Chapter 3
The Young and Old Fold Belts

3.1 Introduction

With the basic concept of plate tectonics and mountain building processes discussed in Chap. 1, we now attempt to understand the evolution of the fold belts or mobile belts or orogenic belts in light of the plate tectonics theory. We believe that orogenic belts, being fold-thrust belts, formed at convergent margins where the adjacent plates had moved toward each other and collided. These margins are called subduction zones and are classified as either oceanic or sub-continental, depending upon whether the crust of the overriding plate above the subduction zone is oceanic or continental. Oceanic subduction zones are marked by a trench and by an arcuate chain of volcanoes on the overriding plate, called island arcs or ocean island arcs. Subcontinental convergent margins are also marked by an oceanic trench and by an arcuate chain of volcanoes that are built on the continental crust. These margins are called continental arcs or Andean-type continental margins. The Andes is the simple orogenic belt that formed when oceanic lithosphere (Nazca plate) subducted eastward under a continental margin of South America. This mountain belt is a 900 km long along the western margin of South America. The prominent features are the active Peru-Chile trench and subduction zone, the active volcanic arc on land, and an active fold-thrust-belt along the eastern margin of the mountain chain. The Nazca plate is subducting beneath South America, along much of the length of the Andes. The Andes have a long Pre-Mesozoic to Recent history (Dewey and Bird, 1970).

Continental mountain ranges can also develop as the result of a collision between an island arc and a continent, but such mountain belts should be smaller than those formed by continent-continent collision. Arc-continent collision zones are relatively rare, as they usually represent an intermediate step in the closure of an ocean and hence are short lived.

Collision fold belts typically develop when subduction to a continental margin brings a continent that makes up an integral part of the subducting oceanic lithosphere. The result is the collision of a continent with another continent. By studying the young collision orogenic belts, we can discover the relationship between orogenic structures and related plate tectonic activity. The younger orogenic belt preserves most of the sedimentation and deformation history.

R.S. Sharma, *Cratons and Fold Belts of India*, Lecture Notes in Earth Sciences 127,
DOI 10.1007/978-3-642-01459-8_3,

3.2 Himalaya as an Example of the Young Fold Belt

We now look at the characteristics of an orogenic belt that formed by continent-continent collision and the Himalayan mountain range is the best example to understand the events of the plate tectonics in the development of the youngest collision fold belt on the earth. In the evolution of this mountain belt, like other collision fold belts, the geological record of collision is always preceded by Andean-type orogenesis. The Himalayan orogenic belt forms a part of the Alpine-Himalayan mountain chain that stretches from Spain in the west to Indonesia in the east. The Himalaya is among the youngest mountains of the world formed in the framework of plate tectonics theory and a classical example of continent-continent collision. The Himalaya offers an unusually good opportunity for understanding the complex interplay of thermal and mechanical processes that lead to the distribution of strain pattern and magmatic and metamorphic rocks that characterize the different parts of the fold belt. It is here that one can recognize the colliding plates, subduction zone (s), Arc etc. The evolution of Himalaya would thus help us in organizing our observations and developing our ability to apply plate tectonic in the evolution of the Proterozoic fold belts of India described in equal details in this book. Before we take up the evolution of the Himalayan mountain belt we should know the geological setting of this young orogenic belt of India.

3.2.1 Geological Setting

The Himalayan orogenic belt in Indian sector is 250–300 km wide and about 3000 km long from Nagaland State of India in the east to Kashmir in the western India and beyond up to central Afghanistan. The eastern boundary of the Himalaya is taken as the major bend or syntaxis, the Namche-Barva syntaxis, although doubted by some geologists. A similar syntaxial bend is seen in the western part of the Himalaya and is known as the Kashmir syntaxis or Nanga Parbat syntaxis. The Himalayan belt comprises a series of lithologic and tectonic units that run parallel to the mountain belt, maintaining a constant character for long distances (Fig. 3.1). This figure is a compilation after Auden (1937), Heim and Gansser (1939), GSI (1962), Gansser (1964) and Valdiya (1977).

The main geological features of the Himalaya (Fig. 3.1) from north to south are:

1. Tibetan or Asian Block
2. Trans-Himalayan Batholith (Early Cretaceous to Early Oligocene)
3. Indus-Tsangpo Suture Zone (Early to Middle Cretaceous)
4. The Tethys (Tibetan) Sedimentary Zone (Late Precambrian to Early Eocene)

   ——-South Tethyan Detachment Fault, STDS (Trans Himadri Fault of Valdiya, 1989)—-

5. Higher Himalayan Crystalline (HHC) (Late Precambrian with Old and Young Granites incl. Miocene)

   ——————————-Main Central Thrust (MCT)—————-

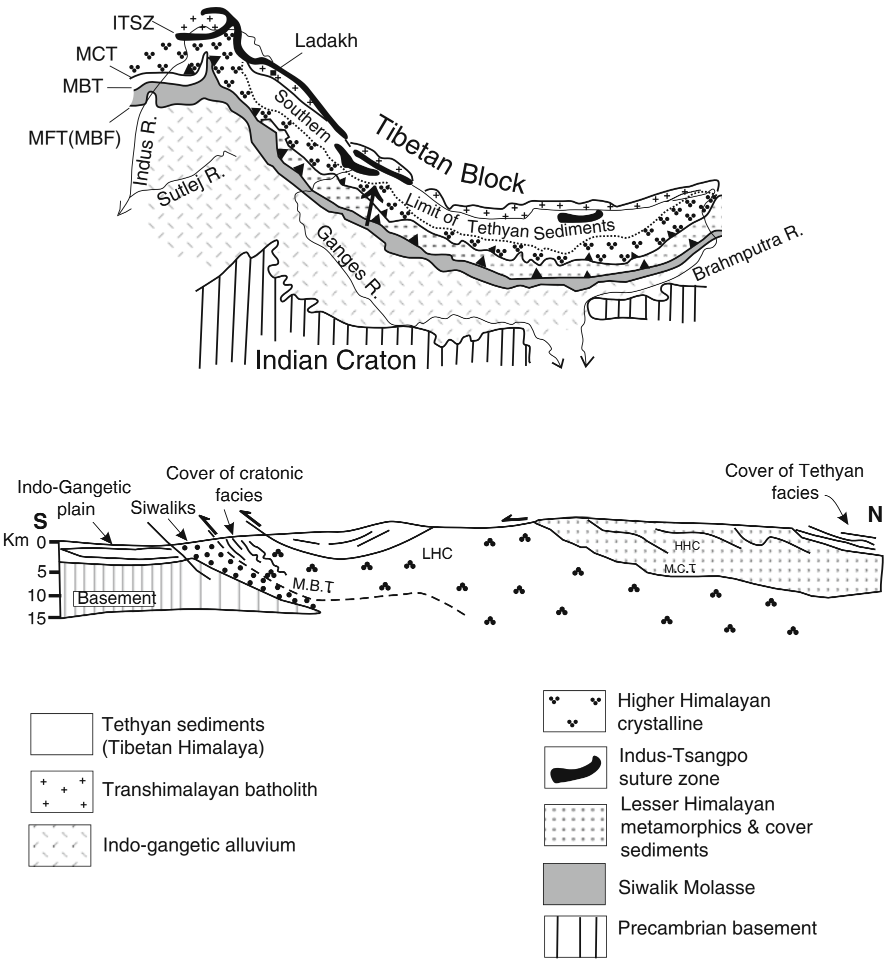


Fig. 3.1 Generalized geological map of the Himalaya showing the principal geological divisions (litho-tectonic units), based on the compilation map of GSI (1962), Gansser (1964), and Valdiya (1977). Bottom figure is a crustal section showing the relationship of different shear zones. Abbreviations: HHC = Higher Himalayan Crystallines, ITSZ = Indus-Tsangpo shear zone, MBT = Main Boundary Thrust, MCT = Main Central Thrust, MFT (MBF) = Main Frontal thrust (Main Boundary fault)

6. Lesser Himalayan Crystalline (LHC) and Cover Sediments (Proterozoic to Palaeozoic/Caledonian Granitoids)

———————————Main Boundary Thrust (MBT)———————————-

7. Subhimalaya or Outer Himalaya (Cenozoic Molasse: Siwaliks)

————————- Himalayan Frontal Fault ——————————

8. Indian Plate (northern margin covered with alluvium)

The major units of Himalaya occur mainly in the Indian sector where the Himalaya is divided as Lesser or Lower Himalaya (elevation of 1500–3000 m) and Higher Himalaya with elevations from more than 3000 to 8000 m. The Lesser Himalayan rocks (LHC) consist of both Precambrian metamorphic and cover sediments (weak to unmetamorphosed) and are thrust over the sub-Himalayan rocks along Main Boundary Thrust (MBT), whereas the Higher Himalayan Crystallines (HHC) have been thrust over the LHC rocks along the Main Central Thrust (MCT) which is a zone of ductile deformation of several kilometers thick (bottom sketch of Fig. 3.1). The metasediments at higher levels of the HHC unit are intruded by granites of Miocene age, which originated by melting of the pelitic schists in upper crustal conditions when the HHC unit was exhumed along the south-vergent, north-dipping Main Central Thrust (MCT). The southward thrusting of the Himalayan rocks commenced mostly in Middle and Late Tertiary and some movements continue even today (see Auden, 1937).

The Himalaya is also divided into three geographic divisions from west to east as Western, Central and Eastern Himalaya. Western Himalaya encompasses mountain ranges (Ladakh, Zanskar, Dauladhar, Pir Panjal etc.) of the State of Jammu and Kashmir, Himachal Pradesh, Uttarakhand (Kumaun and Garhwal Hills) and the adjoining Karakoram, Kohistan, Kailas, Mansarovar ranges. Central Himalaya essentially covers Nepal. The mountain ranges of Sikkim, Darjeeling, Bhutan and Arunachal Pradesh are included in the Eastern Himalaya.

The Himalaya is bordered on the south by the Indian shield covered under the molasse and alluvium of Indus and Ganges river system. Prior to the occurrence in the Himalaya, all of the HHC and the Lesser Himalayan crystallines were once a part of the northern Indian cratonic rocks, mostly of Rajasthan-Bundelkhand craton, Chhotanagpur Gneiss Complex and the Megahalaya craton. These rocks were a variety of Neoarchaean to Proterozoic gneiss, granite and older metasediments which were involved, along with their sedimentary cover in the Himalayan orogeny and were variously deformed and metamorphosed. The sedimentary successions of Eocambrian, Paleozoic and Mesozoic periods were extensively developed in various parts of the Himalaya, notably in Kashmir, Chamba, Zanskar-Spiti and Kinnaur-Kumaun areas of the Western Himalaya. The Lesser Himalaya in Kumaun has also several Proterozoic sedimentary basins, namely the Krol belt, Shali belt (Deoban-Garhwal) and Chamoli-Tejam belt. Most of these sediments were deposited in intracratonic and pericratonic basins (corresponding to the Purana platform basins of the Indian shield) when India was a member of the Eastern Gondwana located in southern hemisphere. The present configuration of these basins is due to the Tertiary Himalayan orogeny and subsequent erosion along various structural levels (Bhargava and Bassi, 1999).

The Indus-Tsangpo (Zangpo) Suture Zone (ITSZ) is considered to mark the southern junction of the Indian and Asian plates in the Himalaya (Gansser, 1964) along which the Neo-Tethys oceanic lithosphere was subducted as early as 140 Ma ago (Searle et al., 1987; Ahmad et al., 2008). The suture is a few kilometers narrow zone chiefly of ophiolite mélange that dips steeply south. The ophiolites are not continuous and at places are replaced by sediments typical of fore-arc environment.

The ITSZ continues westward into Pakistan where it is known as Main Mantle Thrust (MMT). Rocks within the suture are mostly Cretaceous or Early Tertiary in age. Geological evidence shows that all of the Himalayan rocks to the south of the suture were once part of the Indian plate and not derived from the Asian plate. Immediately to the north of the ITSZ is a Trans-Himalayan Batholith for almost entire length of the Himalaya. The batholith is a magmatic arc and formed as a result of melting of the Neo-Tethys oceanic floor which subducted beneath the Asian plate, prior to India-Asia collision. The Batholith received different names, Kohistan batholith in Pakistan, Ladakh batholith in Indian Territory; Kailas and Gandese pluton in southern Tibet; while in Arunachal Himalaya it is called Lohit batholith. The island arc rocks of Ladakh in India (and of Kohistan arc in Pakistan) are Cretaceous to Early Tertiary in age, ranging from 140 to 60 Ma. These rocks invade and overlie the Precambrian-Early Mesozoic continental crystalline rocks. However, the volcanic rocks in Kohistan arc appear to be part of an intra-oceanic island arc.

To the north of the ITSZ is the Southern Tibetan plate with cover of Palaeozoic and Mesozoic sediments that are intruded by the Cretaceous-Eocene (Powell and Conaghan, 1975) Trans-Himalayan granite batholith, the Kangdse Granite, formed in response to the northward underthrusting of the Tethyan Ocean, prior to India-Tibet (Asia) collision. Its equivalent unit in the Himalayan sector of Pakistan is the Kohistan arc that is formed within the Tethys in Mid-Cretaceous times, ultimately to become squeezed between the converging Indian and Asian Plates at about 100 Ma ago. It should be mentioned that the Kohistan arc is separated from the Karakoram Granite batholith to the north by a suture called Bangong-Nuijiang (Anduo) suture. Further north, the Northern Tibetan Block (Quangtong) is separated from the Southern Tibetan Block by an E-W suture, called the Kakoxili suture (Chang et al., 1986). The Northern Tibetan Block is apparently equivalent to the Central Pamir block as it is characterized by Precambrian basement overlain by a Silurian-Permian sedimentary sequence of Gondwana affinity, extensive Cretaceous calc-alkaline volcanics and a thick Tertiary sequence.

Before we discuss plate tectonic evolution of the Himalaya as a single fold belt, we give below the geological characteristics based on agreed observations and geochronological data.

1. The Himalayan lithounits are bounded by thrusts/shear zones, running along the length of the Mountain belt.
2. The Himalayan thrusts are north dipping.
3. The Higher Himalayan Crystallines (HHC or GHS) are characterized all along by inverted metamorphism wherein higher grade metamorphism is in higher structural levels and not in the deeper nappes.
4. Metamorphic isograds are nearly parallel to the regional foliation in the Himalayan Metamorphic Belt (HMB).
5. SW-vergent folds and NNE-SSW stretching lineations are widespread in the (HMB).
6. Higher Himalayan Crystallines (HHC) were thrust along the Main Central Thrust (MCT) on the Lesser Himalyan rocks (LHC). Erosion of the hanging

wall rocks (i.e. HHC) left them as Klippe (erosional remants), while exposures of footwall by erosion of the HHC show the LHC as tectonic windows at several places.

7. Despite being overridden by HHC, the LHC rocks do not show higher pressure assemblages!
8. The basement to the Himalayan sediments is Archaean to Proterozoic metamorphic rocks, similar in nature to the rocks of the northern Indian continental crust.
9. Although protolith ages of the Himalayan metamorphic rocks of Higher Himalayan Crystallines (HHC) and Lesser Himalayan Crystallines (LHC) are Precambrian, their mineral ages are Tertiary.
10. Leucogranite of 18–20 Ma age occurs all along the length of the HHC in the upper levels, near the contact of the HHC with the Tethyan sedimentary zone along Trans-Himadri Fault.
11. Subduction polarity in the Himalaya is similar to that in the Alps, but convexity of the Himalaya is just opposite to that of the Alps and so is the metamorphism; Alps shows normal Barrovian zones while the Himalaya shows inverted metamorphism.

With this basic information, we now describe evolution of the Himalayan fold belt.

3.2.2 Evolution of Himalaya

The Himalaya formed in response to the collision of India with Asia. The collision was brought about by the northward migration of the Indian plate at about 180 Ma ago. As the intervening oceanic crust of Tethys Sea was subducted below the southern margin of the Asian plate, India's journey started northward for about 2500 km. Before the evolution of the Himalaya or main continent-continent collision of Indian and Tibetan plate, there was a successive collision of terrains (or microcontinental blocks), derived from the northern margin of the Gondwanaland. This is evidenced by the presence of many E-W trending ophiolite belts that showed southward younging (ascertained by radiometric dating). Impressed by this feature, Chang and Cheng (1973) suggested that there was a successive collision of several crustal blocks after the progressive closure of several small oceans during Phanerozoic (Fig. 3.2). First, the Northern Tibet welded to Eurasia along Kokoxili suture (KS) at about 140 Ma. It was followed by a step-back in the subduction zone to the south of North Tibet. This underthrusting brought the Southern Tibet plate into juxtaposition with the North Tibet and eventually welded the North Tibet and South Tibet along the Bangong-Nujiang suture (BNS). A further step-back in underthrusting preceded the main collision event. India collided with Southern Tibet along the Indus-Tsangpo suture zone (ITSZ). The successive collision of these blocks and southward stepping of north-dipping suture zone led Allegre et al. (1984) to propose the evolutionary history of the Himalaya, depicted in Fig. 3.2. It is obvious

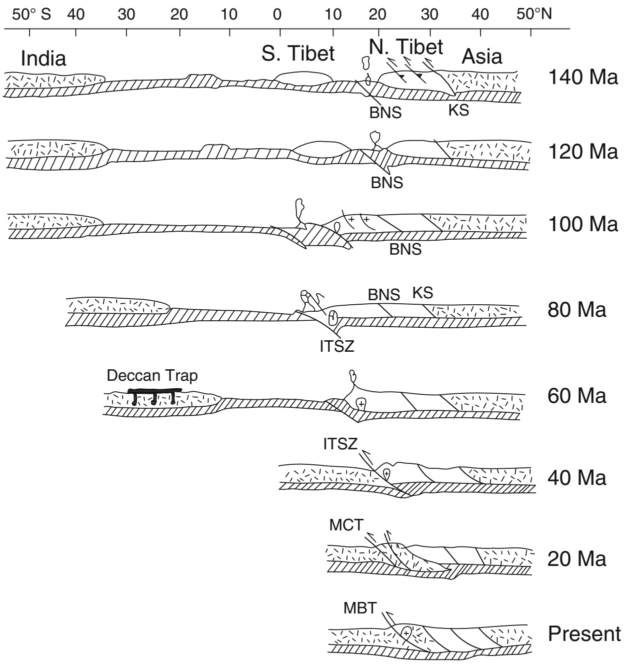


Fig. 3.2 Sequence of events in the evolution of the Himalaya (redrawn after Allegre et al., 1984). Abbreviatons: BNS = Bongong-Nujiang suture, ITSZ = Indus-Tsangpo suture zone, KS = Kokoxili suture, MCT = Main Central Thrust, MBT = Main Boundary Thrust

from this that the Himalaya contains at least two microcontinents (North Tibet and South Tibet) that represent suspect terrains.

Evolution of the Himalaya began when India, after separation from the Gondwana assembly, migrated northward to collide with the accreted Asian plate. A large area of intervening oceanic plate of ca. 2500 km—the site of Tethys which separated India and Asia—was destroyed by subduction under the Asian plate. The subducting ocean floor basalt (and some thin veneer of sediments that had deposited on the ocean floor) partially melted to produce andesitic/basaltic lava with granitoids crystallizing below it as calc-alkaline magmatic arc. This arc first developed as an oceanic Island arc, exposed as Kohistan-Ladakh arc to which the southern margin of the Asian plate (Lhasa or Tibetan Block) accreted along what is now the Shyok suture. The north-moving Indian plate also accreted to the Arc here and its eastern extension along Indus-Tsangpo suture zone (ITSZ). The collision is considered to have occurred obliquely in the NW, and the first impingement of Indian plate was in Kohistan-Ladakh sector (65 Ma ago), followed by anticlock-wise rotation of Indian plate until final collision with South Tibet (Lhasa Block of Asia) at 50 Ma ago (Dewey et al., 1989). Accretion further east developed Andean-type batholith or continental-type magmatic arc (Sharma, 1987), called the Trans-Himalyan Batholith (100–60 Ma old). This island arc ourcrops almost continuously along the southern margin of the Asian plate for a distance of about 3000 km from Pakistan to Nagaland

State of India (see Fig. 3.1). The magmatic-arc plutonic complex from Kailas mountain area is dated by U—Pb method on sphene at 107 ± 1.4 Ma with a lower intercept of 33.8 ± 7.3 Ma (Miller et al., 2001). The magmatic arc, although undeformed, experienced an episodic uplift after the Indian-Asian collision, possibly due to the underthrusting of Indian lithosphere and also because of rapid exhumation of ultra high pressure rocks of Tso-Morari dome (see further). With continued subduction, a mixture of pillow lavas, dunites, peridotites and serpentinites as hydrated ultramafic rocks transported from their birth place arrived to the final position within a convergent plate boundary where this mixture increased in volume by several hydration reactions (giving rise to serpentinite). Here, the mixture along with Cretaceous-Miocene flysch (deposited on the basaltic crust of the Neo-Tethys) was subjected to load pressure of the overlying Asian plate and developed glaucophane in basalt and greywacke compositions under high pressure and low temperature, giving rise to blue schists. The heterogenous mixture of crystallized and recrystallized rocks and marine sediments (as accretion wedge) piled up due to "bulldozing effect" by the leading edge of the Asian plate. This amalgamated pile called *ophiolite complex* was emplaced as scrapped off material of the subducting plate along weak zones (sutures) in the overlying accreted Asian plate, due to combined effect of tectonics and buoyancy. By this emplacement, the ophiolite complex occurs as tectonic fragments on both sides of the Island arc and is familiarly known as ophiolite mélange zone. The mélange on the east formed the Shyok ophiolite zone while on the west the ophiolite mélange zone became the Indus ophiolite complex that is believed to be the site of the trench zone and hence called the Indus-Tsangpo Suture Zone (ITSZ). The geology of the ITSZ ophiolite complex is comparable to that of the Shyok suture, but the two occurrences differ somewhat in their geochemistry (Ahmad et al., 2005). The ITSZ is demarcated by an ophiolite complex that is a remnant of the subducting oceanic crust, thrusted up at some stage onto downgoing continental crust of India. The ophiolites here are not well preserved complexes as elsewhere, but are dismembered lithologic sequences emplaced as tectonic slices or mélanges. The ophiolite complex not only occurs as tectonic slices or isolated blocks, but also appears as pebbles within continental detrital series (e.g. Shergol conglomerate).The emplacement of the ophiolite mélanges is believed to have raised the Trans-Himalayan Batholith at about 60 Ma ago and build the lofty mountain of the Himalaya. The occurrence of ophiolite is considered to demarcate the contact of Indian and Asian plates hence the ITSZ denotes the join between these two continental blocks. It is a matter of debate as to how the ophiolite complex was emplaced. According to one view, the ophiolites were emplaced onto the continental margin during arrested subduction of the buoyant Indian plate. To attain isostasy the Indian plate rose up, lifting on its margin a torn off slice of overriding plate as ophiolite complex. An alternative model proposes that the ophiolite complex was emplaced onto an active Indian plate by a process called obduction. Here, a part of the subducting oceanic plate detached along a pre-existing fault and was shoved onto the Indian continental margin as the Indian plate continued to subduct. Many ophiolites can thus be tectonically emplaced not only on continental crust but also over island arc, normally with a thin layer of amphibolite at the base. The radiometric age data on the glaucophane yielded 100 ± 20 to 67 ± 12 Ma (Sengor, 1989), implying that the blueschist facies metamorphism occurred much before the main Himalayan collision.

With further subduction, India came close to the accreted Asian plate and continent-continent collision of the India and Asia commenced in the Late Eocene (about 45 Ma ago). Collision along the ITSZ occurred between 45 and 55 Ma, as evidenced by paleomagnetic data (Klootwijk et al., 1992), paleontological criteria of similarity of fauna and fossils between India and Asia after collision (Sahni et al., 1981), the end-phase of subduction-related magmatism (Searle et al., 1987), stratigraphic record of the termination of Neo-Tethys ocean floor crust (Beck et al., 1995), and onset of purely terrigenous molasses along the ITSZ (Powell and Conaghan, 1973) as well as the last marine sediment in the Himalayan region as nummilitc limestone of Early Eocene age in the Lesser Himalaya (Subathu Formation) and in the northern parts of the Tethyan Himalaya, marking the regression of Tethys sea. Estimates showed that northward movement of India continued for at least a further 1500 km after the initial collision. This movement seems to have been accommodated by the crustal shortening within the colliding plates and the effects are observed in the development of folds and thrust in the Proterozoic and Lower Palaeozoic sequences of the northern Indian plate. It is estimated that the N-S compression generated due to collision of Indian-Asian plates gave rise to crustal shortening of 300–700 km within the colliding plates (cf. Kearey and Vine, 1996; see also Saklani, 1993). Structures attributed to this deformation include folds, thrust belts, strike-slip faults, and extensional rift system. The N-S compression produced immense crustal thickness in the Proterozoic and Lower Palaeozoic cover sequences of the northern Indian plate margin during underthrusting of the Indian plate. This crustal thickening was followed by thermal relaxation (Himalayan metamorphism, 32–30 Ma ago) so that the sediments and the basement rocks recrystallized (cf. Dewey and Burke, 1973), developing Barrovian regional metamorphism up to kyanite isograd in these structurally duplicated rocks, attaining maximum pressure of 8–11 kbar at temperatures up to 650°C (Searle et al., 1987).

While India began to underthrust the Asian plate, the subducting lithosphere may have anchored some continental pieces of this northern Indian margin, which recrystallized at greater depths, corresponding to the P-T stability of coesite and diamond. The metamorphosed thick crust disturbed the isostasy in this part of the Earth. The lithostatic pressure of 30 km (equivalent to 10 kbar) was due to the load of the ophiolite-bounded magmatic arc exerting on these "Indian" rocks below. As a result, the HHC metamorphic rocks exhumed only as mid crustal rocks, leaving the Indian lower crust at its own place in the colliding plate. This explains why no rocks typical of lower crust have been exhumed south of ITSZ, despite metamorphic assemblages indicating upper mantle depths (Sharma, 2008). At the time when the HHC rocks were undergoing regional metamorphism, the Lesser Himalayan rocks near the leading edge of the Indian plate also recrystallized up to garnet grade with normal metamorphic progression It is completely ruled out that the inverted metamorphism found all along the Himalayan Metamorphic Belt (HMB) was due to reverse geothermal gradient in the deep crust, for there is no physical explanation for reversal of the isotherms in the deep crust/mantle, particularly when regional metamorphic belts are commonly characterized by normal Barrovian metamorphism. The HMB rocks of the Himalaya are, unlike most regional metamorphic terrains, associated with continental collisions, where normal progressive metamorphism is found from deeper tectonic levels (showing high grade rocks) to successively higher structural levels (showing lower grade rocks). It implies that the inversion of metamorphic isograds in the Himalayan Metamorphic Belt (HMB) occurred later for which different explanations have been advanced, as discussed below. To restore isostasy, there occurred crustal extension of the deformed and crystallized continental rocks of the Indian plate margin. The Higher Himalayan Crystalline rocks were exhumed along the south directed, north-dipping Main Central Thrust (MCT), as evidenced by NE-SW trending mineral stretching and S-C fabrics in HHC rocks. As a result, the HHC were thrust upon the Lesser Himalayan rocks made of Proterozoic to Lower Cenozoic sedimentary cover of the Indian continent. Cartlos et al. (2004) state that the MCT shear zone in the Sikkim Himalaya was active during 22–10 Ma and show that the average monazite ages decrease toward structurally lower levels. They also showed that the peak metamorphism in the Lesser Himalaya occurred at 610°C/7.5 kb and 12–11 Ma and that of the MCT zone at 525°C/6 kb at 14–12 Ma. Some geologists consider it a possibility that the SW-directed stacking of the HHC was contemporaneous with the NE-directed stacking of the Tethyan Sedimentary zone. During this exhumation, the HHC rocks, particularly its pelitic schists, underwent decompression melting, forming 18–22 Ma old leucogranites that occur in the higher levels of the HHC near its contact with the Tethyan sedimentary zone. It is the dehydration melting reaction of muscovite that gave rise to the formation of migmatite and sillimanite isograd (4–6 kbar/750 $\pm$ 50°C) much after the Barrovian index mineral kyanite in the HHC rocks. The sillimanite zone thus formed occurs at the top of the inverted Barrovian isograds. Harris et al. (2004) documented decompression melting of kyanite zone metapelites in western Sikkim at 23 Ma to generate leucogranites.

Both HHC and Lesser Himalayan metamorphics exhibit *inverted metamorphism*, as stated earlier. Various models or explanations have been suggested for this inverted metamorphism. These can be grouped into three categories: (i) recumbent folding of the isograds; (ii) syn- to post-metamorphic thrusting; and (iii) shear heating. Some even suggested that ductile deformation caused translation of material along ductile shears, which consequently caused inverted metamorphism (Jain and Manickavasagam, 1993; Jain et al., 2002). In this model, translation of material occurs by grain-scale processes across a zone. This translation is accommodated by the distributed ductile shears. An analogue for this regional non-coaxial deformation could be a card deck where small amounts of slip between individual cards is likely to accommodate much larger displacements across the whole deck. This ductile deformation is assumed to translate material from deeper crustal levels whereby higher metamorphic grades overlie shallower rocks of lower grade, causing inverted metamorphism. The displacements on ductile zone will re-align pre-existing markers, e.g. lithological contacts, isograds etc, into approximate parallelism with the shear zone boundary. In this model of ductile shearing, the isograds need not be synchronous with the shearing and it is a question whether the entire Himalayan Metamorphic Belt represents a vast intracontinental ductile shear zone. Since recumbent folding is uncommon in the Himalayan metamorphic belt, unlike the Alps, inverted metamorphism cannot be attributed to a huge recumbent fold. In the model of post metamorphic thrusting, the higher temperature slab is assumed to thrust over the lower temperature slab (Le Fort, 1975, 1989). To preserve inverted metamorphism, exhumation (uplift and erosion) would have to be extremely rapid. The model of shear heating invokes generation of frictional heat during thrusting of one rock slab over the other, whereby this heat increases T in the lower slab in the immediate vicinity whilst upper slab cooling. Preservation of this inverted metamorphism requires rapid exhumation. Also, if the shear zone is wide, as in the MCT, thermal effects are unlikely to occur in the broad zone of 5–10 km, because wide ductile shear zone contains abundant micas and tourmalines which on shearing would liberate volatile phase (fluid fluxing) and hence unlikely to allow frictional heating. Joshi and Rai, (2003), after analyzing the P-T data for the shear zone transects across the MCT, demonstrated that the P-T distribution is chaotic and suggested post-metamorphic fractal displacements in the MCT zone as the cause of inverted metamorphism. The cause of the inverted metamorphism is still not completely clear and it is very likely that different models may be appropriate in different parts of the Himalaya. Dasgupta et al. (2004) investigated Lesser Himalaya of Sikkim for which P-T conditions are found to increase from 500°C/5 kb at garnet isograd to 715°C/7.5 kb at the sillimanite-K feldspar isograd, contrary to the common belief that P decreases progressively toward higher structural levels.

Recently, Sharma (2005) linked the inversion of metamorphic isograds with the southern convexity of the Himalaya; the two regional phenomena appear to be linked for this mountain belt. It is suggested by Sharma (loc. cit.) that the Indian plate during final collision was too buoyant to be carried down further (England and McKenzie, 1982). Consequently, the northward push of the Indian plate resulted in southward movement or thrusting along the Main Boundary thrust (MBT) of the deformed-metamorphosed rocks that eventually rested over the cold rock of the colliding Indian plate. This formed the Lesser Himalayan Crystallines (LHC) with their (meta-) sediments. This southward thrusting of the LHC reduced the drag load at the tip of the down-going Indian plate and the Indian plate was raised and simultaneously retreated during the ongoing collision. Sharma linked this with flipping up of the Indian plate whereby causing both reversal of the curvature of the Himalayan arc and southward movement of the hotter and deeper rocks. This happened when thrusting of the Higher Himalayan Crystallines (HHC) occurred over the lower grade rocks of the Himalayan Metamorphic Belt (HMB) and resulted in the inversion of metamorphic grade. In this event, the Indian plate once again rose and retreated, whilst the northward push continued. While the HHC moved along the MCT, the result was a decrease of P and hence decompression melting of the continental crust, which produced the 18–20 Ma old leucogranite that occur all along the HHC, close to the major structural discontinuity between HHC and the lower grade metasediments of the Late Proterozoic Haimantas toward the north (Sharma, 2005). This mechanism is shown diagrammatically in Fig. 3.3a–d, with thrusting

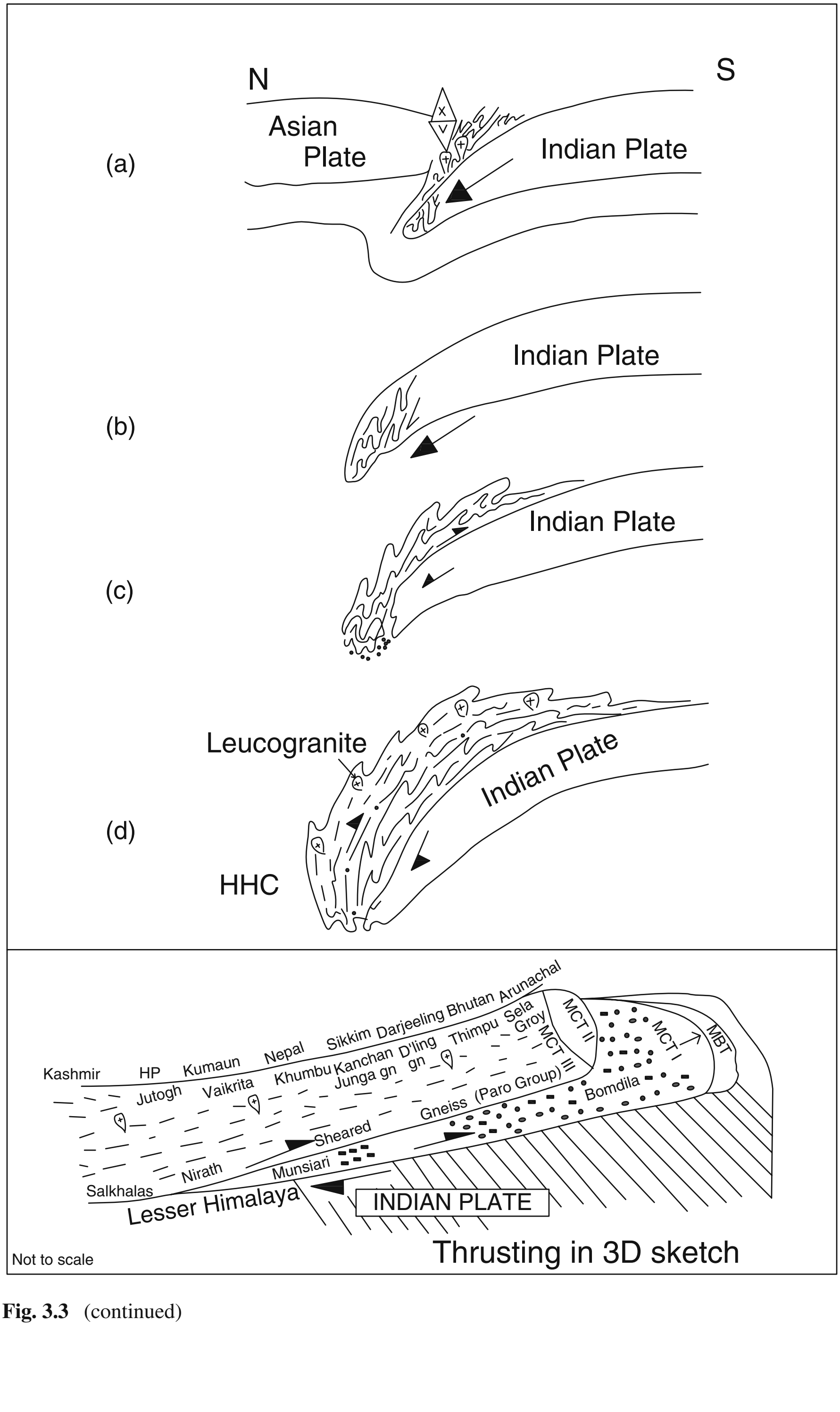


Fig. 3.3 (continued)

mechanism in a three-dimensional sketch (in bottom of Fig. 3.3). In this model, Sharma presumes that the metamorphism was normal but the reversal occurred due to successive thrusting of the sheets wherein the high grade one was overlain by the next lower grade one which in turn was overridden by still lower grade and so on, linking this inversion with the southern convexity of this lofty mountain belt.

To explain the inversion of isograds and exhumation of HHC, a newer mechanism called *channel flow model* has been proposed (see Grujic, 2006). According to this model, the HHC channel between MCT and STDS attained critical low viscosity at depth and under focused denudation and gradient in lithostatic pressure (due to high elevation of Tibetan plateau against low elevation southwards) the HHC rocks yielded to lateral flow and hence extrusion from mid-crustal depths. The lateral flow is the consequence of pure shear or Poisuille (pipe) flow in the middle and Cuette flow (simple shear) at the margins of the channel, thereby causing thinning of the channel, inversion of the isograds and exhumation of the mid-crustal rocks of the Higher Himalaya. According to this model, the inverted metamorphic isograds are not primary features but are tectonic artifacts that reflect post-metamorphic thrusting or imbrication. The HHC between SW-vergent, north-dipping Main Central Thrust (MCT) and noth-dipping South Tibet Detachment System (STDS), juxtaposed the HHC against the weakly metamorphosed Tethyan Sedimentary zone. The Himalayan inverted metamorphic rocks experienced a rapid decompression. As a result, dehydration melting of muscovite in pelitic schists occurred in the higher levels (4–6 kbar, 700 ± 50°C) of the HHC in the age range of 15–22 Ma, giving rise to the leucogranite and migmatite in the upper levels of the HHC near the North-Himadri Fault, i.e. South Tibet Detachment System (STDS).

In this model of channel flow, how much is the role of isostasy and erosion (including that by antecedent river system) is unknown, although both play a significant role in exhumation of the deep-crustal rocks, possibly without retaining their high pressure/high temperature mineralogy.

As an alternative to the channel flow model, Sharma (2007) advanced the *slab breakoff model* wherein it is considered that termination of the collision occurred

Fig. 3.3 (continued) A plausible model (after Sharma, 2005) for the inverted metamorphism in the Himalayan metamorphic belt (HMB), shown in stages: (**a**) Indian-Asian plate collision, but Indian plate was too buoyant to be carried down further, (**b**) Plastic deformation and recrystallization of the leading edge of the Indian plate, being at greater depth and hence at higher P & T, (**c**) Northward push of the subducted Indian plate resulted into southward thrusting of the plastically deformed and recrystallized rocks over the Precambrian rocks (Lesser Himalayan/basement) and overlying sediments. Emplacement of these rocks reduced drag load at the "tip" of the subducted continental crust, retreating and raising up the Indian plate, and this flip also caused southern convexity of the thrust belt all along the Himalayan length, (**d**) Continued northward push of the Indian plate resulted into renewed southward thrusting of still higher grade Higher Himalayan Crystalline (HHC) rocks during Oligocene-Miocene. Upthrusting of HHC caused decompression melting (later solidified as the Leucogranite), occasional folding of isograds and parallelism of "S" and "C" fabrics in the HHC rocks. Bottom figure shows thrusting in 3-dimension, differentiating MCT into I and II with the detachement of the Indian lithospheric slab whereby the HHC wedge destabilized, exhuming the HHC rocks either in pulses or in a single event when the ultra high pressure rocks were rapidly excavated from deeper levels.

As stated earlier, the fragments of the continental crust anchored by the lithosphere were dragged down to 100–130 km depth and developed ultra high pressure (UHP) mineralogy with coesite in garnet (Mukherjee and Sachan, 2001) or in silicate rocks and diamond in graphite-bearing rocks. It is conceived by this writer that end stages of continent-continent collision experienced slab detachment. As a result, the UHP rocks, including the eclogites (formed from basaltic parent), which had been subducted deep at the coesite and diamond stability between 55 and 45 Ma, had a rapid return due to rebound effect of the slab breakoff and resided as "nodules" in the rocks near the ITSZ. The not-so-deep subducted continental rock returned earlier and got emplaced as a distinct unit, called the Tso-Morari Crystalline (TMC) that occur as a dome between low-grade Tethyan Sedimentary zone and ITSZ. During its ascent the TMC may have possibly elevated the Tethyan sedimentary zone. Because of its early arrival and location, the TMC became the host for the UHP fragments coming later from greater depths. By this model of slab breakoff, we can understand the geological setting of the Tso-Morari Crystallines (TMC) that was once a part of the HHC but during subduction process changed its mineralogy with high pressure assemblages (Sharma, 2008). Due to its limited subduction (reflected in its mineralogy) it disengaged first from the downgoing slab and ascended rapidly to mid crustal depths and emplaced amidst Tethyan zone near ITSZ. The proposed mechanism finds support from the occurrence of post-orogenic, undeformed basic rocks (being products of decompression melting of the upwarped asthenosphere as a consequence of slab detachment) intruding the Himalayan metamorphics (Sharma, 1962) and from the deep mantle anomalies recently discovered by seismic tomography at the base of lithosphere in West Pakistan (van der Voo et al., 1999).

The geological events for the Himalayan evolution are enumerated below

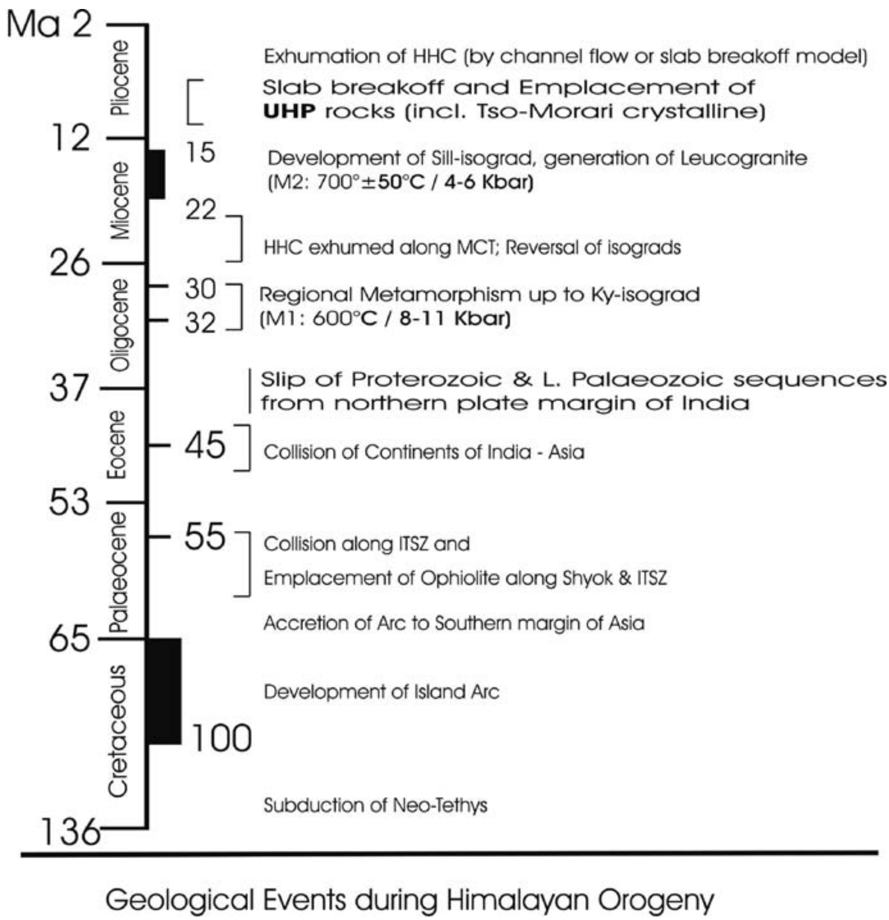
The paradigm of slab breakoff is an efficient mechanism for exhumation of the Himalayan rocks and is independent of the parameters of the channel flow model that can be challenged on several accounts Sharma (2007, p. 40) as given below.

1. Continuity of lithologic units in the HHC (GHS) for over 1000 km along strike (Gansser, 1964) indicates lack of internal stratigraphic disturbance.
2. The Model is beset with uncertainties:
   (a) Whether HHC represents a *complete section of mid crust*

   or

   An *extruded segment* of a cooling channel?
   (b) Whether the HHC bounding faults associated with *late-stage* exhumation of
   or

   Formed at depth beneath the Tibetan Plateau?



(c) Whether the fabric within HHC are related to *flow during channeling*/extrusion

or

Pre-date the Himalayan event?

3. The Model presumes that the SW-directed stacking of the HHC (GHS) was contemporaneous with the NE-directed stacking of the Tethyan Sedimentary Zone. This is invalidated by the available geochronological data which suggest that STDS was active in the interval of 17–14 Ma whereas the majority of HHC-Leucogranite, formed during slip on MCT, were emplaced between 24 and 19 Ma (Harrison et al., 1999).
4. Geochronology of metamorphic monazites from MCT in Central Nepal indicates that MCT shear zone was active at 6 Ma (Harrison et al., 1997) while ductile deformation seems to have terminated during the Early Miocene (Schelling and Arita, 1991).

5. Denudation playing a predominant role in the Model, implies that insufficient denudation is responsible for the absence of (Lower crust) granulites. If granulites are anhydrous residue of deep crustal melting, why channel flow did not operate?.
6. The Model requires rapid and large magnitude of denudation for the minor effect that decompression has on melting of the source composition.
7. The Model requires generation of multiple anatectic phases via decompression, which seems difficult and does not fulfill the requirement (by the Model) of definite timing linking Slip with STD and Anatexis.
8. Omission of parts of the Tethyan section across STDS at the top of HHC slab shows that some fault movements clearly post-date peak metamorphism (Steck et al., 1993).
9. The exhumation of HHC rocks metamorphosed first (in M1) up to kyanite grade at P of 8–11 kbar and T below anatexis of crustal rocks (Searle et al., 1992) could not have occurred by channel flow but requires another process.

The recently discovered crustal thickening beneath the Himalaya is attributed to on-going underthrusting of India. It has been proposed that the thickening was accomplished by delamination of the Indian crust from the lithospheric mantle along a brittle-ductile transition at the level of Moho. This means that some process of "mantle peeling" has taken place beneath the Asian plate so that Indian crust could be placed directly beneath the Tibetan (Asian) plate. This is revealed by the seismic tomographic analysis of Roecker (1982) and by the seismic studies of Hirn et al. (1995), both showing that no Indian lithosphere is present at depth north of the ITSZ. Therefore, another model is proposed by Molnar (1984) in which only limited underthrusting of the Indian plate is postulated and no crust-mantle decollement is required. In consideration of this proposition and that of McKenzie (1969) that the Indian crust was too buoyant to be subducted deeper under the Asian plate, Butler (1986) suggested that crustal thickening was accomplished by thrusting along a detachment climbing from the Moho. According to this model the footwall section of the crust would be depressed below the level at which the granulites of the lower crust undergo a phase change to eclogite. The increased density of the basal eclogite could then decrease the buoyancy of the crust and allow limited subduction to take place. Bouguer anomalies are found increasingly negative northwards from India, suggesting that the crust thickens in this direction. A detailed imaging of crustal structure across the Himalaya and Tibet shows crustal thickness of ca. 35 km of the Indian shield, and this increases to 55 km under the Himalaya and to 70 km beneath Tibet. Furthermore, the Moho topography is not smooth and exhibits a number of breaks or steps and at places overlapping of Moho at depth. This suggests that crustal thickness did not occur by simple underplating and that India did not descend smoothly and coherently beneath Asia; rather it took place in response to intracontinental thrusting affecting both the crust and the upper mantle.

A combined interpretation of satellite images and focal mechanism solution of earthquake has revealed that the pattern of faulting in the Himalayan region is such that thrust faulting is restricted to a relatively narrow belt north of the Himalayan

Frontal Faults. Strike-slip faulting is dominant in a region some 1500 km wide to the north of the Himalaya and eastward into Indo-China. The MBT and MCT thrusts can be cited as two most spectacular examples of thrust faulting. The high-grade crystallines (HHC) which behaved like a basement to the Tethyan sediments were also transported on the Indian continental margin, along MCT, to rest on the Lesser Himalayan low grade metasedimentaries and seen further south to cover even the Eocene rocks. It needs to be pointed out that there is a serious confusion in the use of the term LHC. The LHC are believed by most workers as Lesser Himalayan Crystallines equivalent of HHC which are geographically located in the Lesser Himalaya but are equivalent to HHC.

Two major strike faults are remarkable on the west and east end of the Indian plate. At its western end, the Indus suture is terminated by a large sinistral strike fault, the Quetta-Chaman fault that continues into the Indian Ocean as Owen Transform Fault. At the eastern end of the Indus-Tsangpo Suture there is a large dextral strike fault, the Saquang Fault, which continues through Burma to connect with the Indonesian subduction zone. This fault perhaps also continues into the Indian Ocean as the now largely inactive Ninety-East Ridge Transform. The Indian plate, therefore, may be regarded as a two-pronged wedge that has driven northwards into the Tibetan plate between two large transform faults. An analogy has been drawn between the pattern of faulting in Asia when it is indented by a rigid indenter of India and the lines of failures (slip lines) developed in a plastic medium experimented by Tapponier and Molnar (1976). Alternative to this Indentation model, there is another model, known as Continuum model that places greater emphasis on the role of forces arising from crustal thickening (Tapponier et al., 1982). Both indentation model and continuum model have been challenged by England and Molnar (1990) who suggested that the major sinistral strike-slip faults are a consequence of the presence of a wide deforming zone between India, China and Eurasia (see also, Lyon-Caen and Molnar, 1983). If the faults rotate in a clockwise sense, as suggested by recent palaeomagnetic measurements, their sinistral motion arises from their location in a broad region undergoing dextral shear in the N-S direction. This model explains the eastward motion of Tibet relative to Asia and India, and its northward movement relative to SE China (Fig. 3.4).

The Himalaya is undergoing rapid uplift at rates between 0.5 and 0.4 mm per year. It is, therefore, experiencing rapid erosion, with deposition of these clastic sediments in a foredeep or foreland basin. These Siwalik molasses reach a thickness of as much as 6–8 km near the Himalayan front, and the coarseness of the basin fill decreases away from the mountain front. Conglomerates pass into sandstones and shales, extending southwards into the Ganga basin where they are bounded by the Main Frontal thrust (MFT). In some foredeep, the clasts in the conglomerates and sandstones reveal an unroofing sequence in which stratigraphically younger deposits contain debris from successively deeper levels in the mountain.

Geophysical evidence suggests that in the Himalaya, oceanic portions of the Indian plate are still being underthrust at the Indonesian arc (Zhao et al., 1993). In the west of India, oceanic crust of the northwest Indian Ocean is still subducting beneath the Makran region of southwest Pakistan and Iran (see Moores and Twiss, 1995).

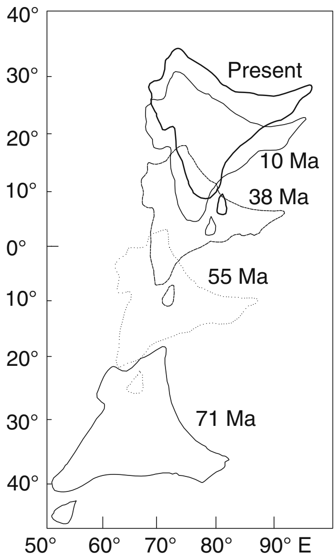


Fig. 3.4 Northward drift of India with respect to Asia from 71 Ma to the present, determined from magnetic lineations in the Indian and Atlantic oceans (redrawn from Molnar and Tapponier, 1975)

3.3 Old Fold Belts

After having discussed the plate tectonics theory and its application to the evolution of the Himalayan orogenic belt, we now look at the characteristics of the older fold belts, mainly Proterozoic fold belts of India. Here we consider the plate tectonic events that may have led to their development. The record of the older plate tectonic activity is expected to be found in continental rocks and especially in the Proterozoic fold belts. Here we need to recognize the plates that converged, their direction of movement, the collision suture, as well as the magmatic arc as relict arc on the flanking continent. For the nature of the Proterozoic plate tectonic events, we need to rely on the palaeomagnetic evidence for the plate movement. However, we do not have palaeomagnetic data in any of these mobile belts and their flanking cratonic areas. Therefore, we have to search for other "petrotectonic indicators", such as blueschist and calc-alkaline rocks that suggest subduction, which might have occurred in the concerned fold belts. We may fail to recognize ophiolite in the Proterozoic fold belts for some reasons such as thermal overprinting, although ophiolites and blueschists have been reported from Middle to Lower Proterozoic terrains. Blueschists are characteristics of consuming margins, and ophiolites are typical in collision belts. The lack of blueschists and ophiolites presents problems in any attempt to apply plate tectonics models to the Middle to Early Proterozoic terrains. Their absence in Proterozoic fold belts implies either that the plate tectonics of Palaeozoic type has not operated or that some other petrotectonic indicators have taken the place of blueschists and ophiolites. If the Precambrian geotherm was higher, the metamorphic conditions in the subduction zone might have been a high-P, low-T to form rocks such as kyanite $\pm$ talc schists that occupied the tectonic position of blueschists in the Proterozoic fold belts. It is also possible that the deformed ultramafic complexes that occur in some ancient fold belts represent the subducted oceanic crust during Precambrian times. Despite these problems, mantle convection cannot be denied for the Precambrian times and some form of plate tectonics has to be conceived as pleaded by Kroener (1981).

In considering the nature of global tectonic activity in Precambrian times, Kroener (1981) suggested three approaches. First, a strictly uniformitarian approach can be taken in which Precambrian tectonics originated by the same mechanism of plate tectonics actively operated in Phanerozoic times. Second, a modified uniformitarian approach can be postulated in which plate tectonic processes in the Precambrian were somewhat different from present because the physical conditions affecting the crust and mantle have changed throughout geological time (Hargraves, 1981). Third, a completely different tectonic mechanism can be invoked for Precambrian times. Only the first two of these approaches will be considered, as there are no cogent reasons for not doing so.

The oldest orogenic system that originated in plate tectonic time is the Pan African-Baikalian-Brazilian system present in East Asia and throughout much of Africa and South America. It dates back to Neoproterozoic (1000–550 Ma) and has many features that also resemble those of Mesozoic-Cenozoic orogenic regions. Long-distance correlations are not possible but it is suggested that around ca. 1000 Ma ago, a supercontinent formed by the collision of formerly separate continents to develop the Grenville orogenic belt and its correlatives. At this time, North America and East Gondwanaland (Australia and Antarctica) may have been joined. In the latest Precambrian time, this supercontinent (called Rodinia by some geologists) broke up with the separation of Laurentia, and another supercontinent formed, because of the Pan African-Baikalian-Brazilian orogeny. The Canadian geologist, Paul Hoffman suggests that one supercontinent called Gondwanaland may have developed by turning inside out (or extraversion) of a previous supercontinent. This extraversion resulted in the rifting of Western North America away from Antarctica-Australia and the opening of the Protopacific (Panthalasa) Ocean. At the beginning of Palaeozoic times, the plates were rearranged into a supercontinent. In Cambrian times, this supercontinent fragmented with the opening of the first proto-Atlantic ocean. This closed and the supercontinent reassembled in Late Silurian times, producing the *Caledonian orogeny*. In Early and Middle Devonian times, transcurrent movements between the Northern and Southern parts of the supercontinent produced the *Acadian orogeny*. A second phase of opening in the Late Devonian created the second proto-Atlantic, which closed in Middle Carboniferous times, producing the *Variscan orogeny* and the re-assembly of the plates into the supercontinent, the Pangae.

In the early Palaeozoic, orogeny during Middle to Late Ordovician is called Taconian in the USA, Grampian in the Canada and Fennarkian in the Scandinavia. A Middle Palaeozoic orogeny during Silurian-Devonian is known as the Acadian orogeny in North America and Caledonian orogeny in Great Britain and Scandinavia. A third orogeny in Late Carboniferous (Pennsylvanian)-Permian is referred to as the Alleghanian orogeny in the USA and Hercynian-Variscan in southern Europe. The timing of these orogenies is different at different places along the strike of the Appalachian-Caledonian system. This belt is probably the best example of an orogen that conforms to the composite orogenic model, evolved in the plate tectonic times. The presence in it of foredeep sediments, fold-and-thrust belts, a core with nappes and ophiolites, as in Newfoundland and Scandinavia, are interesting features that need to be looked in also for Proterozoic fold belts of India.

3.4 Proterozoic Fold Belts of India

As stated earlier, the Proterozoic fold belts are of special interest from the geotectonic point of view because it is believed that Proterozoic plate tectonics/ convergence were responsible for the evolution of these fold belts (Condie, 1982).

In India the Proterozoic fold belts are well developed in Rajasthan (western India), in Singhbhum-Orissa (eastern India), Madhya Pradesh and Maharashtra (central India), Eastern Ghats and Southern Granulite Terrain. All the fold belts are peripheral to the cratonic areas made up of Archaean gneiss-amphibolite-migmatite-granite association that served as basement for the supracrustal rocks forming the fold belts. These basement rocks also crop out amidst the deformed and metamorphosed supracrustal rocks of the fold belts, suggesting that both cover and basement were folded together in the mountain building orogeny.

For the Proterozoic mountain building process (orogenesis), there could be Phanerozoic style of plate tectonics in which lithospheric plates from far off distances approached towards each other because of subduction of the intervening oceanic crust (Black et al., 1979). In this mechanism it is very important to identify the colliding plates, a suture zone along which subduction occurred, a magmatic arc which formed of melts generated from subducting plate, and other features like ophiolite belt—a remnant ocean floor. This belt is the most direct evidence of a collision suture, but often unrecognizable in the Proterozoic fold belts. Some sutures, on the other hand, are characterized by mylonitic or ductile shear zone. Without collating other evidence, these sutures do not necessarily imply the presence of a subduction. These sutures are likely to form in ensialic orogenesis, slightly different from that by plate tectonics where orogeny is associated with ocean floor subduction. In the ensialic orogenesis, the rift basin is floored by continental rocks and not by oceanic crust. The structures in older terrains on either side of the fold belt are virtually unaffected (albeit overprinted) through the younger mobile belt. On the other hand, the sialic blocks, welded together by plate tectonics process, show markedly different stratigraphy or tectonic history and have a discontinuity in the orientation or style of structures. Nearly all the Proterozoic fold belts of India seem to have evolved by the ensialic orogenesis that is described below.

3.5 Ensialic Orogenesis Model

The ensialic orogenesis is a modified version of plate tectonic theory, appears suitably applicable for the evolution of the Proterozoic fold belts of the Indian shield. The ensialic orogenic model initially starts with segmentation of once large sialic block due to crustal thinning by ductile stretching or rifting. Extension and thinning of the continental crust eventually causes the floor of the rift valley to drop below sea level. This depositional basin then receives sediments that are laid upon sialic continental basement, and nowhere upon mafic oceanic crust. The volcanic lava or basic flows, generated by decompression melting of the upper mantle are likely to interrupt the sedimentation processes. Subsequently, these supracrustals (sediment + lava) along with their basement are compressed, deformed, and metamorphosed and even partially melted at depth. The whole rock complex is finally raised in an orogenic or mobile belt that is accreted to the converging sialic blocks. The ensialic orogenesis involves only a reworking of the older continental material and is thus incompatible with the crustal accretion from far off distances as in the plate tectonic mechanism. However, in the ensialic orogenesis, a limited subduction is envisaged which is called Ampferer subduction (abbreviated A-subduction). Here the mantle part of the lithosphere is conceived to detach from the crust and the crust eventually undergoes heating from the mantle magma to result in metamorphism of both basement and cover rocks of the region. Thus, it seems that in the ensialic orogenesis model the structures in older terrains on either side of the fold belt are virtually unaffected (albeit overprinting) through the younger mobile belt. Sharma (2003) explained the reversal of the movement direction of the crustal blocks that had drifted away from the site of continental rifting. He suggested that sediment load and decay of the underlying “plume” or mantle diapir and consequent change in vertical component of stress are the main cause for reversal of the movement direction of the separating crustal blocks. Geological evidence in nearly all Proterozoic fold belts of India suggests that they owe their origin to ensialic orogenesis model that is a modified plate tectonic process in Phanerozoic time. It is believed that the stable Archaean cratons were subdivided by mobile belts in which deformation is almost wholly ensialic.

Toward the end of Archaean, 2.5 Ga ago, major global changes seem to have occurred, and one of the significant changes was demise of Archaean processes, reflected in global stabilization of continental areas or cratons. The Archaean crust subsequently underwent ductile stretching or rifting because of drag effects of sub-lithospheric convection currents. This resulted in the formation of "geosynclinal" basins whose floor was made of continental crustal rocks or oceanic crust (true geosyncline). These geosynclinal basins became the site for deposition of submarine basalts (poured out as a result of decompression melting of upper mantle) and of shallow to deep-water facies sediments and clastics derived in part from the pre-existing terraines. These supracrustal rocks of the basin were subsequently deformed by convergence of opposing continental blocks/lithospheric plates. This deformation resulted into thickening of the crust that led to heating (radiogenic and even by heat from igneous intrusions) and hence metamorphism of the rocks which ultimately were raised into lofty mountains or fold belts. Thermal modelling of crustal overthickening tectonics (England and Thompson, 1984) has indicated that increased heat production and insulation are the most important factors for the perturbed geotherm, whereas the effects of erosion and isostasy are important in terminating metamorphism.

Heating of a crustal segment could also take place by mere lithospheric thinning and magma underplating/interplating. Because of the magma addition below or within the crust, the affected rocks are heated to very high temperatures, designated as ultra high temperature (UHT) metamorphism, resulting in the formation of granulite facies rocks. These rocks are found to show anticlockwise (ACW) trajectory in the P-T space, unlike the rocks of geosyncline that document clockwise P-T path. There is no reasonable thermal scenario for creating anticlockwise P-T path with simple burial. Such paths clearly appear to require magmatic heating early in their evolution. This heating can be brought about by (a) accretion of large volumes of mantle-derived magmas in the lower/middle crust (Harley, 1988), and (b) lithospheric thinning or crustal extension due to which basic magma generated by decompression melting of mantle would move upwards to underplate the lower crust (Sandiford and Powell, 1986). Such mechanisms are proposed for the Eastern Ghats belt (Dasgupta and Sengupta, 2002), discussed later. Admittedly, extreme mantle thinning is capable of generating granulite facies conditions but excessive crustal lithosphere thinning would give rise to such rapid heat loss from the crust that granulite facies conditions cannot be attained. It means that granulite metamorphic conditions can only be generated by extreme lithosphere thinning with minimal crustal thinning. Such a scenario is most realistically generated by asymmetric lithosphere extension (Sandiford and Powell, 1986).

The anticlockwise P-T path for most granulite facies rocks is often beset with a problem of their excavation from deep crustal levels. Some granulite terrains with ACW path show isothermal decompression which implies a rapid erosion or gravitational spreading/stretching of tectonically thickened crust (cf. Mohan et al., 1997). Difficulties arise, however, when the retrograde path of the ACW is an isobaric cooling. This implies that these granulites had their residence at great depths and in order to be exhumed they must be involved in another orogeny or upthrusted along a shear zone. This controversy of the Eastern Ghats belt has been discussed by Bhattacharya and Gupta (2001). In order to resolve a type of tectonic setting during metamorphism, the P-T data, obtained from calibration of geothermobarometric models and from petrogenetic grids based on experimental mineralogy and textural relationship, are considered in conjunction with geochronological data. However, these critical data are not always available for all the Proterozoic fold belts of India, believed to have evolved through the ensialic orogenesis.

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