumed (see Hossack, 1979). Another important feature of this method is the assumption of existing a horizon of non-deformation - decollement which can be ascertained from seismic profiles or from drill-hole data. From the figure it is clearly evident that under plane strain,
Fig. 7.8. Simplified schematic diagrams of area balancing based on the principle of balanced cross section under plane strain. A. Crustal thickening as a function of shortening (simplified from Coward, 1992). B. Crustal thickening above a decollement (simplified from Hossack, 1979).

A.

Mountain Range

Foreland

Original width = $A \times \frac{1}{s}$, s-shortening

B.

Decollement/Thrust

$A$
AHFG = BCEH = A = the excess area.

So, initial depth to decollement, \( t_o = \frac{A}{S} = \frac{A}{L_o - L_f} \) \hspace{1cm} (7.2)

and, shortening, \( S = \frac{A}{t_o} \) \hspace{1cm} (7.3)

where \( L_o \) is the initial length, \( L_f \) is the final length.

Hence, either by knowing the initial depth to decollement \((t_o)\) or the shortening \((S)\), we can find out the value of the other parameter, and the rocks within the area, DEFG may be deformed by any style of folding and/or thrusting (Hossack, op.cit.).

To compute the original or initial length \((L_o)\), the following relations can be used

1) \( L_o = \frac{A}{t_o} + L_f \) (Elliot, 1977; see also Gwinn, 1970) \hspace{1cm} (7.4)

2) \( L_o = \frac{L_f t_1}{t_o} \) (Keifer and Dennison, 1972) \hspace{1cm} (7.5)

a. Computing Tectonic Shortening Across the Manipur Hills

In order to estimate the crustal shortening across the hills of Manipur, the method outlined by Coward is adopted. Because, the method given by Hossack is practically difficult to employ since, for the region depth to decollement is not known. Only the area balancing method is used, sinuous bed or line length balancing method is also not applied. Because, the geological profile (Fig. 2.2) across the state is drawn purely based on the field data collected by the author and a number of places along the profile line is inaccessible and so, dip-strike data are projected using the values obtained along the strike continuity. No subsurface data e.g. seismic profiles are available to corroborate the information given in the profile. Moreover, there are no specific marker horizons that are well known as well as mapped which can be used for line length balancing. Lack of specific marker horizon is also probably due to the monotonous nature of the lithounits (e.g. Disangs, Barails) present in the region. The procedures for computing the crustal shortening using the method mentioned above can be summarised as follows.
1. A topographic profile across the hills of Manipur state is drawn normal to the regional tectonic trend (WNW-ESE, the profile runs nearly parallel to 24°45'N and 24°30'N line). Survey of India (SOI) degree-sheets on 1:250,000 scale are used in order to plot the curve quite accurately, and the vertical scale is exaggerated 5 times to make easier in calculating the area (Fig. 7.9).

2. Datum line corresponding to the foreland elevation of the IMR is defined which is as low as 30-40m above the sea-level in Jiribam area. So, sea-level (0m) is taken as the general elevation of foreland for practical simplicity.

3. The area beneath the curve and the datum is then calculated using simple graphical technique. To simplify this, the curve is plotted on a centimetre graph sheet. The area between the Makru river (X) on the west and the topographic break between the Tamu/Kabaw Valley and Eastern Manipur Hills (Y) is then evaluated. It is found to be 118.46Km². Electronic or manual planimeters can also be used for area calculation.

4. The final length or width between the two reference points (X and Y) within which the area was estimated is evaluated. This length is found to be 123.75 Km along the profile line.

Now after evaluating the area beneath the curve, the original width or length of the area lying between X-Y line in the profile can be computed. In order to do so, however, we need to have the values of H and T. The value of H, the crustal thickness outside the mountain belt has been taken as 20Km as per the diagram provided by Mukhopadhyay and Dasgupta (1988), Mukhopadhyay and Krishna (1992). In the information given in their diagrams the crust below the Bengal Basin has an approximate thickness of 20Km or less. The foreland region of the IMR, the western part of Manipur state and the Cachar Plains are very similar to the Bengal Basin. So, the value of H given above is taken into consideration. This value of 20Km may be quite reasonable because in Andaman region also thickness of the crust is reported to be 20Km by Kumar (1981, 1990). A thicker crust is indicated by Kayal (1995). However, the average crustal thickness of mainland India is about 36 to 40Km
Fig. 7.9. Topographic profile across the hills of Manipur in a WNW-ESE direction nearly parallel to 24°45'N and 24°30'N latitude line and to the tectonic transport direction. X-Y is the region for which shortening is computed.
(Qureshy, 1970). Hence, in the foreland part of the IMR, the crust must be thinner since sialic layer is likely to be very thin or absent in this sector. The value of \((T+H)\), thickness of the crust in the IMR is taken as 53Km from Sitaram and Barua (personal communication). They have indicated a value of 52.4Km for the North Cachar Hills and 53.9Km for the IMR with a possible error of \(\pm 10\%\). From all the above deliberations, finally the values can be given as follows.

The crustal thickness outside the mountain belt, \(H = 20\text{Km}\).

The crustal thickness in the mountain belt, \(T+H = 53\text{Km}\).

\[\therefore\text{the excess crustal thickness, } T = 33\text{Km}.\]

Now employing equation (7.1), the value of original width can be computed as

\[
\text{Original width, } L_o = \frac{\text{Area beneath the curve and datum (A)}}{H} \times T
\]

Substituting the values of \(A\), \(H\) and \(T\) in the above equation, we get

\[
L_o = 118.46 \times \frac{33}{20} = 195.46\text{Km}.
\]

\[\therefore\text{shortening, } S = L_o - L_f = 195.46 \text{Km} - 123.75 = 71.71\text{Km}.\]

Expressing the shortening (S) in terms of percentage using the principle of measuring engineering strain, \(e\) we have

\[
e = \frac{L_f - L_o}{L_o} \times 100
\]

\[
= \frac{123.75 - 195.46}{195.46} \times 100 = -36.69\% \approx -37\%
\]

\[\therefore\text{shortening, } S = 37\%.\]

The value obtained above i.e. 37\% may be substantially underestimated. Because, the profile across the state passes through the Imphal Valley (A-B) where the area loss between the A-B line and the topographic curve (see Fig. 7.9) is found to be 37.55Km\(^2\). This accounts to 49\% of area loss in this sector. On the other hand, the area loss from the usual topographic unevenness or ruggedness in the sector (A-C) is found to be only 33\%. So, the excess loss of 16\% cannot be neglected, for it can change the shortening figure of the region considerably. Now subtracting 33\% of the area, 37.55Km\(^2\) and adding the remaining value, which is worked out to be 25.16Km\(^2\), to the area, 118.46Km\(^2\), shortening of the state can be computed again as follows.
Original width, \( L_0 = 118.46 + 25.16 \times \frac{33}{20} = 236.97 \text{Km.} \)

\[
\therefore \text{shortening,} \ S = 236.97 - 123.75 = 113.22 \text{Km.}
\]

i.e. shortening, \( S \) expressing in percentage = 47.78% \( \approx \) 48%.

From the above, the area between the Makru river on the west and the margin of Eastern Hills of Manipur (X-Y) might have been as wide as 237Km before shortening.

The assumption of plane strain in balanced cross-section is criticised by a number of workers (e.g. Geiser, 1978). The argument is that there is considerable reduction in volume thereby leading to the reduction in area of the profiles. The factors that contribute in volume loss or area reduction are tectonic compaction, pressure solution, elongation along the tectonic strike and of course, erosion. In this context Hossack (1979) argues that, the loss in area is evident and sometimes by no means small. And therefore, the value of shortening estimated by employing any method will be the minimum. So, the values can be used always as a first approximation for the purpose of tectonic studies. True balancing and restoration of the cross-sections to their original disposition may be carried out as and when more data on volume reduction are available. In case of the present study also, 48% shortening may still be slightly underestimated. Because, parameters causing area/volume reduction though, are not known, rate of erosion in the region seems to be quite high. High rate of both mechanical and chemical weathering and erosion of the rocks is due to intense fracturing and heavy precipitation in the region. Hence, of the 33% of topographic ruggedness expressed above, 20% may account for erosion only. Thus, 48% of shortening of the state calculated above may be used as a first approximation. Future refinement can be incorporated with increase in availability of both subsurface as well as surface geological data.

b. **Computing Shortening From Minor Structures**

Minor compression structures such as folds, reverse/thrust faults can be employed to compute the shortening in a localised extent. This principle is exemplified by the figure 7.10, and can be referred to any standard structural geology
Fig. 7.10. Diagrams showing how small structures such as folds and faults can be used to calculate shortening. A. Small fold for computing shortening (negative extension). B. Small reverse/thrust faults for computing shortening.

A.

\[ e = \frac{L_f - L_o}{L_o} \times 100 \]

Extension

B.

\[ s = D \cos \phi \]

Shortening
Fig. 7.11 An eroded small anticline in Thanga hills and reconstructed using dip-strike data for measuring shortening. A. Photograph of the eroded anticline, taken facing south. B. Reconstructed fold profile.
text book e.g. Park (1989), Twiss and Moores (1992), Davies (1984), etc. In figure 7.10A, a folded layer is considered for computing shortening in terms of engineering strain (ve extension, \( e \)) as mentioned in equation (7.6). On the other hand, figure 7.10B employs a small reverse or thrust fault for computing shortening. From the figure it is seen that shortening can be expressed as follows

\[
\text{shortening, } S = D \cos \phi
\]  

(7.7)

where \( D \) is the fault displacement, \( \phi \) the fault angle.

However, expressing this shortening in terms of percentage is impractical since the original length is unknown. So, it basically provides only the approximate value of shortening of that fault and may not be comparable with the regional shortening.

Figure 7.11A is an eroded antiform found in the Thanga hills of Loktak Lake where 7.11B is the sketch of a sandstone band 2.5 to 3m thick. The fold is completely eroded but limbs have been preserved in the hill slopes. Profile of the fold is constructed based on dip-strike data measured in the field (Eastern limb, N-S, 60°/E and Western limb, 025°, 76°/W). Angular, but not sharp hinge is assumed based on the general fold geometry as discussed in chapter 6. This fold has the following data as on the photograph, \( L_f = 9.6 \text{cm}, \ L_o = 18.3 \text{cm} \).

\[
:e = \frac{L_f - L_o}{L_o} = \frac{(9.6-18.3)}{18.3} = -47.54 \approx -47\%.
\]

In the same manner, figure 7.12 represents a minor thrust from Ngariyan road section where 7.12A is the photograph and 7.12B is the corresponding sketch showing the displacement of the marker bed. Due to difference in fault attitudes, the faults provides variable magnitudes of shortening. Using the relation given in equation (7.7), the following contractions have been estimated.

For \( T_1, 003°, 81°/W, \phi = 81°/W, D = 94 \text{cm} \). \( \therefore S = 14.7 \text{cm} \).

Expressing the shortening in terms of \( D \) as percentage, \( S = 16\% \).

For \( T_2, 004°, 50°/W, \phi = 50°/W, D = 44 \text{cm} \). \( S = 28.3 \text{cm} \). i.e. \( S = 64\% \).

The values of shortening estimated using the minor compression structures mentioned above are found to be widely variable although, all the analysed data are
Fig. 7.2. Small/minor thrust from Ngariyan road section (Ngariyan hills) for estimating shortening. A. Photograph showing the minor thrust. B. Sketch of the thrust above.
not included here. But, small folds generally show relatively less variation than the
minor thrusts. For instance, figure 7.13A is also another small fold found near the
Ilang river (along NH 53) providing a shortening of 44%. The variation in shortening
given by these folds usually ranges between 40 to 50% with very few exception. On
the other hand, minor thrusts give highly variable shortening ranging between 16 to
85%. Figure 7.13B is such another minor fault developed in a competent layer as a
pre-buckle thrust (see also Price and Cosgrove, 1990). This fault ($\phi = 61^\circ$/$180^\circ$, $D$
= 101cm) gives a shortening value of 48% but the fault is highly rotated and oriented
almost E-W. The high variation of shortening values of these small thrust faults may
be due to their post deformational rotations leading to wide difference in their
attitudes. However, in a generalised manner, the shortening given by the minor
structures, specially the folds, are quite compatible with that of the region (48%)
computed above.

7.5.2 Crustal Stretching (Extension)

When the principle of balanced section was introduced in the studies of the
thrust-fold belts by Dahlstrom (1969), there was a revolution in the entire gamut of
tectonic or orogenic belts analysis. Similarly, since the introduction of the principle of
crustal stretching in basin formation by McKenzie (1978), there was also a sea-change
in the research and study of basin evolution and analysis. The simple principle is that
when the lithosphere of the earth is stretched or extended by a pure shear mechanism
with a pulling force, the upper elastic layer is deformed into normal faults by brittle
failure. With the pull, the faulted blocks not only rotate but thinning of the crust also
takes place. This in turn causes subsidence of the stretched crust first by the faulting
as an instantaneous and permanent process and then, more importantly followed by
thermal subsidence. The tilted blocks are evidently observed from seismic profiles of
the crust. This concept practically has revolutionised again the earlier ideas of basin
evolution, e.g. geosyncline theory. Similar basin formation models, but of different
deformation mechanisms - simple shear and heterogeneous simple shear stretching
have also been respectively discussed by Wernicke (1985) and Coward (1986). A

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Fig 7.13. A. Photograph showing a small fold (antiform) near Irang river along NH 53, providing a shortening of 45%. B. Photograph showing a minor fault, like pre-buckle thrust, developed in relatively competent sand layer, along NH 53 near Tupul.
detailed account on the mechanics of basin formation can also be had from Allen and Allen (1990).

The tilted fault blocks can be employed for computing crustal stretching as observed in the seismic profiles. There are a number of methods and techniques based on which crustal stretching can be estimated e.g. subsidence history, crustal thickness changes, tilted/rotated faults, extension from strain analysis, offset patterns on transform faults/structures (a comprehensive account of this can be obtained from Allen and Allen, op. cit.). Among the methods mentioned above, rotated as well as non-rotated normal faults are most commonly used and the principle is illustrated in figure 7.14. Detailed discussion on crustal stretching, subsidence, footwall uplift as well as other aspects associated with normal faulting can also be referred to Wernicke and Burchfiel (1982), Jackson and McKenzie (1983), Barr (1987), etc. Figure 7.14A depicts a tilted or rotated normal fault where $\phi$ is the fault angle and $\theta$, the bedding angle. AB is the final length ($L_f$) and BC the original length ($L_o$). Then according to the formula of stretching, we get

\[
\text{Stretch, } \beta = \frac{AB}{BC} = \frac{L_f}{L_o} \quad (7.8)
\]

From the $\Delta$, ABD we have, $AB = BD/sin \phi$ and $BC = BD/sin (\theta + \phi)$. Substituting these values in the above equation,

\[
\beta = \frac{BD / sin \phi}{BD / sin (\theta + \phi)} = \frac{sin(\theta + \phi)}{sin \phi} \quad (7.9)
\]

So, by knowing the tilted fault and bedding angles, crustal stretching can be estimated. From the relationship it is clearly evident that, as $\phi \to 0^\circ$ and $\theta \to 90^\circ$, the extension of the crust may be infinitely large. That is, in other words, since normal faults develop at relatively high angle (say $60^\circ$) in the beginning, they will become gentle dip with more and more extension or stretching of the rocks/crust.

Similarly, figure 7.14B is a normal fault that is practically irrotational. So, the bedding remains more or less horizontal after faulting. From this, extension can be computed simply as

\[
\text{Extension, } e = D \cos \phi \quad (7.10)
\]
Fig. 7.14. Diagrams showing how normal faults can be employed to estimate crustal stretching and extension. A. Tilted and rotated faults and beddings. B. Nonrotated fault and horizontal beddings.

Stretch, $\beta = \frac{\sin(\theta + \phi)}{\sin \phi}$

Extension, $e = D \cos \phi$
where D is the displacement along the fault, $\phi$, the fault angle.

2. Computing Stretching

Crustal stretching or extension of the Imphal Valley and its adjoining region is calculated using the methods mentioned above. But in the processes of calculation, a number of problems have been encountered. For instance, in the tilted/rotated fault block model, it is generally assumed that the beds were initially horizontal. In the valley, however, the rocks have already been deformed as well as tilted to a very high angle. Therefore, the dip angles cannot be directly used. Although, they are supposed to appear horizontal in the fault profiles, they often don’t so, because quite frequently extension took place along the pre-existing shear fractures. So, the stretching direction is not ideally parallel to the regional extension direction, NNE-SSW. Other problems that are commonly faced is the unavailability of adequate subsurface data. The small hillocks within the valley though, seem to be remnants of horsts or half horsts (uplifted footwall blocks), nothing can be precisely confirmed except inference based on field study. Lack of marker horizon is another problem in order to calculate the stretching factor ($\beta$). Because, without known beds, the final length cannot be accurately obtained and in addition to it, due to high weathering and erosion rates, the fault surfaces are often eroded thereby making difficult to measure the fault attitude. In spite of these practical difficulties, the normal faults or faults that show extension features have been employed to compute the crustal stretching, rather a possible range of crustal stretching, of the rocks of the Imphal Valley.

Figure 7.15 is a fault that shows considerable extension. Probably it was an antithetic shear fracture along which dip-slip has occurred during reactivation. Figure 7.15A represents the fault (331°, 30°NE) characterised by silification. This fault has two slickenide striations - one plunging 15°/SE while the other 40°/NE on the fault trending 337°. The low plunging slickenide may correspond to the earlier strike-slip component while the 40°/NE plunging slickenide to the dip-slip of latter reactivation associated with extension (Fig. 7.15B). The difference in the fault angles of 30°/NE and 40°/NE suggests that, possibly they have extended and rotated, at least
Fig. 7.15. Photographs showing normal fault and associated features in Pechi hills near Yairipok.
A. Reactivated pre-existing fault. B. Slickensides showing two senses of slipping, pencil with blue tip indicates 15°/137° plunge while the other indicates 40°/068° plunge.
for a couple of times. In this locality the bedding has dip-strike of 274°, 30°/N and oriented nearly normal to the regional extension. Since the faults are not oriented parallel to the regional compression direction (WNW-ESE) computation of stretching from the data is practically difficult. Hence, in order to simplify it, the faults and the bedding are plotted on a stereographic (equal area) net. Then the angular relationship between the faults and bedding along the slip direction (N68°E) and along the intersection directions (which are also roughly parallel to the regional extension direction) are evaluated (Fig. 7.16A). The angular and calculated stretch factor of the two faults are presented in Table 7.1.

<table>
<thead>
<tr>
<th>Fault</th>
<th>Fault angle, $\phi$</th>
<th>Stretching factor ($\beta$)</th>
<th>40°/N67°E (F1)</th>
<th>Stretching factor ($\beta$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Along dip-slip, 40°N68°E</td>
<td>30°</td>
<td>1.46</td>
<td>48°</td>
<td>1.26</td>
</tr>
<tr>
<td>Bedding angle, $\theta$</td>
<td>17°</td>
<td></td>
<td>14°</td>
<td></td>
</tr>
<tr>
<td>Along intersection direction</td>
<td>28°</td>
<td>1.77</td>
<td>30°</td>
<td>1.73</td>
</tr>
<tr>
<td>Bedding angle, $\theta$</td>
<td>28°</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

From the table, it is clearly seen that the stretching factor, $\beta$ varies from a minimum of 1.26 to a maximum of 1.77. The figures, obtained above, may not be incorrect because, the rocks in the valley might have undergone a couple of (possibly more) stretching phases. The maximum figure may be a cumulative value of the stretching. When the faults are rotated to a low angle, high values of stretching factor have been indicated by a number of researchers also e.g. Proffett (1977), McKenzie (1978), Pichon and Sibuet (1981).

Figure 7.16B is a geological profile map of a small area in Thoubal district (between Kaina and Angha hills). Here a marker sandstone bed is found displaced, the intervening land being occupied by agricultural field. The slope in the N-S direction is roughly parallel to the tilted bedding angle which is found to be 23°/NNE. The fault surface is not well preserved due to erosion, but the slope angle
Fig. 7.16. A. Stereographic plot of the Pechi faults shown in figure 7.15. F₁ has 337°, 40°/067° and F₂ has 331°, 31°/061° attitudes (see text for more details). B. Geological profile along Kaina-Angtha hills showing a possible stretching factor (β) of about 1.46 based on field observation.
Fig. 7.17. A. Photograph showing a small normal fault indicating extension along the Thongjaorok stream. B. Sketch of the fault above. The fault trends 104°, dips 68°/S and the bedding has tilted by 21°/N.
corresponding to the fault surface which is found to be $36^\circ$/SSW is assumed to be the fault angle. Then by employing the formula given in equation (7.9), we have

$$\beta = \frac{\sin(23^\circ + 36^\circ)}{\sin 36^\circ} = 1.46.$$ 

Figure 7.17A is a photograph of a small normal fault found along the Thongjaorok stream while 7.17B is the sketch of the fault. The attitude of the fault is $104^\circ$, $68^\circ$/S, while the bedding as appeared in the fault profile is $21^\circ$/N. From the above data, using the equation (7.9), we can get

$$\beta = \frac{\sin(21^\circ + 68^\circ)}{\sin 68^\circ} = 1.07.$$ 

This value is comparatively less than those estimated for rocks in the central part of valley as computed above. The less value may be simply an indication of low stretching of the competent Barails in the western periphery of the valley. This value may, therefore, account only for the tectonic elongation along the orogenic strike which may range 5-15% (Hossack, 1979). So, 7% extension ($e = \beta - 1$) mentioned above may not be indicative of the stretching of the valley in the central part or otherwise, the fault may become relatively gentle in deeper part of the crust thereby providing larger extension at depth because such possibilities have been indicated by McKenzie (1978). This latter idea can, however, be verified only by subsurface seismic data.

Whatever may be the variation in the values of stretching computed above, a generalised outcome is that rocks of the Imphal Valley might have undergone more than a single phase of extension. There are evidences of very recent stretching or extension of the crust of the valley even in the last century (see next chapter, 8). Therefore, the stretching factor estimated above may not be incorrect and probably the value of may range from 1.25 to 1.77. And in absence of any precise subsurface data, these values may be used as a first approximation.