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The longest voyage: Tectonic, magmatic, and paleoclimatic evolution of the Indian plate during its northward flight from Gondwana to Asia

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ABSTRACT

The tectonic evolution of the Indian plate, which started in Late Jurassic about 167 million years ago (~167 Ma) with the breakup of Gondwana, presents an exceptional and intricate case history against which a variety of plate tectonic events such as: continental breakup, sea-floor spreading, birth of new oceans, flood basalt volcanism, hotspot tracks, transform faults, subduction, obduction, continental collision, accretion, and mountain building can be investigated. Plate tectonic maps are presented here illustrating the repeated rifting of the Indian plate from surrounding Gondwana continents, its northward migration, and its collision first with the Kohistan–Ladakh Arc at the Indus Suture Zone, and then with Tibet at the Shyok–Tsangpo Suture. The associations between flood basalts and the recurrent separation of the Indian plate from Gondwana are assessed. The breakup of India from Gondwana and the opening of the Indian Ocean is thought to have been caused by plate tectonic forces (i.e., slab pull emanating from the subduction of the Tethyan ocean floor beneath Eurasia) which were localized along zones of weakness caused by mantle plumes (Bouvet, Marion, Kerguelen, and Reunion plumes). The sequential spreading of the Southwest Indian Ridge/Davie Ridge, Southeast Indian Ridge, Central Indian Ridge, Palitana Ridge, and Carlsberg Ridge in the Indian Ocean were responsible for the fragmentation of the Indian plate during the Late Jurassic and Cretaceous times. The Réunion and the Kerguelen plumes left two spectacular hotspot tracks on either side of the Indian plate. With the breakup of Gondwana, India remained isolated as an island continent, but reestablished its biotic links with Africa during the Late Cretaceous during its collision with the Kohistan–Ladakh Arc (~85 Ma) along the Indus Suture. Soon after the Deccan eruption, India drifted northward as an island continent by rapid motion carrying Gondwana biota, about 20 cm/year, between 67 Ma to 50 Ma; it slowed down dramatically to 5 cm/year during its collision with Asia in Early Eocene (~50 Ma). A northern corridor was established between India and Asia soon after the collision allowing faunal interchange. This is reflected by mixed Gondwana and Eurasian elements in the fossil record preserved in several continental Eocene formations of India. A revised India–Asia collision model suggests that the Indus Suture represents the obduction zone between India and the Kohistan–Ladakh Arc, whereas the Shyok-Suture represents the collision between the Kohistan–Ladakh arc and Tibet. Eventually, the Indus–Tsangpo Zone became the locus of the final India–Asia collision, which probably began in Early Eocene (~50 Ma) with the closure of Neotethys Ocean. The post-collisional tectonics for the last 50 million years is best expressed in the evolution of the Himalaya–Tibetan orogen. The great thickness of crust beneath Tibet and Himalaya and a series of north vergent thrust zones in the Himalaya and the south-vergent subduction zones in Tibetan Plateau suggest the progressive convergence between India and Asia of about 2500 km since the time of collision. In the early Eohimalayan phase (~50 to 25 Ma) of Himalayan orogeny (Middle Eocene–Late Oligocene), thick sediments on the leading edge of the Indian plate were squeezed, folded, and faulted to form the Tethyan Himalaya. With continuing convergence of India, the architecture of the Himalayan–Tibetan orogen is dominated by deformational structures developed in the Neogene Period during the Neohimalayan phase (~21 Ma to present), creating a series of north-vergent thrust belt systems such as the Main Central Thrust, the Main Boundary Thrust, and the Main Frontal Thrust to accommodate crustal shortening. Neogene molassic sediment shed from the rise of the Himalaya was deposited in a nearly continuous foreland trough in the Siwalik Group containing rich vertebrate assemblages. Tomographic imaging of the India–Asia orogen reveals that Indian lithospheric slab has been subducted subhorizontally beneath the entire Tibetan Plateau that has played a key role in the uplift of the Tibetan Plateau. The low-viscosity channel flow in response to topographic loading of Tibet provides a mechanism to explain the Himalayan–Tibetan orogen. From the start of its voyage in Southern Hemisphere, to its final impact with the Asia, the Indian plate has experienced changes in climatic conditions both short-term and long-term. We present

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a series of paleoclimatic maps illustrating the temperature and precipitation conditions based on estimates of Fast Ocean Atmospheric Model (FOAM), a coupled global climate model. The uplift of the Himalaya–Tibetan Plateau above the snow line created two most important global climate phenomena—the birth of the Asian monsoon and the onset of Pleistocene glaciation. As the mountains rose, and the monsoon rains intensified, increasing erosional sediments from the Himalaya were carried down by the Ganga River in the east and the Indus River in the west, and were deposited in two great deep-sea fans, the Bengal and the Indus. Vertebrate fossils provide additional resolution for the timing of three crucial tectonic events: India–KL Arc collision during the Late Cretaceous, India–Asia collision during the Early Eocene, and the rise of the Himalaya during the Early Miocene.

Contents

1. Introduction

The tectonic evolution of the Indian plate from its original location in the Gondwana supercontinent during Permian through Middle Jurassic time, its sequential separation from other Gondwana continents, its continual fragmentation, its northward motion as an island continent, and its successive collisions, first with the Kohistan–Ladakh Arc and then with Asia, represents one of the longest journeys of all continents, about 9000 km in 160 million years (Dietz and Holden, 1970; Chatterjee, 1992; Chatterjee and Scotese, 1999, 2010). India has had one of the most complicated tectonic histories of all of the Gondwana continents. This tectonic history was shaped by complex breakup and dispersal events, and modified by flood basal volcanism and collision events.

The rifting events of the Indian plate can be reconstructed from the magnetic anomalies of the Indian Ocean floor. The Indian Ocean is one of the most diverse oceans on the face of the globe, both in relief and origin of seafloor features, and contains every type of plate boundary. The origin and evolution of the Indian Ocean is the result of the breakup of the Gondwana supercontinent from Early Jurassic onwards (~167 Ma), and by the northward movement of the Indian plate, which began colliding with Asia about 50 Ma. Its formation involves the drifting of the Gondwana fragments of various sizes including Africa, India, Australia, Antarctica, Arabia as well as several smaller continents such as Sri Lanka, Madagascar, Seychelles, and the Laxmi Ridge. Although it opened some 125 Ma, the Indian Ocean had taken on its present configuration by 36 Ma (Royer and Coffin, 1992; Reeves and de Wit, 2000).

The tectonic development of the Indian Ocean is complex, but well understood (McKenzie and Sclater, 1973; Reeves and de Wit, 2000). On India's passage to the north during the Cretaceous, its eastern and western margins were marked by two long, linear, hot spot tracks (Duncan, 1981). The successive eruptions of the Rajmahal Traps covered part of eastern India in the Early Cretaceous and the Deccan Traps at the Cretaceous–Tertiary transition covered much of western and central India (Courtillot, 1999). Environmental catastrophes induced by Deccan volcanism may be linked to the mass extinction at 65 Ma (Keller et al., 2011). The Indian plate underwent an unusual sudden acceleration during its northward journey from Late Cretaceous (67 Ma) to Early Eocene (50 Ma), about 20 cm/year, and then dramatically slowed to 5 cm/year during its collision with Asia (McKenzie and Sclater, 1971; Patriat and Achache, 1984; Lee and Lawver, 1995). The collision of India with Asia is perhaps the most profound tectonic event of the Cenozoic, and is responsible for the uplift of the Himalayan–Tibetan Plateau.

India provides an elegant natural laboratory for the study of continental rifting and continental collision. The present-day configuration of the coastal margins of India is the consequence of five episodes of sequential rifting events since the Early Jurassic (~167 Ma) time. Each rifting event is associated continental flood basalt volcanism. India became smaller and smaller during its rifting, leaving behind several smaller continents such as Sri Lanka, Madagascar, Laxmi Ridge, and Seychelles. In this paper we bring together recent data on the spatial association and temporal sequence of rifting of the Indian plate from Gondwana, the recurrent eruption of flood basalts through time and space, and reconstruct the thermal history of the evolving continental margins of India. We review the repeated rifting events of the Indian plate from magnetic anomalies, paleomagnetism, and hot spot tracks.

During its long journey, the Indian plate underwent both divergent and convergent tectonic regimes and the fragments were welded to form a larger continent. The collision of India with the Kohistan–Ladakh Arc, and then with Asia, contributed to the development of the Eurasian supercontinent. The popular model of subduction of the Neotethyan ocean floor along the Indus-Suture Zone is oversimplified. The recognition of two subduction zones within the Neotethys suggests that India first collided with the Kohistan–Ladakh Arc at the Makran–Indus Trench and then to Asia at the Shyok–Tsangpo Trench (Reuber, 1986; Allègre, 1988; Van der Voo et al., 1999; Ali and Aitchison, 2008; Jagoutz et al., 2009a, 2009b; Chatterjee and Scotese, 2010; Burg, 2011). We have proposed a model for the evolution of the Indus–Tsangpo Suture as the final zone of collision between the Indian plate and the Lhasa terrane.

The Himalayan–Tibetan orogen is an ideal place for the study of continent–continent collision because this orogen is seismically active and has produced a variety of geologic features such as subduction, large-scale thrust, strike-slip and normal faults, magmatism, channel flow, and regional metamorphism (Gansser, 1964; Molnar and Tapponnier, 1977; Hodges, 2006a, 2006b; Aitchison et al., 2011; Zhu et al., 2012). New seismic tomography provides a snapshot of the deep earth structure beneath the Himalayan–Tibetan Plateau and suggests that subduction of the Indian lithosphere under Asia played an important role in the tectonic evolution of the Himalaya–Tibetan Plateau and surrounding regions (Van der Voo et al., 1999; Kind et al., 2002; Li et al., 2008; Replumaz et al., 2010).

We have used a dynamic climatic model (Fast Ocean Atmospheric Model or FOAM) to illustrate the paleoclimatic evolution of India during its northward journey. The FOAM is a fully coupled mixed-resolution, general circulation model designed to address climate science questions with high-throughput (simulated years/day) while still providing a good simulated mean climate (Jacob et al., 2001). Examples of other widely used Atmosphere–Ocean General Circulation Models (AOGCMS) include CCSM (from the National Center for Atmospheric Research), CGSM (from the Canadian Center for Climate Modeling and Analysis), GFDL (from the Geophysical Fluid Dynamics Laboratory of National Oceanic and Atmospheric Administration), HadCM (from the Met Office, UK's National Weather Service) etc. We have discussed how the uplift of the Himalaya–Tibetan Plateau has greatly influenced the Neogene climate of Asia and created the monsoons (Hodges, 2000; Kent and Muttoni, 2008; Zhisheng et al., 2011) and triggered Pleistocene glaciation (Kuhle, 2002).

2. Continental breakup and dispersal

2.1. Supercontinent cycle

Throughout the Earth's history, the continents have undergone a process of collision and rifting several times in an episodic fashion. Smaller continental blocks collide and merge into larger supercontinents, which then later break apart. There have been approximately six major continental assembly and breakup events, about 300–500 million years apart, which have caused dramatic sea level changes, destroyed and recreated ecological niches, and affected the global climate and evolution of life. These supercontinents began with Ur (3.0 Ga), followed by Kenorland (2.7–2.5 Ga), Columbia (1.9–1.8 Ga), Rodinia (1.1 Ga), Pannotia/Gondwana (0.54 Ga), and Pangea (0.25 Ga). Other postulated supercontinents such as Valbaara (3.2 Ga) lack critical evidence to support their existence (Santosh et al., 2009).

Currently the Earth is in a collision cycle in which Africa, Arabia, and Australia are in collision with Eurasia. Our geologic record of supercontinent cycles is only well documented for the last two cycles: Rodinia and Gondwana–Pangea. The Proterozoic supercontinent Rodinia broke apart to form smaller continents during the Early Paleozoic, which were then reunited to form the supercontinent Pangea in the Late Paleozoic. Similarly, Pangea began to disintegrate into smaller plates during the Late Jurassic, which drifted apart to form the modern ocean basins. The plate motions are still continuing. Continental collision makes supercontinents, while rifting makes numerous, smaller continents with intervening new oceans. Throughout geologic history, the lithosphere has been dynamic and has remained continuously in a state of flux. The supercontinent cycle must have changed the configuration of continents and oceans episodically, renewed the lithosphere, and influenced the climatic and biological evolution.

2.2. Rifting and dispersal of continents

Since the tectonic evolution of the Indian plate is largely a story of continual breakup from Gondwana and its dispersal through time, at the beginning we may address this issue of intra-continental rifting by asking, “Why do the continents break up and drift apart?” Earth history is punctuated by recurrent episodes of continental assembly and continental breakup, but the causes of continental rifting and dispersal are still debated (Storey, 1995; Segev, 2002). Wegener's (1915) theory of continental drift is one the greatest paradigm shifts of the Earth Sciences. He envisaged the continents were moving through the oceanic crust and proposed that the cause of the continental rifting might be the centrifugal force of the Earth's rotation. This driving mechanism was criticized by physicists of that time who believed that these centrifugal forces were much too weak to move the continents. No doubt Wegener's inability to provide a convincing mechanism for continental drift contributed to the rejection of his theory. Though plate tectonic theory has been accepted today the mechanism that causes plates to break apart is not widely known and misconceptions still exist concerning the role the mantle plays in plate motions.

Some have suggested that the rotation of Earth and its oblate ellipsoidal shape may be a contributing factor (Doglioni, 1990). When it was shown by plate tectonics that continents and ocean floor move together, not separately, and that new lithosphere forms at spreading ridges by rising magma, and that volcanoes are common along the trench boundaries, most geologists accepted that some type of convective flow—in which subducting slab of cold, oceanic lithosphere sinks, and upwelling of mantle plume rises—might be responsible for surface plate motion (Condie, 2001). The question still remains, however, "What exactly drives the plates?" Both plate-driving forces (Forsyth and Uyeda, 1975) and plume-assisted forces (Morgan, 1981) have been proposed for rifting and drifting of the thick continental lithosphere.

2.3. Plate tectonic forces and continental rifting

The relationships between ridges, trenches, transform faults, and continents continuously change on a sphere, thereby changing the boundary conditions and the interplate stresses (Anderson, 1994). It is generally believed that some type of convective flow driven by the sinking of the cooler, denser oceanic lithosphere (slab pull) is the principal driving force of plate tectonics. In the slab-pull mechanism, the cold subducting slab of oceanic lithosphere, being denser than the surrounding warmer asthenosphere, pulls the rest of the plate with it as it descends into the asthenosphere. A number of other lithospheric forces such as, the push from the mid-ocean ridges (ridge-push), slab suction towards trench, basal drag, and collisional resistance at plate boundaries also have been proposed (Forsyth and Uyeda, 1975; Kearey and Vine, 1990).

At ocean ridges the ridge-push forces act on the edges of the rifting plate. As a result of rising magma, the oceanic ridges are higher than the surrounding oceanic floor, and acting on an incline, forces the oceanic lithosphere away from the higher spreading ridges and toward the trenches. This force may be two or three times greater if a mantle plume underlies the ridge. Both slab-pull and ridge-push are gravity driven but still depend on thermal differences within the Earth. Although slab-pull cannot initiate subduction, once a slab begins to sink, the slab-pull force rapidly becomes the dominant force for continuing subduction. Thus the plate moves like a conveyor belt, where the push is created at a ridge, while the pull occurs at the trench boundary (Forsyth and Uyeda, 1975; Kearey and Vine, 1990; Condie, 2001).

Along with the slab-pull mechanism, another force arises from the drag of a subducting slab on the adjacent mantle. The result is an induced mantle circulation that pulls both subducting and overriding plates toward the trench. Because mantle flow tends to draw in nearby plates it is called the slab-suction force (Forsyth and Uyeda, 1975; Kearey and Vine, 1990).

2.4. Mantle plume and continental rifting

A mantle plume is a narrow cylindrical thermal diapir of low-density material that originates deep in the mantle, either from the mantle–core boundary (at a depth of about 2900 km), or from the 670 km discontinuity at the base of the upper mantle. A mantle plume may be stable for several hundred million years (Morgan, 1981; Davies, 1999). Some plumes may act as fixed reference frames for plate motion. A second type of plume, also originating from the core–mantle boundary, can be bent and moved relative to the global circulation of the mantle. Courtillot et al. (2003) identified three kinds of plume, namely: 1) primary or deep plumes, originating from the core–mantle boundary; 2) secondary plumes, originating from the top of the large domes of deep plumes; and 3) tertiary shallow plumes, originating near the 670 km discontinuity and linked to tensile stresses in the lithosphere.

Continental rifting and flood basalts are linked to upwellings of deep mantle plumes. Mantle plumes impinge on the base of the Earth's lithosphere in all plate tectonic settings, thus producing domal uplift of 500–1000 m, lithospheric thinning, extensional stress fields, and a thermal anomaly centered on plume (Condie, 2001). The surface uplift occurs a few million years before the eruption of flood basalt volcanism. Heating the base of the lithosphere by mantle plumes may lead to partial melting and the formation of mafic magma. When a rising plume head arrives at the Earth's surface, massive volcanic eruptions spew out voluminous lavas that may promote the breakup of supercontinents and often result in the formation of volcanic margins (Fig. 2A).

Morgan (1981) championed the idea that mantle plumes may play an important role in the breakup of continents. However, some plumes do not result in continental breakup, as in the case of Columbia River flood basalt or the Siberian Traps. When continents rift to form new ocean basins, the rifting is sometimes accompanied by massive igneous activity. Along many continental rift margins, huge areas of flood basalts are often extruded onto the surface at the same time, as the continents break apart. Many researchers have argued that there is a close temporal and the causal link between mantle plumes, flood basalts and continental breakup (Morgan, 1981; Storey, 1995; Courtilot et al., 1999; Tackley, 2000; Segev, 2002).

Magma from mantle plumes may be extruded as lava flows on the Earth's surface to form continental flood basalts that represent the largest outpourings of lava flows in the Earth's history. Flood basalts are often referred to as "large igneous provinces" or "LIPs," which are known for marked thinning of lithosphere, often ending with a continental breakup (Coffin and Eldholm, 1994). These large eruptions create a series of separate overlapping lava flows that give many exposures a terrace-like appearance, called traps. A characteristic feature of flood basalt volcanism is that it takes place quickly, often in less than a million years. Eruptions of flood basalts are often associated with rifting, but establishing the relative timing of the two phenomena is often difficult.

Subaerial flood basalt rocks (~3 to 6 km thick) that usually characterize the volcanic rifted margins are found on opposite shores of India. These rifted margins formed when India began to break away from Gondwana.

2.5. Active and passive continental rifting

The rifting of the continents and evolution of ocean basins are fundamental components of plate tectonics, yet aspects of the process of continental breakup remain controversial. The rigid continental lithosphere comprising the crust and upper mantle is about 200–300 km thick. By comparison, the oceanic lithosphere is only 100 km thick. It is interesting to note that though the continental lithosphere is thicker than the oceanic lithosphere, it is considerably weaker. This is due to the fact that the continental lithosphere has a lower average melting temperature than the oceanic lithosphere. Therefore, at 100 km, the continental lithosphere is closer to its melting temperature. Also the thick continental keel is very weak because of the great temperatures and pressures at depth.

During the breakup of a continental tectonic plate, the lithosphere thins, weakens, ruptures, and eventually is replaced by a new oceanic lithosphere. There is a debate about the cause and consequences of the rifting. Some models suggest that lithospheric extension and magmatism occur in response to subduction (Storey et al., 1995), plate tectonic reorganization (Anderson, 1994, 2001), convective partial melting (Mutter et al., 1988), or the mantle plumes (Morgan, 1981; White and McKenzie, 1989; Courtilot et al., 1999).

Clearly, for rifting to take place, there has to be extension. The continental plate has to be stretched and thinned, allowing the buoyant mantle material to rise toward the surface and occupy the newly created space. Does extension come first, initiating volcanism? or is it a thermal anomaly in the mantle that promotes both volcanism and extension? Some active rifts are sites of major underlying plumes as evident by extrusion of flood basalts associated with considerable extension; others are not. There are non-volcanic or passive rifts with moderate extension, initiated by plate forces acting over large distances.

There are two mechanisms for breaking up continental lithosphere: active rifting and passive rifting (Sengor and Burke, 1978; Turacotte and Emerman, 1983). In the active rifting model (Richards

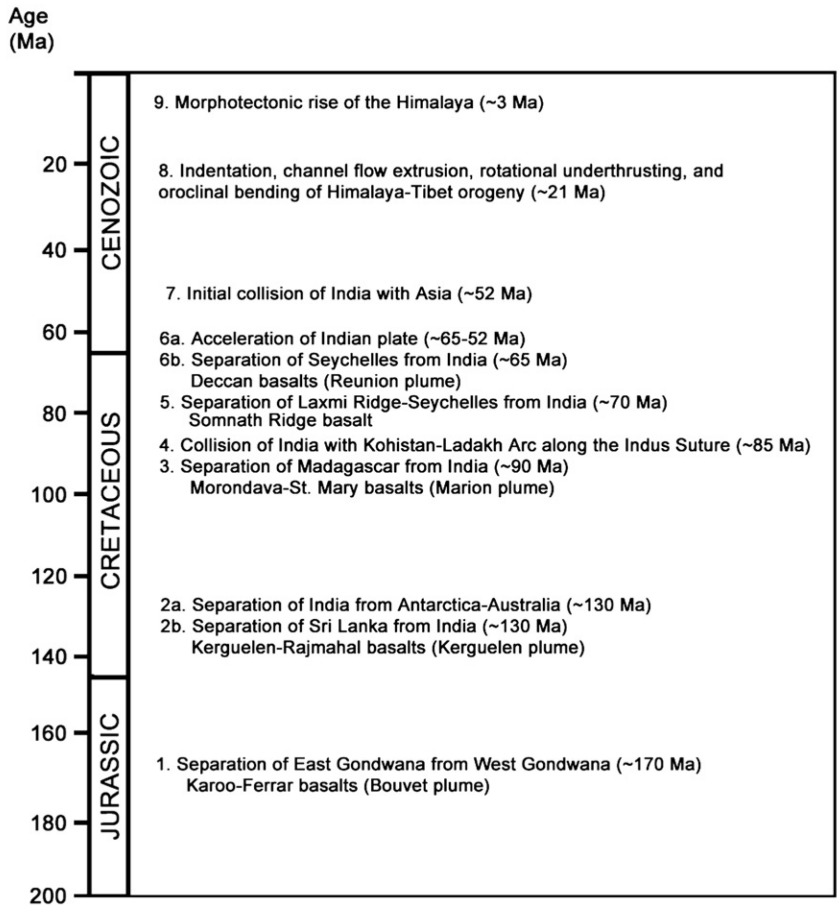

Fig. 1. Evolution of the Indian plate during the Late Mesozoic and Cenozoic periods associated with plate tectonic and magmatic events showing 9 major stages.

et al., 1989), the buoyant material of a hot plume, which probably originated at the core–mantle boundary, rises through the mantle like a thermal diapir and impinges on the base of the continental lithosphere, where it spreads laterally against the barrier. The plume undergoes widespread decompression melting to form large volumes of basaltic magma and forces the lithosphere to dome upwards (Fig. 2A). This explains why rifts often occupy broad topographical rises, like the Rio Grande and Baikal Rifts.

The arriving head of the plume, which may be as large as ~1000 km in diameter, flattens to a disk at the base of the lithosphere, causing extension. It thins, weakens, and cracks the lithosphere. The head then melts and erupts rapidly onto the surface as continental flood basalts over an area of 2000–2500 km in diameter. The sequence of events during the development of an active rift is doming, followed by flood basalt volcanism, and then finally continental rifting (Campbell and Griffiths, 1990; Condie, 2001).

In the passive rifting model (White and McKenzie, 1989), plate tectonic forces stretch, thin, and ultimately rupture the lithosphere, thereby allowing the underlying mantle to rise by decompression melting of the asthenosphere, triggering volcanism. In this model, rifting is the controlling force and is driven by extensional processes where the lithosphere is stretched laterally rather than vertically (Fig. 2B). Passive rifting precedes the associated volcanism. Here continental breakup is commonly preceded by a series of extensional episodes, which have an effect on the subsequent volcanic nature of the rifted margin. The lithosphere is stretched, thinned, and finally ruptured because of regional stresses related to plate boundary forces and geometry. The ultimate force of lithosphere thinning and stretching comes from the plate tectonic forces such as slab pull. In passive rifting, the rifting event precedes the volcanism.

The relative timing of rifting and rift-related volcanism is generally used to discriminate between these two basic types of rifting (Sengor and Burke, 1978). In active rifting, flood basalt volcanism predates breakup event. There may be a time lag of 10–20 Myr between initial phase of volcanism and subsequent continental breakup as revealed by the oldest ocean magnetic anomalies (Hill, 1999). In the passive rifting model, rifting events predate volcanism. The passive mode of rifting appears to be more common in the geological record than the active rifting.

The passive and active rifting mechanisms have been recently polarized into two opposing camps: plate vs. plume paradigms (Foulger, 2010). The plate hypothesis endorses exclusively the passive rifting model and is thus conceptually inverse of the plume hypothesis. It attributes volcanism to shallow, near-surface processes associated with plate tectonics, rather than active processes arising at the core–mantle boundary (Anderson, 2001).

The resolution of the debate lies in the recognition that "passive rifting" versus "active rifting" is a false dichotomy. The answer is neither "passive rifting", nor "active rifting", but rather a combination of both. There will always be mantle plumes and hot spots, but their geographic location and timing is largely independent of plate motions. Similarly, there will be times when plate motions cause stresses which cause rifting and extension of the continental lithosphere. It is clear that when continental lithosphere is put under extension it will break along pre-existing zones of weakness. The most conspicuous zones of weakness are: 1) old continental collision zones where the lithosphere is still relatively warm, and 2) areas recently weakened by mantle plumes and hot spots. Plumes and hotspots are associated with continental rift margins because they represent fundamental zones of weakness in the lithosphere. The thermal uplift associated

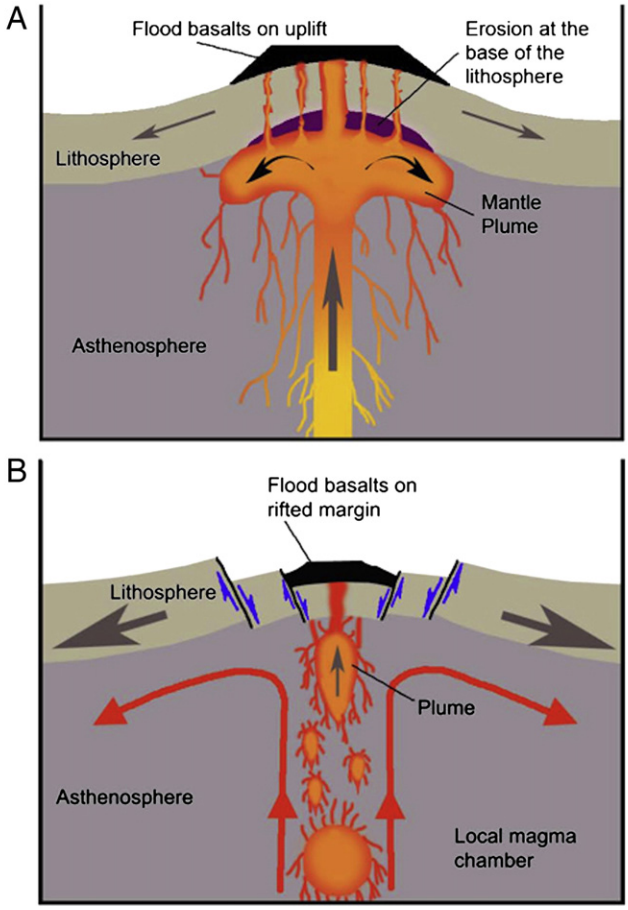

Fig. 2. Two possible mechanisms for continental breakup. (A) In the active rifting model, uprising plume causes doming of the lithosphere that triggers flood basalt volcanism and subsequent rifting; flood basalt predates the rifting event. (B) In the passive rifting model, plate tectonic forces stretch, thin, and ultimately rupture the lithosphere, that could lead to partial melting of the underlying asthenosphere to rise and trigger volcanism; rifting precedes the flood basalt volcanism. In our view, the "passive rifting" versus "active rifting" debate appears to be a false dichotomy. It is clear that if the continental lithosphere is put under extension it will break along pre-existing zones of weaknesses. Plumes and hotspots are associated with continental rift margins because they represent fundamental zones of weakness in the lithosphere.

Fig. 3. Distribution of Mesozoic flood basalts and their ages implicated for the sequential rifting of the Indian plate from Gondwana; locations of flood basalts are shown in a Late Triassic reconstruction of the Gondwana map (~220 Ma).

with a plume also adds a deviatoric force, similar to ridge push, that helps the rift propagate in the direction of least stress. For these reasons, it is clear that though plumes are associated with rifting events, they do not cause rifting, but rather mantle plumes and hot spots promote and accelerate the rifting process driven by plate tectonic forces.

2.6. Mesozoic Gondwana flood basalts

Mesozoic continental flood basalts dominate the landscapes of the Gondwana continents. Flood basalts are found on opposite sides of South Atlantic in Brazil and South Africa (Storey, 1995; Courtilot et al., 1999). Similar volcanic rift margins can be seen on opposite sides of the Indian Ocean, between South Africa and Antarctica, between East India and Australia, and between West India and Madagascar. On the two sides of the Indian peninsula, the Rajmahal and Deccan traps covered a significant fraction of India's land surface (Figs. 3, 5). In both cases the eruption of continental flood basalts occurred at the same time as rifting along the margins of the Indian plate. Therefore, India is a unique natural laboratory for investigating rifting mechanisms associated with plumes and plates. The breakup of East Gondwana from West Gondwana, beginning some 167 Ma as revealed from ocean floor anomalies, followed by repeated rifting of India provides a crucial framework for testing the timing and relative roles of plumes vs. intraplate stresses for continental breakup. The breakup was complete by 65 Ma when India separated from the Seychelles and migrated northwards on a collision course with Asia.

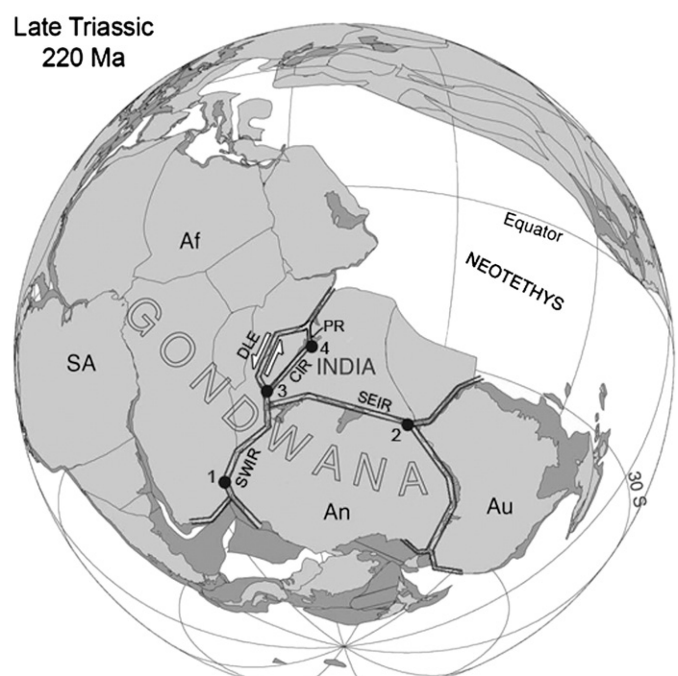

Fig. 4. Paleogeographic reconstruction of Gondwana during the Late Triassic (~220 Ma) showing the future locations of mantle plumes and the ages of breakup of eastern Gondwana. 1, Bouvet plume (~180 Ma); 2, Kerguelen plume (~118 Ma); 3, Marion plume (~88 Ma); and 4, Reunion plume (~65 Ma). Abbreviations: Af, Africa; An, Antarctica; Au, Australia; CIR, Central Indian Ridge; DLE, Davie and Lebombo–Explora transforms; PR, Palitana Ridge; Sa, South America; SEIR, South East Indian Ridge, and SWIR, Southwest Indian Ridge.

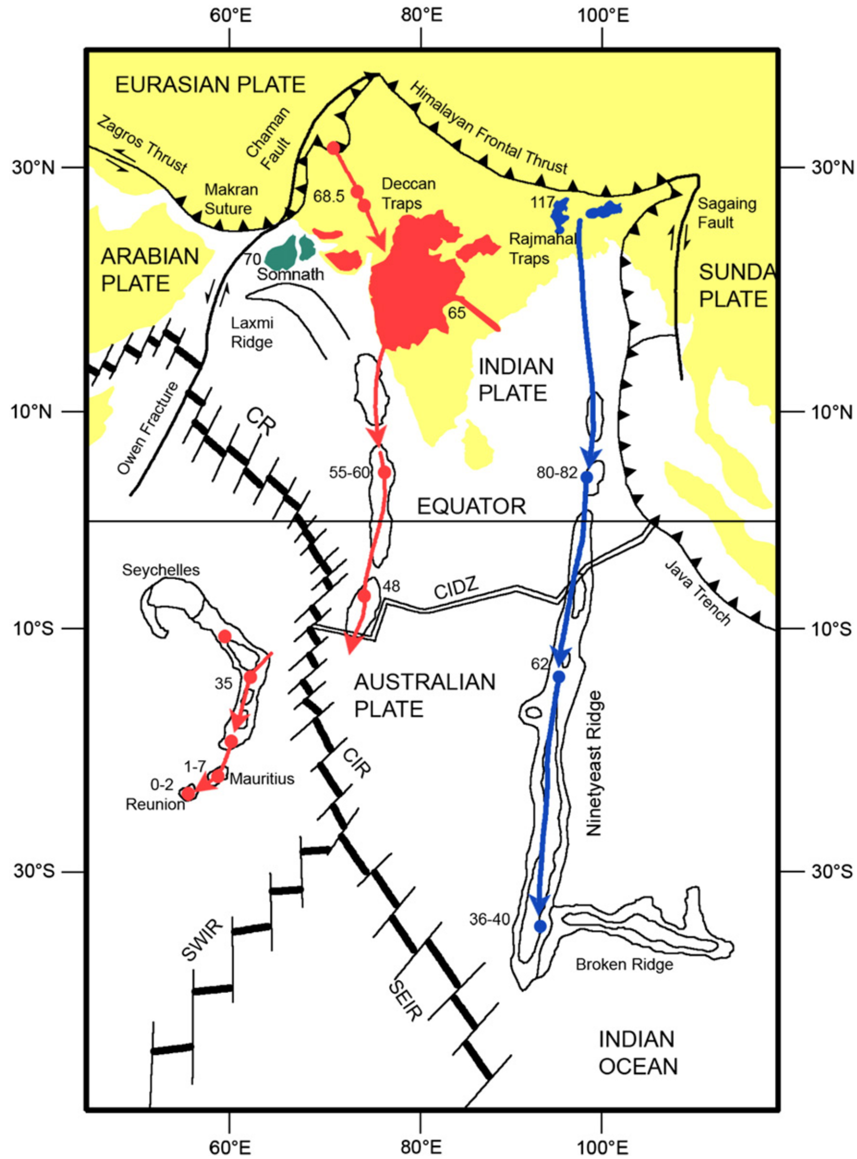

Fig. 5. Major tectonic and magmatic features of the Indian Ocean that provide clues to the tectonic evolution of the Indian plate. Abbreviations: CIDG, Central Indian Deformation Zone; CIR, Central Indian Ridge; CR, Carlsberg Ridge; SEIR, South East Indian Ridge; and SWIR, Southwest Indian Ridge. Compiled from McKenzie and Sclater (1973), Powell (1979), Schlich (1982), Chatterjee (1992), Royer and Coffin (1992), and Bouysse et al. (2004). Two spectacular hotspot trails, the Rajmahal–Kerguelen trail on the east side of the Indian plate (blue line) and the Deccan–Reunion trail on the west side of the Indian plate (red line) are shown. West of the Deccan Trap exposure, the offshore Somnath volcanic platform (green) is shown north of the Laxmi Ridge that erupted about 5 million years earlier than the Deccan.

3. Chronology of the breakup of the Indian plate from Gondwana and its collisions

3.1. Sequential breakup of India from Gondwana and associated flood basalts

During the Mesozoic, there were enormous volcanic outpourings of continental flood basalts in Gondwana. The continental breakup and dispersal of Gondwana is marked by a series of mantle plumes or hotspots (Storey, 1995). As Gondwana began to break up into separate continents during the Jurassic and Cretaceous periods, extensive intra-continental rifting took place in association with catastrophic flood basalt events.

The role of India in the dispersal of Gondwana is well constrained by linear magnetic anomalies from the ocean floor and numerous paleomagnetic poles (Seton et al., 2012). The distribution of Mesozoic volcanism in the Indian plate is the surface expression of the underlying mantle plume activity. Both the eastern and western coasts of India are volcanic rifted margins, characterized by subaerial volcanic rocks.

The two major hotspots in the Indian Ocean, the Réunion and the Kerguelen hotspots, have left behind an unusually complete record of mantle plume activity (Duncan, 1981). The present day configuration of coastal margins of the Indian peninsula is the consequence of five episodes of continental flood volcanism and sequential rifting events since the early Jurassic time. In the following section we review the association of flood basalts, mantle plumes, and rifting events of the Indian plate through time and space and reconstruct the thermal histories of the evolving Indian margins.

The breakup and dispersal of the Indian plate from Gondwana illustrates the varied temporal and spatial relationships that exist between continental rifting and magmatism. The great event that triggered the breakup of the supercontinent Gondwana into Africa, Antarctica, Australia, and India about 167 Ma, and consequently the opening of the Indian Ocean, is thought to have been plate tectonic events related to subduction of the Tethyan ocean floor beneath the southern margin of Eurasia (Scotese, 1991).

Following the breakup of Gondwana during the Early Cretaceous, the Indian craton was subject to several major rifting events. The Bouvet, Marion, Kerguelen, and Réunion plumes all would leave their marks on the Indian plate (Fig. 6). Plume activities continued for 115 million years as India became smaller and smaller, trimming its rifted continental margins, and shedding several smaller continental blocks such as Sri Lanka, Madagascar, Seychelles, and Laxmi Ridge. Madagascar and Seychelles were stranded in oceanic crust, whereas Sri Lanka and Laxmi Ridge were accreted and traveled with India during its long northward journey.

We have identified five time-progressive flood basalt volcanic episodes that affected the Indian peninsula (Fig. 2). These include: (1) Karoo–Ferrar basalts (Bouvet plume, ~182 Ma); (2) Kerguelen–Rajmahal basalts (Kerguelen plume, ~118 Ma); (3) Morondava–St. Mary basalts (Marion plume, ~88 Ma); (4) Somnath Ridge basalt (~70 Ma), and (5) Deccan–Reunion basalts (Reunion plume, ~65 Ma) (Figs. 3, 4). We believe mantle plumes and hotspots do not break apart continents. They just make the job easier. The prime reason for continental rifting is stretching and thinning of the lithosphere by slab pull forces that can lead to partial melting of hot, ductile rock of the mantle, which wells up and erupts as spectacular flood basalt volcanism (White and McKenzie, 1989). Thus plate forces combine opportunistically with pre-existing mantle plumes to produce massive, flood basalt volcanism.

3.2. Collisions during India's northward journey

Other than rifting, the Indian plate shows two successive collision events that shaped its northern margin: (1) collision of India with the Kohistan–Ladakh Arc during Late Cretaceous (~85 Ma) along the Indus Suture; and (2) collision of India with Asia during Early Eocene (~50 Ma) along the Shyok–Tsangpo Suture, followed by post-collisional shift to Indus–Tsangpo Suture. The India–Asia collision led to the rise of the Himalayan mountain range and the uplift of the Tibetan Plateau.

4. Material and methods

We have used the pre-breakup configuration of Gondwana during the Late Triassic (~220 Ma) as the starting point to trace the tectonic evolution of India in space and time (Chatterjee and Scotese, 1999, 2010) (Fig. 3). We show the probable extent of the "Greater India" subcontinent, which rifted away from the western margin of Australia during the Early Cretaceous (Powell et al., 1988; Ali and Aitchison, 2005). As discussed by Powell and Conaghan (1973), Greater India was subducted beneath Asia during the initial phases of collision, and now lies buried beneath the Tibetan Plateau.

The dating of the breakup of Gondwana and separation of its fragments can be approached several ways. Linear magnetic anomalies and deep sea drilling data provide fairly accurate information about the continental breakup, but there are no anomalies during the long Late Cretaceous Quite interval (83–119 Ma). Information from different fields of continental geology and paleomagnetism also provide additional evidence. For example, the eruption of melt-dating flood basalts along the continental rift margins may help to constrain the timing of rifting. The plate tectonic model described in this paper is based on the on-going work of the PALEOMAP Project (Scotese, 2011a, 2011b, 2011c, 2011d, 2011e, 2011f) and can be best understood by reviewing the computer animation http://www.scotese.com or the iPad application "Ancient Earth" (Moore and Scotese, 2012).

For dynamic climate modeling to very long time scales we have used the Fast Ocean–Atmosphere Model (FOAM) (Scotese et al., 2007, 2008, 2009, 2011; Goswami, 2011).

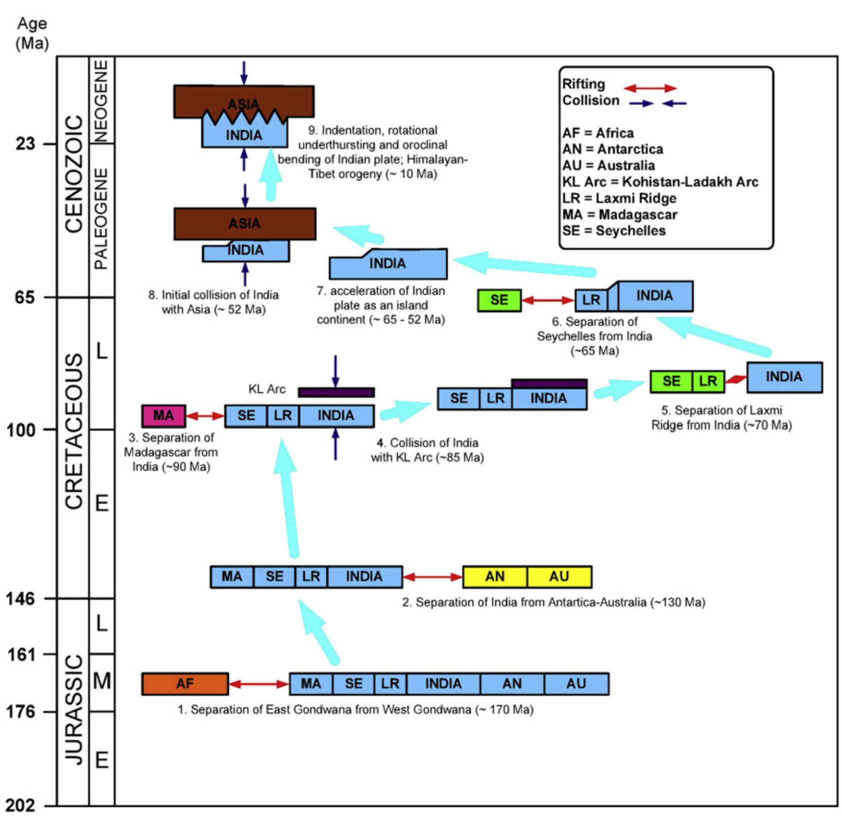

Fig. 6. Summary of nine major tectonic evolutionary stages of the Indian plate during its 9000 km-voyage from Gondwana to Asia (see Fig. 1 for explanation). Rectangles represent diagrammatic cross-sectional views of plates and microplates, which originally comprised Gondwana. They show sequential rifting of the Indian plate from Gondwana and its subsequent collisions with the Kohistan–Ladakh Arc (KL Arc) and Asia.

5. Evolution of the Indian plate during the Jurassic Period

5.1. Tectonic setting of the Indian Ocean

The seafloor of the Indian Ocean is dominated morphologically by a system of active midocean ridges: the Carlsberg Ridge (CR), Central Indian Ridge (CIR), Southwest Indian Ridge (SWIR), and Southeast Indian Ridge (SEIR) (McKenzie and Sclater, 1973). The spreading rates of these ridges, corresponding to the quantity of magma supplied at their axes during a same period of time, are very different (Bouysse et al., 2004). These spreading ridges converge midocean at the Rodriguez Triple Junction located at 70°E and 26°S, which resembles an inverted Y. The ridges form the boundaries of three plates: the Nubia/Somalia (or Africa) plate to the west, the India–Australia plate to the east, and the Antarctica plate to the south. One arm of the inverted Y, the Southwest Indian Ridge (SWIR) separates Africa and Antarctica and connects with the Mid-Atlantic Ridge. It is characterized by a very slow spreading rate, less than 2 cm/year. The other arm, the Southeast Indian Ridge (SEIR) separates the Indo-Australian plate from Antarctica and joins with the East Pacific Rise. It is the fastest of the three ridges, with a spreading rate of 7 cm/year. North of the triple junction, the Central Indian Ridge (CIR) runs almost due north as a series en echelon spreading centers and fractures zones before turning to the northwest as the Carlsberg Ridge (CR). Its spreading rate is 3 cm/year. The Owen Fracture Zone offsets the northern end of the Carlsberg Ridge and the Sheba Ridge about 300 km. The Sheba Ridge is the plate boundary between India and Arabia and extends for hundreds of kilometers in a north-northeast—south-southeast direction and connects with Chaman strike-slip fault along the Baluchistan Arc (Fig. 5).

Since the closure of the Neotethys Ocean around 50 Ma, India and Australia moved as a single plate (Scotese et al., 1988). Around 8 Ma, the Indian and Australian plates began to break into two plates due primarily to stresses induced by the collision of the “Indo-Australian plate” with Asia (Van Orman et al., 1995). The Indo-Australian plate —long identified as the single plate on which both India and Australia lie—appears to be divided by a diffuse plate boundary, the Central Indian Deformation Zone (CIDZ).

The modern Indian Ocean has the fewest trenches of any of the world's oceans. The narrow (~80 km) volcanic, and seismically active Java trench (~6000 km long) runs from southwest Java and continues northward as the Sunda trench along the southern rim of the Sunda Island Arc (Indonesia) and the Andaman and Nicobar Islands. A smaller subduction zone, the Makran Trench (~900 km long) is located south of the shores of Baluchistan (Bouysse et al., 2004).

There are three important areas of mantle plume activity in the Indian Ocean, namely, Kerguelen, Marion, and Reunion hotspots. The breakup of Gondwana and dispersal of several of its component continental fragments—Africa, India, Australia, and Antarctica—created the Indian Ocean (Reeves and de Wit, 2000). In most cases, extensive flood basalts accompanied these rifting events. Two linear, aseismic submarine ridges, the Ninety East Ridge and the Chagos–Maldive–Laccadive Ridge, lie on either sides of the Indian plate (Fig. 5). These islands and seamounts are traces of Kerguelen and Reunion hotspots, and were formed during the rapid northward movement of India during Cretaceous and Tertiary time (Duncan and Pyle, 1988).

The Indian Ocean had begun to open by the time of magnetic anomaly M25 (~156, Late Jurassic), which is the oldest anomaly between Africa and East Gondwana. It has experienced, along with three main phases of seafloor spreading, two major plate reorganizations from the Late Jurassic to Present. The first phase of spreading started with India's movement away from Antarctic–Australia during the Early Cretaceous. This separation created the Early Cretaceous oceanic crust with Mesozoic anomaly sequence M11 through M0. During the Late Cretaceous (118–84 Ma), the Indian plate moved northward across the Indothethys to collide with the Kohistan–Ladakh Arc (Khan et al., 2009).

5.2. Separation of East Gondwana from West Gondwana (~167 Ma)

The configuration of the Indian plate changed dramatically during the Late Jurassic (Figs. 4, 7). During the Late Jurassic (~167 Ma), Gondwana began to rift apart into two roughly equal, smaller continents separated by the Southwest Indian Ridge (SWIR) (between South Africa and Antarctica) and the Davie and Lebombo–Explora (DLE) transforms (between North Africa and Madagascar–India), producing a narrow seaway between West and East Gondwana. This is the first rifting event in Gondwana along the SWIR-DR spreading center separating Africa from India–Antarctica (Reeves and de Wit, 2000) (Fig. 7).

West Gondwana consisted of Africa and South America, and East Gondwana composed of Madagascar, India, Australia, and Antarctica (Coffin and Rabinowicz, 1987). The oldest ocean floor separating the western Gondwana from eastern Gondwana is Late Jurassic, approximately 167 Ma (Lawver et al., 1991) (stage 1, Fig. 6). The first oceanic crust appeared in the Late Jurassic as the Africa–South America plate moved northward relative to the India–Sri Lanka–Seychelles–Madagascar–Australia–Antarctica plate. The initial phase of rifting was signaled by the eruption of widespread flood basalts in Gondwana, such as Karoo Group in South Africa, Ferrar Group in Antarctica, and Tasman Group in Australia (Duncan et al., 1997; Fig. 2A).

6. Indian plate motion and climate change

A meteorologist by profession, Alfred Wegener was particularly interested in ancient climates. One impressive line of evidence presented by Wegener (1915) in support of his continental drift theory is the past distribution of climatic indicators. During the Late Carboniferous Period, a continental ice sheet covered parts of South America, southern Africa, India, southern Australia, and Antarctica. Such a huge ice sheet could only mean that southern continents were joined together in cold latitudes surrounding the South Pole. With the advent of plate tectonics, it became clear that the Earth's climate has varied throughout geological time as a result of the movement of the continents through the Earth's climatic belts, as well as global climatic change.

Fig. 7. Paleogeographic reconstruction showing the initial breakup of East Gondwana from West Gondwana around ~160 Ma and the location of Bouvet plume. Abbreviations: DLE, Davie and Lebombo–Explora transforms.

Plate motions reconfigure continents and oceans over the course of million years, shift latitudinal position, and generate new topography and relief, which can affect both local and global patterns of climate and atmosphere–ocean circulation. The horizontal and vertical displacements associated with plate tectonics play a fundamental role in climate change over a wide range of time scales.

Since India drifted about 9000 km from its Gondwana home in Southern Hemisphere to Asia in Northern Hemisphere, it has experienced different types of climatic regimes during its northern journey. As tectonic plates move, so do the subaerial continents, which control the geometry of the oceans. These changes of the configurations of lands and seas have an important effect on the transfer of heat and moisture across the globe, and therefore, in determining global climate.

Tectonic uplift is often cited as an important contributor to long-term climate change, and the Himalayan–Tibetan orogen has been implicated as primary contributors to the onset of monsoon, and Cenozoic cooling and Northern Hemispheric glaciation. The flight of the Indian plate from southern tropical zone to northern tropical zone via the equatorial belt and the orographic barrier of the Himalayan–Tibetan Plateau provides a unique model to study plate motion and its effect on climate changes.

We have used FOAM to better understand the paleoclimatic changes of the Indian plate during its long latitudinal trek from its Gondwana abode to its final destination in Asia (Scotese et al., 2007, 2008, 2009, 2011; Goswami, 2011).

6.1. Jurassic climate

During Early Jurassic (~167 Ma) as the Pangea started to breakup, the Indian subcontinent was situated in the southern Subtropical Arid Belt (30°S). Consequently, the climatic conditions were very warm and dry across all of India. As estimated by the Fast Ocean Atmospheric Model (FOAM), except for the distant northwestern province of the Indian subcontinent the overall mean annual precipitation was less than 2 cm/month (Fig. 8). The eastern part of the peninsular India, as well as the eastern part of India (including Bangladesh and part of Burma), were relatively cooler with annual mean temperature of about 7.5 °C–10 °C. The temperature gradually increased from eastern to western parts of India and ranged between 10 °C and 17 °C as estimated by FOAM (Scotese et al., 2007, 2008, 2009, 2011; Goswami, 2011).

The Late Jurassic (~160 Ma) climate of India was little different than the Early Jurassic (~180 Ma). As part of East Gondwana, the center of the Indian subcontinent was located near 32°S and 32°E. The overall precipitation was still less, and the western part of the then peninsular India was very dry. As FOAM estimates (Scotese et al., 2007,

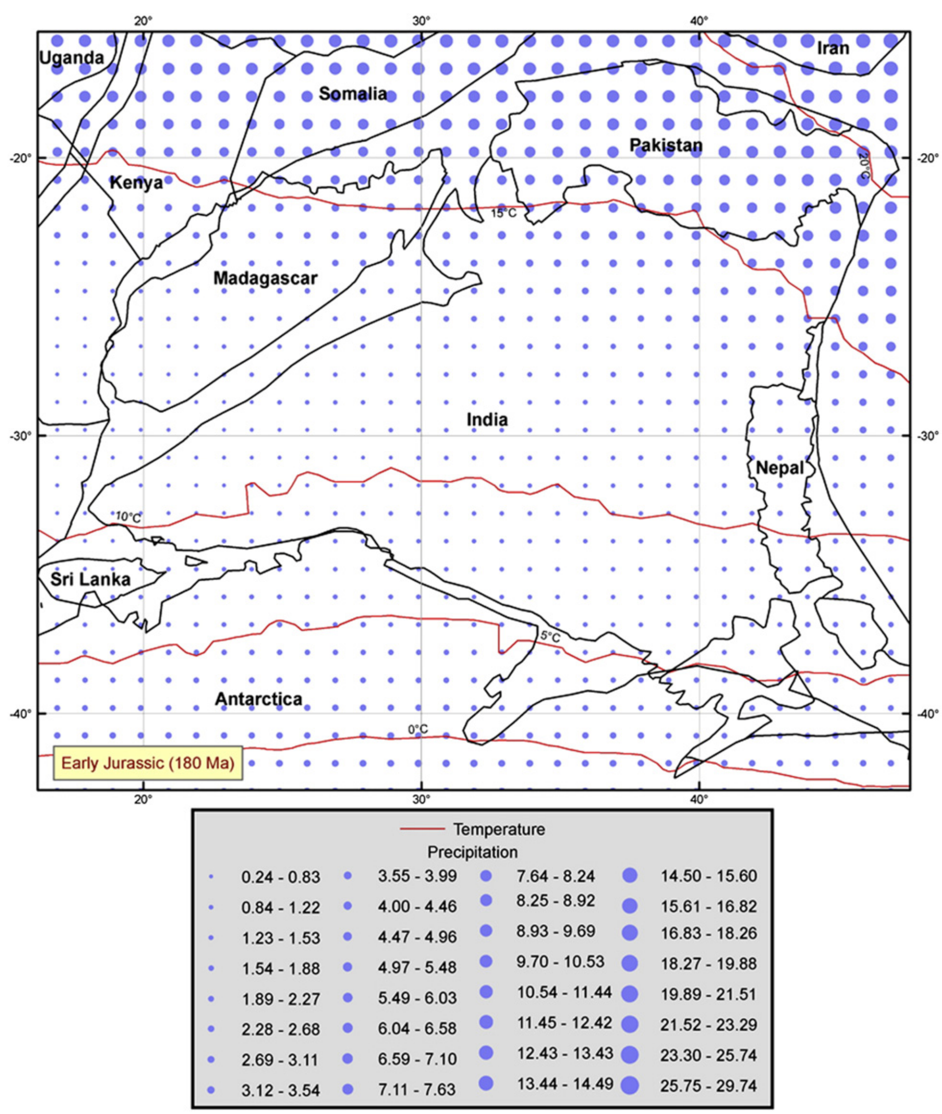

Fig. 8. Mean annual precipitation and mean annual temperature maps of the Indian plate during Early Jurassic (~180 Ma) time.

2008, 2009, 2011; Goswami, 2011), the mean annual precipitation gradually increased to the northwestern and northeastern direction and ranged between 3 cm/month and 7 cm/month. The mean annual temperature for the peninsular and eastern India was around 10 °C–15 °C and the northwestern part was even warmer (16.5 °C–25.0 °C) (Fig. 8).

7. Evolution of the Indian plate during the Cretaceous Period

7.1. Separation of India from Antarctica–Australia (~130 Ma)

During the second phase of Gondwana breakup history in Early Cretaceous time (~130 Ma), East and West Gondwana separated into two nearly equal halves. In East Gondwana, the conjoined Antarctica–Australia rifted from the smaller Sri Lanka–India–Laxmi Ridge–Seychelles–Madagascar (SL–I–LR–S–M) fragment, opening the Central Indian Ocean from northeast to southwest (stage 2, Fig. 6) by spreading along the SEIR (Fig. 9). The rifting created the modern coastline of eastern India. By the mid-Cretaceous, seafloor spreading in the Somali Basin had stopped, and India became fixed to the African plate. The uplift at the India–Antarctic boundary reached a maximum in the Early Cretaceous, when a variety of flood basalts were emplaced: Rajmahal–Sylhet basalts in eastern India, Bunbury basalts in Australia, and Kerguelen basalts in Kerguelen Island, all produced by the Kerguelen plume that commenced at 118 Ma and continued throughout the Cretaceous and Cenozoic (Kent et al., 1997). Today Rajmahal flood basalts in eastern India are exposed over an area of more than 4000 km^2 and occur in the subsurface beneath the Bengal Basin. All of these basaltic rocks were most likely contiguous with the Bengal and Sylhet traps, and therefore may have originally covered an area of at least 200,000 km^2 (Baksi et al., 1987). The Kerguelen plume left a 4500 km-long hotspot track along the Ninety East Ridge, Broken Ridge to the Kerguelen Archipelago in the eastern Indian Ocean beneath (Duncan, 1981; Morgan, 1981) (Fig. 5).

The timing of the separation of India from Antarctica/Australia is not well constrained. Anomaly M10 in the Perth and Cuvier Basin indicates that rifting between Australia and India began around 133 Ma

Fig. 9. Paleogeographic position of Gondwana showing the separation of India from Antarctica–Australia (~130 Ma) along the Southeast Indian Ridge and the location of the Kerguelen plume. For abbreviations, see Fig. 4.

(Johnson et al., 1980). Recently Mesozoic magnetic anomalies in the Enderby basin of Antarctic margin indicate that the India–Antarctica and India–Australia rifting events were nearly contemporaneous, ranging from 124 to130 Ma (Gaina et al., 2007). This correspondence between magnetic anomalies and flood basalt volcanism suggests that early breakup event between India and Antarctica occurred, at least 6 Myr prior to the surface expression at the Rajmahal Trap (Figs. 3, 5).

7.2. Limited separation of Sri Lanka from India (~130 Ma)

The similarity of the geology of Sri Lanka with that of South India has long been recognized. Katz (1978) used the pre-Cretaceous boundary fault along the eastern margin of South India and western margin of Sri Lanka to reassemble these two Gondwana fragments in the closing of the Gulf of Mannar and Palk Strait. He believed that the India–Sri Lanka rift might be contemporaneous with the separation of India from East Antarctica in the Early Cretaceous (~130 Ma). Recently Desa et al. (2006) identified the oldest Mesozoic magnetic anomaly as M11 (~134 Ma), thus supporting the idea that separation of Sri Lanka occurred simultaneously with the rifting of India from Antarctica. Apparently, it was a failed rift that was aborted around ~125 Ma. Since then, Sri Lanka has become an integral part of the Indian plate (SL–I–LR–S–M).

7.3. Rifting of Madagascar from India (~90 Ma)

In the third phase of Gondwana breakup, starting in Late Cretaceous time, Gondwana fragmented into four parts: Seychelles/Greater India and Africa/Madagascar broke apart, as did Australia and Antarctica. The Seychelles and Laxmi Ridge were originally "sandwiched" between Madagascar and India prior to the breakup of Gondwana. During the Late Cretaceous, the Sri Lanka–India–Laxmi Ridge–Seychelles (SL–I–LR–S) block rifted away from Madagascar (M), opening the Mascarene Basin (stage 3, Fig. 6). Madagascar drifted southward in relation to India. This right-lateral strike-slip motion created a long, relatively straight, rifted passive margin. As with the rifting of Antarctica–Australia from eastern India, the Madagascar rifting event was preceded by uplift, which reoriented the fluvial systems on the subcontinent (Gombos et al., 1995). This event was associated with significant flood basalt volcanism in Madagascar induced by the Marion plume underlying southern Madagascar during the Late Cretaceous (Albian and Turonian) around 93 Ma (Storey et al., 1995; Bardintzeff et al., 2010).

The Morondava flood basalts must have covered the entire surface of the Madagascar Island at the time of rifting from Greater India. Today, these volcanics occur along the east and west coasts of Madagascar (Fig. 3). The St. Mary flood basalts along the western coast of India that have yielded ages between 84 and 93 Ma (Storey et al., 1995; Torsvic et al., 1998; Bardintzeff et al., 2010) are linked to the initial breakup (Fig. 10). Seafloor spreading in the Mascarene basin started at about 88 Ma (White and McKenzie, 1989). Thus, there is a minor gap in time between the earlier flood basalt volcanism and later continental fragmentation.

One of the striking features of the fit between India and Madagascar is the linearity of the coastline of Madagascar (>1500 km). It seems likely that this linearity is due to strike-slip movements prior to rifting in the Mascarene basin. In this model, approximately 600 km of right-lateral strike-slip movement took place between India and Madagascar, as India, together with Madagascar, slid southwards away from Somalia (180 Ma–120 Ma).

7.4. Collision of India with the Kohistan–Ladakh (KL) Island Arc (~85 Ma)

The collision and accretion of numerous crustal fragments to the southern margin of Asia and northern margin of India preceded the India–Asia collision event as the Neotethyan Ocean contracted by a

Fig. 10. Paleogeographic reconstruction of Gondwana fragments showing the India–Madagascar rift (~88 Ma) and the location of the Marion plume. India was converging northward to the Makran–Indus Trench across the Indothethys. For abbreviations, see Fig. 4.

series of asymmetric subduction events. The Kohistan–Ladakh (KL) Arc represents an island arc around the Nanga Parbat syntaxis of the Himalaya that was accreted to the Indian plate during its northward journey (Fig. 11). The arc is juxtaposed to the north along the Shyok Suture with Asia, and against the Indian plate to the south along the Indus Suture. The KL Arc is a key area to reconstruct the various steps of convergence of the Indian and Asian continents since the Late Cretaceous. How and when the KL Arc was sandwiched between India and Asia is a controversial topic that needs to be resolved if we want to reconstruct the collision chronology of India with Asia.

So far, we have discussed the repeated rifting events at the divergent boundaries that separated the Indian plate from other Gondwana continents and the development of the three principal spreading ridges: the Southwest Indian Ridge, Southeast Indian Ridge, and the Central Indian Ridge. However, there is another aspect of the evolution of the Indian plate, where the fragments were welded to form a larger continent at convergent margins. The collision of exotic terranes and island arcs along convergent continental margins is a well-accepted process for continental growth. For example, Asia is a mosaic of small continents and terranes that have been sutured by collision to form a larger tectonic collage. Most of the collisional sutures in Asia appear to be older than 200 million years, therefore predating the breakup of Pangea (Allègre, 1988). Prior to India–Asia collision, several terranes were accreted to southern Asia, of which the southernmost is the Lhasa terrane, which is of prime importance for understanding the collision of India with Asia (Fig. 11). Most likely, the Lhasa block collided with the Qiangtang terrane of Tibet around Early Cretaceous (~110 Ma) and was welded with Asia along the Bangong–Nujiang Suture (BNS) (Allègre, 1988; Decelles et al., 2001). We see a similar mechanism of accretion of island arc at the leading margin of India during its northward trek.

In the fourth phase of the evolution of the Indian plate (see stage 4, Fig. 6), India broke off from Madagascar during the Late Cretaceous (~90 Ma) with the spreading of the Central Indian Ridge (CIR) and began to drift northward. Therefore, subduction necessarily began to take place between India and Asia to accommodate the northward journey of the Indian plate. Two north-dipping subduction zones existed between India and Asia, both active during the middle to Late Cretaceous times (~120 Ma to 90 Ma). The northern zone, the Shyok–Tsangpo Trench, is found north of Neotethys along the southern margin of Asia. The southern zone, the Makran–Indus Trench, fringes the Arabian margin of Africa, and continued eastward across the Neotethys (Allègre, 1988; Besse and Courtillot, 1988; Ali and Aitchison, 2008; Chatterjee and Scotese, 2010). This southern trench, the Oman–Makran–Indus Trench separated Neotethys into two segments: a northern Neotethys, and a southern, short-lived Indotethys (Fig. 10). As discussed later, the southern subduction zone along the Makran–Indus Trench has been revealed by tomographic imaging of the mantle under Himalaya and Tibet, leading support to its existence (van der Voo et al., 1999) (Fig. 12). Subduction of the Indian plate

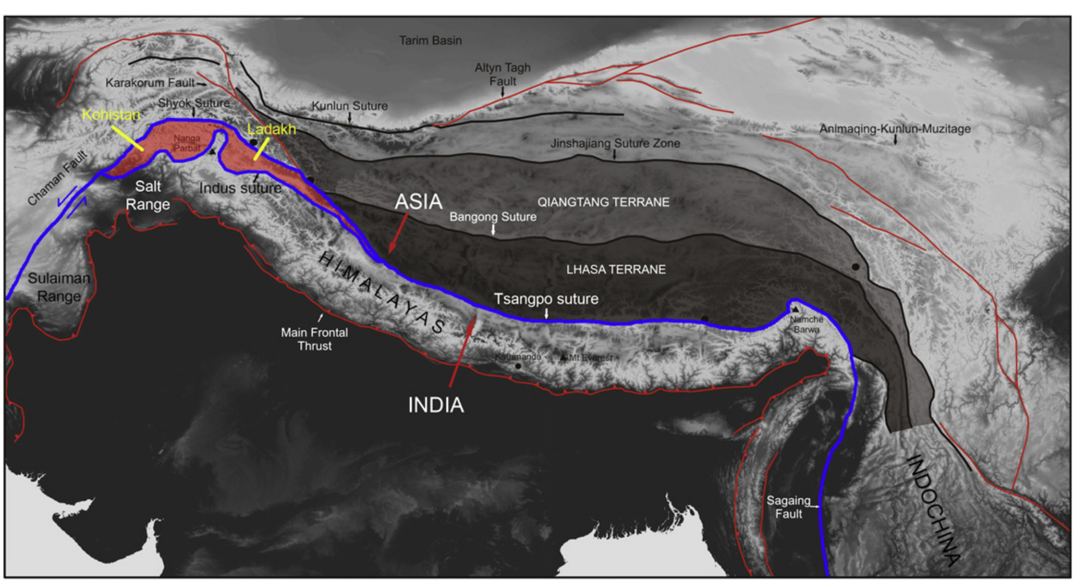

Fig. 11. Digital topography of the India–Asia collision zone showing three suture systems: Shyok, Indus, and Tsangpo. Tibet shows three accretionary microplates—Lhasa terrane, Qiangtang terrane, and Kunlun terrane. The initial collision site was the Shyok–Tsangpo Suture (blue line), which later transferred to the Indus–Tsangpo Suture Zone.
Shuttle Radar topography map from Andronicos et al. (2007) and other sources.

beneath the Indus Trench would give rise to the Kohistan–Ladakh (KL) Arc associated with spectacular obduction of deep oceanic crust onto the Indian subcontinent (Tahirkheli, 1979).

The KL Arc now lies between the Indian and Asian plates, and is bounded by the Shyok Suture (SS) to the north and by the Indus Suture (IS) to the south (Fig. 11). The timing of the collision of the KL Arc with India and with Asia, however, is controversial. Two hypotheses have been proposed for the origin of the KL Arc. Some authors have suggested that the KL Arc first collided with Asia around 95–75 Ma, well before its collision with India (Coward et al., 1987; Clift et al., 2002). In this model, Neotethys stretched in a north–south direction from India's northern (passive) margin to the Tethyan trench of the Asian margin in southern Tibet. It consisted of a single Neotethyan oceanic plate that lay between India and Asia (Powell and Conaghan, 1973).

However, most of the researchers suggest that the KL Arc first collided with India during the late Cretaceous (95–65 Ma) before India's terminal collision with Asia (Reuber, 1986; Allègre, 1988; Van der Voo et al., 1999; Ali and Aitchison, 2008; Jagoutz et al., 2009a, 2009b; Chatterjee and Scotese, 2010; Burg, 2011). In the early collision model of the KL Arc, the Tethyan Ocean between India and Asia is split into two oceanic plates by the Indus Suture: the northern Neotethys, and the newly identified southern, short-lived Indotethys. Both subduction zones, the Makran–Indus Trench and Shyok–Tsangpo Trench were active during the middle to Late Cretaceous (Besse and Courtillot, 1988) (Fig. 10).

Combined evidence from geochronology, paleomagnetism, tomographic imaging, and paleontology along the Makran–Indus Suture Zone suggests that India first collided with the Kohistan–Ladakh Arc during the Late Cretaceous. The Kohistan–Ladakh Arc represents a complete section of an oceanic arc, with rocks from mantle to upper crustal volcanic and sedimentary levels exposed (Burg, 2011). Sedimentary sequences (Yasin group) of Aptian–Albian age (<120–99 Ma) indicate that formation of this arc began in the Early Cretaceous (Pudsey, 1986). The bulk of igneous framework of the arc formed in the Late Cretaceous (90–100 Ma) (Petterson and Windley, 1985). Recent paleomagnetic data suggests that in Late Cretaceous time, the KL Arc occupied a position close to the Equator not far from the Indian plate

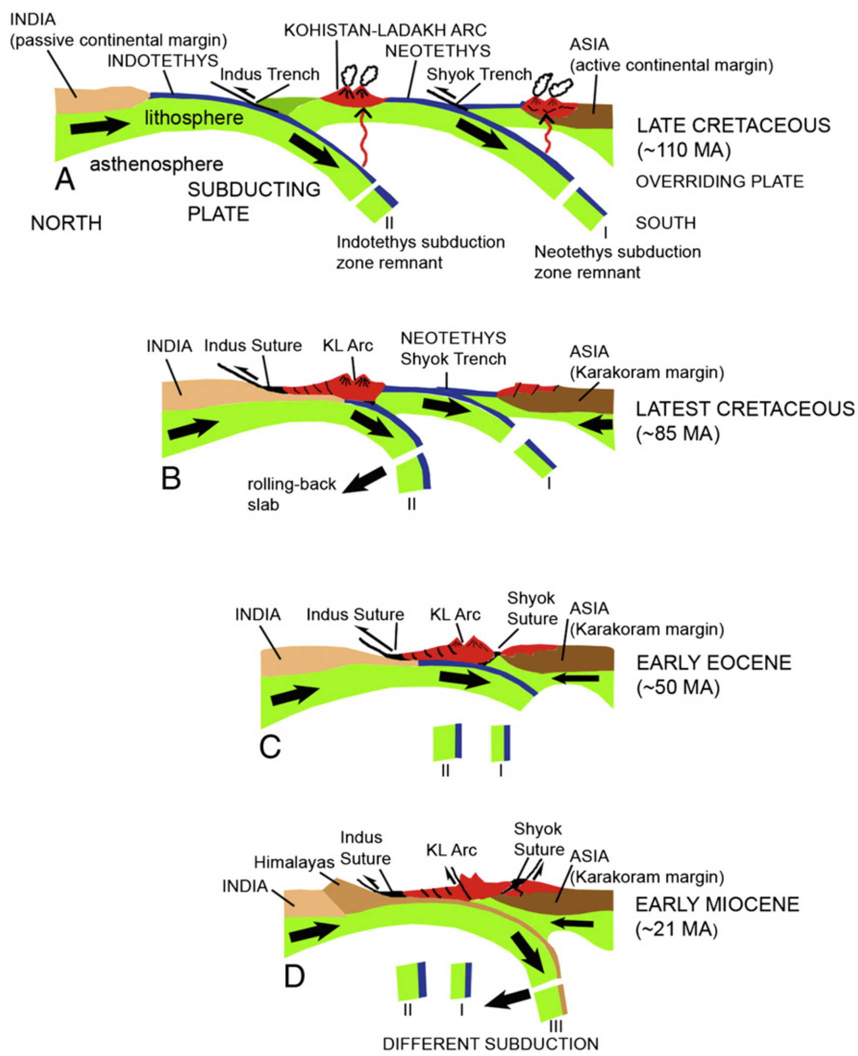

Fig. 12. Cartoon showing continent–arc collision in cross-section and possible sequences of plate tectonic events in the evolution of the Kohistan–Ladakh arc (A–D) based on tomographic imaging (Van der Voo et al., 1999). (A) During the Late Cretaceous the tectonic setting of the Kohistan–Ladakh arc lay between two north-dipping subduction zones when Neotethys comprises two oceanic plates, Indotethys and Neotethys. (B) Around ~85 Ma, India collided with the Kohistan–Ladakh arc with the disappearance of Indotethys. (C) India moved northward with accreted KL Arc in its leading edge towards Asia during the Late Cretaceous–Paleocene time at a superspeed of 15 cm/year. In the Early Eocene (~50 Ma), India collided with Asia with the disappearance of Neotethys and the speed was slowed down to 5 cm/year. (D) Subduction of India under Asia caused about 2500 km crustal shortening with the rise of the Himalaya and Tibetan plateau. Tethyan subduction zone remnants from the Neotethys and Indothethys (I and II) were revealed from tomographic imaging of the mantle.
Panel A–D was modified from Van der Voo et al. (1999), Allègre (1988) and Burg (2011).

(30°S). On the other hand, the Lhasa block was more than 3000 km north (20°N) (Khan et al., 2009). This geographic proximity of the KL Arc with India clearly supports that the arc was sutured to India first (~85 Ma), and then welded to Asia later (~50 Ma). This early collision model of the KL Arc with the India is consistent with tomographic imaging of the mantle (down to ~2000 km) under the India–Tibet collision zone, which identified the obducted oceanic slabs of the Indotethys (Van der Voo et al., 1999) (subduction zone II in Fig. 12). The continuity of the KL Arc ophiolite westward with contemporaneous Semail ophiolite in Oman points to the existence of a continuous, Tethys-wide intraoceanic subduction zone during the Cretaceous. Arabia and India collided with the Oman–KL arc approximately 85 Ma with consequent ophiolite obduction and slab break off (Powell, 1979; Besse and Courtillot, 1988; Dilek and Furnes, 2009; Chatterjee and Scotese, 2010).

Radiometric (Jagoutz et al., 2009a, 2009b) and paleontologic (Baxter et al., 2010) ages from the KL Arc support an earlier collision event. Early Cretaceous radiolarians have been recovered from the Indotethyan sediments associated with the Ladakh volcanic activity, indicating that Indotethys was in existence for most of the Cretaceous (Baxter et al., 2010).

If India were isolated as an island continent as the late collision model predicts, India should have produced a highly endemic Late Cretaceous vertebrate assemblage during its long isolation. To the contrary, Late Cretaceous Indian vertebrates are cosmopolitan and show close faunal similarities with those of other Gondwana fragments such as Africa and South America. We prefer the early collision model, because dinosaur biogeography supports its existence. The KL Arc restored the long isolation of India temporarily as reflected by the homogenous Maastrichtian fossil vertebrates. The KL Arc formed a corridor for migration of several groups of Maastrichtian dinosaurs such as abelisaurs and titanosaurs between India and Africa that clearly indicates that India reestablished its contact with Africa during the Late Cretaceous time (Chatterjee and Scotese, 2010).

A synthesis of the possible tectonic evolution of the Kohistan–Ladakh collision zone is shown in Fig. 12 (Allègre, 1988; Van der Voo et al., 1999; Ali and Aitchison, 2008; Jagoutz et al., 2009a, 2009b; Burg, 2011). Prior to the collision of the KL Arc, the northward movement of India with respect to Asia reflects the subduction zone along the Makran–Indus Trench at about 120 Ma (Schettino and Scotese, 2005). As India (SL–I–LR) moved northward, subduction of the Indotethyan ocean floor beneath the KL Arc and its marginal basin, was followed by the collision, obduction, and suturing of the arc with the Indian continent during the Late Cretaceous (~85 Ma).

The KL Arc evolved as an obducted intra-oceanic island arc above a north-dipping subduction during the Late Cretaceous time (~85 Ma) in the equatorial area of the Neotethys Ocean (Tahirkheli, 1979; Reuber, 1986; Allègre, 1988; Van der Voo et al., 1999; Burg, 2011). The subduction of the Indotethys lithosphere along the Indus Suture underwent a rifting episode to produce the voluminous Chilas–Ladakh intrusion (Khan et al., 1989; Schaltegger et al., 2002). The KL Arc is characterized by remnants of an intraoceanic arc magmatic rock formed during Late Cretaceous times. These magmatic rocks range in age from 110 to 90 Ma, based on radiometric data both in Ladakh (Honegger et al., 1982) and Kohistan (Treolar et al., 1989). These oceanic terranes are mainly remnants of an arc-series, reflecting intra-oceanic subduction of the Indotethys. Continuing subduction resulted in complete consumption of the leading oceanic edge of the Indian plate, ultimately resulting in obduction of the arc onto the Indian continent at the site of the Indus Suture (Burg, 2011) (Fig. 12A).

The KL Arc was accreted to the Indian plate around 70 Ma and formed a biotic dispersal corridor for the migration of Gondwana dinosaurs during the Maastrichtian (Fig. 13) (Chatterjee and Scotese, 2010). It preserves a geochemical record that spans arc inception, arc thickening, and subsequent uplift (Jagoutz et al., 2009a, 2009b; Burg, 2011). It is an eastward extension of the Makran Arc, and its Cretaceous ophiolites represent remnants of the lost ocean of the Indotethyan seaway (Besse and Courtillot, 1988; Dilek and Furnes, 2009; Chatterjee and Scotese, 2010). Subsequently, the KL Arc was split from the Markran Arc during the KT boundary time along the Chaman Transform Fault and moved northward at a rapid rate with respect to Asia. The KL Arc collided and sutured against the Indian plate in the Late Cretaceous with the result that the arc was then converted into the leading edge of the Indian plate, the future NW syntaxis or indenter that would play a major role in the evolution of the Himalaya–Tibet orogeny (Fig. 12C–D).

7.5. Limited separation of Seychelles–Laxmi Ridge from India (~75–68 Ma)

Recent identification of seafloor spreading magnetic anomalies in the Laxmi and Gop basins lying between the Laxmi Ridge in the Arabian Sea and the Indian continent necessitates a change in the reconstructions of the western margin of the Indian plate (Bhattacharya et al., 1994). Associated with a spreading center, a major pre-Deccan volcanic center was located in the Laxmi and Gop basins, north of the Laxmi Ridge. This 7-km-thick plateau of extrusive basalt, called Somnath volcanic platform or Somanth Ridge, fills the basement of the offshore Indus Basin around the Saurasthra Arch, which is interbedded with Late Cretaceous–Paleocene marine sediments (Fig. 14A) (Carmichael et al., 2009; Yatheesh et al., 2009). Because of its offshore location, the Somnath flood basalt is poorly known when compared to the nearby Deccan Traps. The chronology of these two spatially and temporally linked rifting and volcanic events along the western margin of the Indian plate during the KT transition is highly complex and needs to be untangled. This is especially important if we are to reconstruct the early opening of the Arabian Sea.

Around 75 Ma, the Arabian Sea opened in two phases of rifting and flood basalt volcanism within a period of less than 10 Ma (Fig. 14B, C) (Collier et al., 2008; Minshull et al., 2008). The first brief phase of rifting formed the failed Gop Rift at some time between 71 and 64 Ma, when the Laxmi Ridge–Seychelles (LR–S) fragment separated from India (SL–I–KL) (Talwani and Reif, 1998). This rift occurred just prior to the eruption of the flood basalts of Deccan Traps associated with Somnath

Fig. 13. Paleogeographic map showing the collision of the Indian plate with the Kohistan–Ladakh island arc during the Late Cretaceous time along the Indus Trench (~80 Ma). For abbreviations, see Fig. 4.

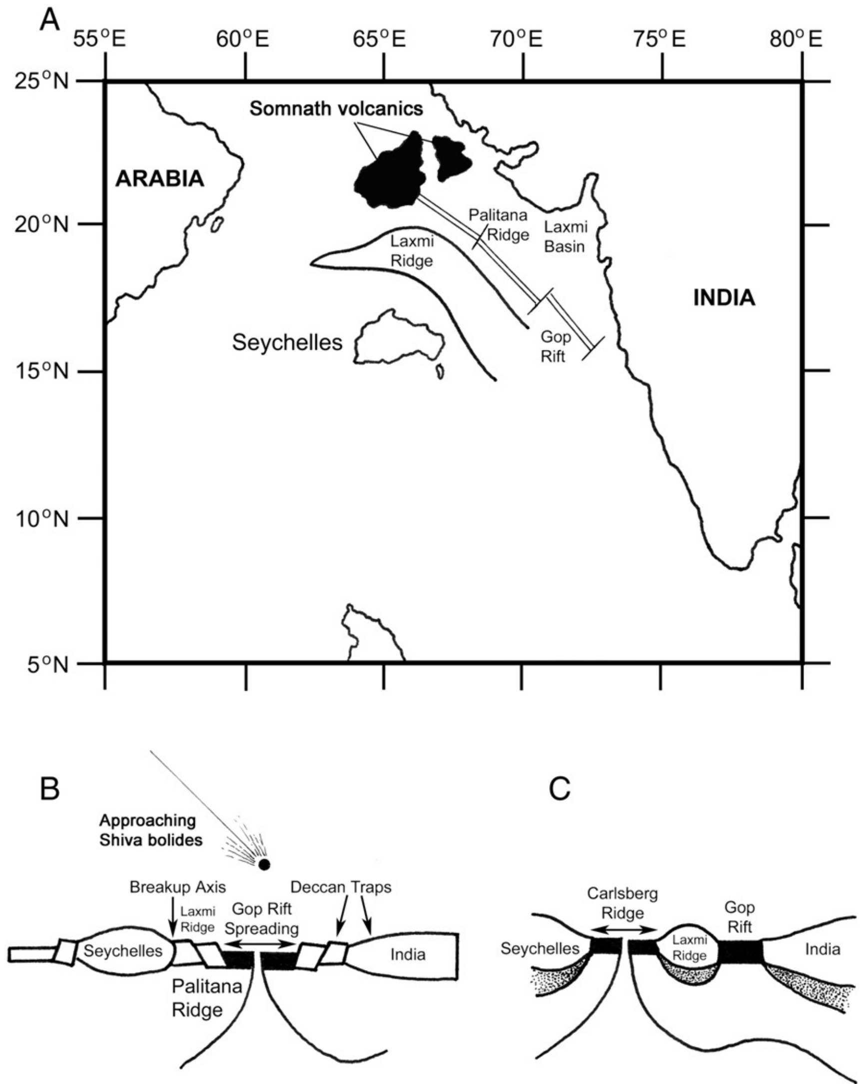

Fig. 14. (A) Location of the Laxmi Ridge, a continental sliver that separated from India around 70 Ma, probably induced by the Somnath volcanism. The spreading center, the Palitana Ridge was short-lived, created the Gop/Laxmi basin and the Laxmi Ridge was accreted to India. (B) Cartoon showing the first phase of rifting, the separation of Laxmi Ridge–Seychelles block from India around 70 Ma; it was a failed rift; (C) around 65 Ma, the second rifting took place between Seychelles and Laxmi Ridge–India, which was linked to Deccan–Reunion plume.
Panel A was modified from Chatterjee et al. (2006), Calves et al. (2008), and Carmichael et al. (2009). Panels B and C were modified from Minshull et al. (2008).

volcanism (Fig. 5). However, the spreading between the Laxmi Ridge and India was short-lived and the Laxmi Ridge remained attached to India. A second rifting event created a sea-floor between the Seychelles (S) and the Laxmi Ridge–India (SL–I–LR–KL) block starting at 63.4 Ma. The Laxmi Ridge–Seychelles rift occurred around the time of the formation of the Deccan flood basalts (~65 Ma), and the two events are generally causally linked. Though the Gop rift margins show extensive magmatism, the Seychelles–Laxmi Ridge rift margins are largely amagmatic.

The Laxmi Ridge is an enigmatic NW–SE trending basement high in the Arabian Sea, about 700 km long and 100 km across, which is buried beneath the sediments of the Indus Fan. It is located about 500 to 700 km off the west coast of India, continues northwards but tapers and bends to the west (Fig. 14A). Although the origin of Laxmi Ridge is controversial, gravity and seismic data indicate that it is a rifted micro-continental fragment that was separated from the Indian plate by an early phase of spreading (Talwani and Reif, 1998). Between the Laxmi Ridge and the Indian continental shelf is a wide region, called Gop Basin and its southern extension Laxmi Basin, whose genesis is important concerning the geodynamic evolution of the Arabian Sea.

Magnetic anomalies indicate that Gop Basin opened between 73 Ma to 68 Ma, which is older than the initial phase of the Deccan volcanism (Malod et al., 1997; Collier et al., 2008). The Palitana Ridge has been identified as the extinct spreading center that created the Gop–Laxmi Basin (Yatheesh et al., 2009). A large pre-Deccan igneous province, the buried Somnath volcanic platform is composed of different volcano-stratigraphic edifices, which were formed around 75–68 Ma in the offshore segment of western India north of the Palitana Ridge. The Somnath volcanics appear to be causally linked with the rifting of the Laxmi Ridge from India (Calves et al., 2011) (Fig. 14A). Most likely the rifting and spreading of the Gop Basin predated the nearby emplacement of the Deccan Traps and was short-lived. The seafloor spreading died at the Gop Basin around 68 Ma, when the Seychelles–Laxmi Ridge microplate became part of India for a brief time. Subsequently, during the outburst of the Deccan volcanism around 65 Ma, a new spreading center, the Carlsberg Ridge would develop southwest of the Laxmi Ridge that would separate the

Seychelles microcontinent from India, while the Laxmi Ridge remained attached to India (SL–I–LR–KL complex).

7.6. Separation of the Seychelles from India (~65 Ma)

The abortive phase of rifting between the Laxmi Ridge and India was soon followed by the second phase of rifting and seafloor spreading between the Seychelles (S) and India (SL–I–LR–KL) complex, which was coeval with the main phase of the Deccan Trap eruption (Fig. 14C). The Deccan volcanic province is one of the largest volcanic eruptions in Earth history and has received global attention for its possible role in KT boundary mass extinction. The outpouring of the enormous continental flood basalts of the Deccan Trap, spreading over vast areas of western and central India and the adjoining Seychelles, cover more than 1,500,000 km^2, with the greatest thickness of about 3.5 km along the Western Ghats escarpment (Figs. 5, 15) (Baksi, 1994). Recent works (Chenet et al., 2007; Keller et al., 2011) recognize three district pulses of Deccan volcanism: the initial pulse of eruption (phase-1) occurs in the late Maastrichtian base of C30n (~68.5 Ma), and is represented by Sarnu–Dandali volcanics in Barmer District and Mundwara volcanics in Sirohi District. Both occur as isolated outcrops in Rajasthan in northern India (Basu et al., 1993). Khan et al. (1999) suggested that the Parh Group of volcanics of northeastern Baluchistan (Pakistan) might represent the earliest (~67.5–75.9) and northernmost manifestation of the phase-1 of Deccan volcanism. The first phase of volcanism was small, initiated on the Neotethyan floor of Parh and continued southward with sporadic eruptions in Rajasthan, which was followed after a quiescence of 2 to 3 Myr by the main phase of Deccan volcanism spanning the KT transition.

The main pulse of volcanism occurs in chron 29R and ending at the KT boundary (~65 Ma) in the main Deccan volcanic province in the western and central part of the Indian shield. The main phase of Deccan volcanics accounts for about ~80% of all the traps. In Jhilmili region of central India, a planktic foraminifer assemblage of Early Danian age has been encountered above the lower Deccan basaltic flow. The last pulse erupted in the upper part of chron 29R and ended within chron 29N in the Early Danian (~62.3 Ma). The exposures of the last phase of Deccan volcanism have been traced eastwards to the Rajahmundry area of the Krishna–Godavari Basin and out into the Bay of Bengal. These traps were overlain by Early Danian age of planktic foraminifera (Keller et al., 2011). The intercanyon flows of the Deccan lava along the interconnected rift basins of Son–Narmada, Cambay, and Godavari drainage systems may explain the enormous areal distribution of Deccan Traps (Chatterjee and Rudra, 1996).

The main pulse of the Deccan basalts erupted very rapidly—probably within 1 Myr—at the KT boundary (~65 Ma), when western India lay above the Réunion hotspot, which is now located east of Madagascar (Courtillot, 1999; Chenet et al., 2007). It was a major tectonic event, which produced one of the largest flood basalt provinces on the Earth's surface. However, the relationship of the India–Seychelles rift and Deccan volcanism is far more complex than the traditional explanation.

When the Deccan Traps erupted, active seafloor spreading was going on in the Mascarene basin, as the Seychelles–Laxmi Ridge–India block drifted away from Madagascar (Fig. 14). Consequently, the plume–rift interaction is thought to have given rise to the rapid eruption of prodigious volumes of Deccan basalt (~10^6 km^3; Courtillot et al., 1986; White and McKenzie, 1989). At the KT boundary time, major reorganization and a rift jump occurred in the Indian Ocean shortly after the emplacement of the Deccan Traps (Malod et al., 1997). During the eruption of the Deccan basalts, spreading died out at the Mascarene basin at the time of chron 28, when the Central Indian spreading ridge in the Mascarene basin jumped 500 km northward to a new location between the Seychelles and the Laxmi Ridge to form the Carlsberg Ridge (Schlich, 1982). At the same, the Palitana Ridge became extinct in the Gop Basin and probably jumped westward to the Carlsberg Ridge (Calves et al., 2011). The beginning of seafloor spreading at the Carlsberg Ridge is well constrained by magnetic anomalies at 63.4 Ma (Collier et al., 2008). Thus the eruption of the Deccan basalts coincided with the separation of the Seychelles from the Laxmi Ridge–India block and followed by rapid seafloor spreading (White and McKenzie, 1989). As a result of this ridge jump, the Seychelles was transferred to the African plate (Fig. 16).

The Deccan flood basalts were formed along the two sides of the rifted margins, western India and eastern Seychelles and its submerged Saya de Malha bank and were implicated for the India–Seychelles rift

Fig. 15. Paleogeographic reconstruction of India–Seychelles at the KT boundary (~65 Ma) time. India–Seychelles separation is generally linked to the Reunion plume (Chatterjee et al., 2006). For abbreviations, see Fig. 4.

Fig. 16. Paleogeographic reconstruction of India in relation to other Gondwana continents during the Cretaceous–Tertiary boundary (~65 Ma). The Reunion plume (~65 Ma), is generally linked to the separation of the Seychelles from the Laxmi Ridge–India block. For abbreviations, see Fig. 4.

(White and McKenzie, 1989). The age of this magmatism in Seychelles is coeval with Deccan Traps, ranging from 66 Ma (Croxton et al., 1981) to 64 Ma (Duncan and Hargraves, 1990). As the breakup between Seychelles and India progressed and India moved northward, a hotspot track formed underlying the Laccadive, Maldive, and Chagos islands; the Mascarene Plateau; and the youngest volcanic islands of Mauritius and Réunion (Duncan, 1981) (Fig. 5).

In spite of close temporal association between Deccan Traps and the Seychelles–Laxmi Ridge rifting, the lack of volcanism at the opposing continental margins of Laxmi Ridge and Seychelles is anomalous. Moreover, the continental rift between Seychelles and India was approximately 1000 km from the Reunion hotspot at the time of breakup, and the oceanic crust seaward of the Seychelles and Laxmi Ridge is thinner than expected. The separation of the Seychelles from the Laxmi Ridge provides a new twist that when two parts of continents break apart, there are not always accompanying massive volcanic eruptions. Perhaps, the Seychelles–Laxmi Ridge separation is a passive rift lying above melt-depleted mantle that led to non-volcanic (magma-poor) margins (Armitage et al., 2010). Most likely, voluminous magmatism in the neighboring Gop rift 6 million years earlier partially exhausted the supply of magma. When the rifting migrated to the new spreading center at the Laxmi Ridge–Seychelles basin at 63 Ma, the thermal anomaly had cooled down, and the mantle was depleted. These factors may explain why the Seychelles–Laxmi Ridge rifting was not associated with magmatism. Despite the apparent proximity of this rift margin to a thermal plume, thin oceanic crust can be generated by stretching and rifting before the arrival of mantle plume (Armitage et al., 2010). After the ridge jump, Seychelles became fixed to Africa, resulting in a shift of the spreading axis as India (SL–I–LR–KL) separated from the Seychelles.

7.7. Acceleration of the Indian plate during the Late Cretaceous–Early Eocene and the subduction of the Neotethys

During most of the Cretaceous, the Indian plate moved northward at a rate of 3–5 cm/year. Plate reconstructions based on paleomagnetic data suggest that the Indian plate suddenly accelerated to 20 cm/year from Late Cretaceous (~67 Ma) to Early Eocene (~50 Ma) (Fig. 15). During this time interval India moved rapidly northward between two great transform faults, the Ninety East Ridge on the east and Owen–Chaman Fault on the west. Some 52 million years ago India slowed to 5 cm/year during the initial stage of collision with Asia and maintained this slow speed throughout its convergence (McKenzie and Sclater, 1971; Patriat and Achache, 1984; Copley et al., 2010).

The rapid acceleration of India during the Late Cretaceous–Paleocene interval makes India unique among the fragments of Gondwana, but the cause of this rapid motion remains unclear. Negi et al. (1986) suggested from heat flow data that the Indian lithosphere was greatly thinned (about one third the thickness of other Precambrian shields), abnormally hot, and lighter during this interval of acceleration, probably triggered by the Deccan volcanism, which had important consequences for mantle rheology. It reduced the drag of the lithosphere against the asthenosphere, resulting in faster northward movement of the Indian plate. According to these authors, the Indian plate decoupled itself from the deeper interior to become more mobile. A similar model for a thinner lithosphere of the Indian plate, about 100 km deep, has been suggested recently that may account for its rapid drift (Kumar et al., 2007). In this model, the Indian plate was thinned by the plume activities (such as Marion, Kerguelen, and Reunion). The loss of the lithospheric roots might have been the reason for its rapid acceleration. India's lithosphere is only half as thick as other Gondwana continents, which is the reason for its high-speed collision with Asia. Recently, Cande and Stegman (2011) suggested the push force of the Reunion plume head might have caused the sudden acceleration of the Indian plate.

After the KT extinction, the Indian plate continued to move northward, carried the accreted Kohistan–Ladakh Arc along its northern edge, severed its connection with the Oman arc along the Quetta–Chaman fault, and became an island continent. It carried its endemic biota like a “Noah's Ark” (McKenna, 1973; Chatterjee and Scotese, 2010). India continued to move northward with the accreted KL Arc (SL–I–LR–KL) across the Neotethys (Fig. 16). Subduction of the Neotethyan ocean floor beneath Asia was followed by an initial collision of the arc with Karakoram along the Shyok Suture in Early Eocene time (~50 Ma). Continuing convergence led to complete suturing of accreted India with Asia along the Shyok–Tsangpo Suture Zone (Figs. 12B–C, 18). The KL Arc was wedged between India and Asia as a result of the closure of the Neotethys (Bard, 1983). Today, the KL Arc is located on the northern edge of the Indian subcontinent on either side of the Nanga Parbat and is bounded by two suture zones: the Shyok Suture (SS) to the north and the Indus Suture (IS) to the south (Bouihol et al., 2011; Burg, 2011) (Fig. 11).

8. Cretaceous climate

The Early Cretaceous was a mild “Ice House” world. There was snow and ice during the winter seasons, and cool temperate forests covered the polar regions. During the Late Cretaceous the global climate was warmer than today's climate, where the average annual temperatures at the Equator topped 38 °C. No ice existed at the poles. During the much of the Cretaceous India resided in the Southern Hemisphere as an island continent, encircled by the ocean. Much of the Tethyan oceanic crust advancing with India from the Southern Hemisphere was covered by a thick carbonate platform deposit, which at the subduction zones released a large flux of CO_2 in the atmosphere to trigger global warming (Kent and Muttoni, 2008).

During the Early Cretaceous (~140 Ma) the center of the Indian subcontinent was situated at 40°S and was rotated 90° clockwise with respect to the present orientation. Situated in between the Subtropical Arid and Temperate Climate Belt, the overall precipitation was low to medium (1.5 cm/month to 12 cm/month). According to FOAM paleoclimate simulations (Scotese et al., 2007, 2008, 2009, 2011; Goswami, 2011) (Fig. 17A), the western part of the present subcontinent India was relatively dry, while the eastern part had higher precipitation. The mean annual temperature ranged between 17 °C (presently in the western part of the Indian subcontinent) and −3 °C (eastern India).

During the Cenomanian–Turonian time (~90 Ma), which is thought to be one of the warmest of all geologic periods, the Indian subcontinent was located at 35°S. The annual mean temperature ranged between 12 °C in the south and 28 °C in the northeast. Mean annual precipitation ranged between 5 cm/month and 13 cm/month. The present day northwestern province of the Indian subcontinent was strictly in the Subtropical Arid Belt and had very low annual precipitation (Fig. 17B).

In the latest Cretaceous (~70 Ma), the Indian subcontinent had drifted northward to 25°S (Fig. 15B). India lay between the Equatorial Rainy Belt and Subtropical Arid Belt, the mean annual temperature ranged between 17 °C and 26 °C (Scotese et al., 2007, 2008, 2009, 2011; Goswami, 2011). The eastern flank of the subcontinent was for the first time drier than the western part. The average mean annual precipitation was estimated to be between 20 cm/month (in the western part of the subcontinent) and 6 cm/month (in the eastern part of the subcontinent).

9. India–Asia collision and Cenozoic evolution of the Himalayan–Tibetan orogeny

After the KT extinction, India had become an island continent moving at a rapid rate of ~20 cm/year in relation to Africa. It enjoyed its splendid isolation during the Paleocene until it collided into Asia in Early Eocene and slowed down to 5 cm/year (Copley et al., 2010). The collision of India with Asia has been extensively studied and refined

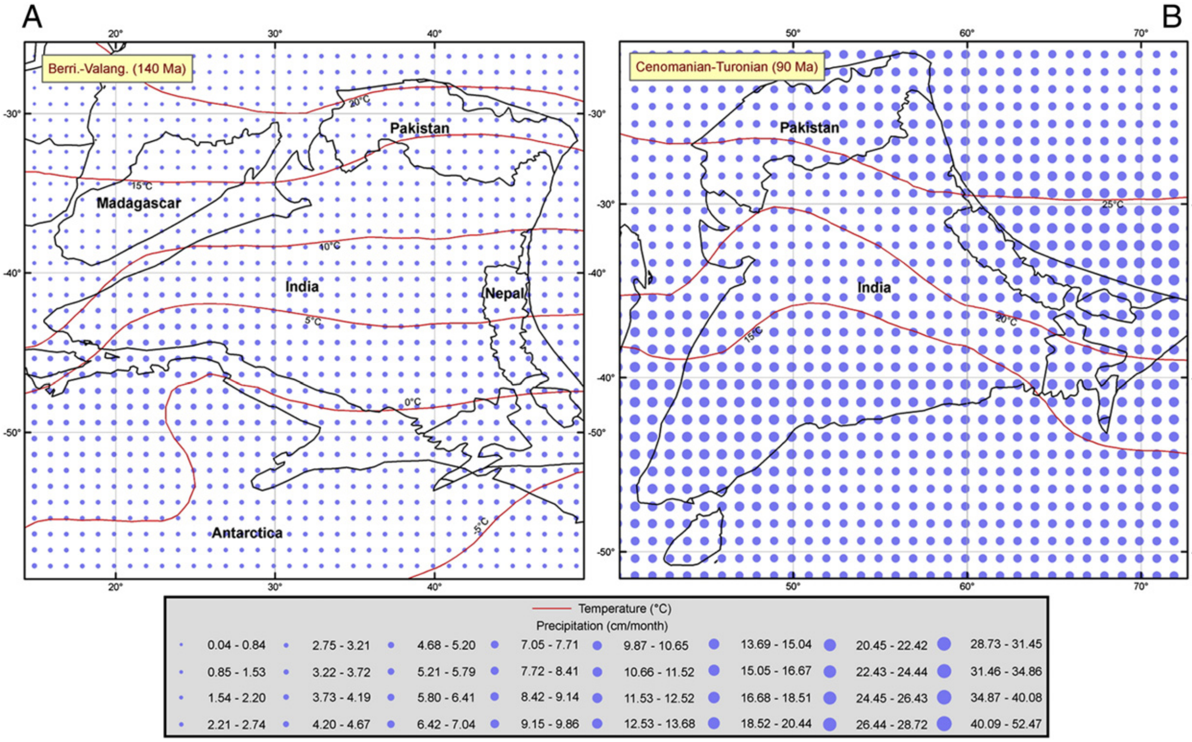

Fig. 17. Mean annual precipitation and mean annual temperature maps of the Indian plate during the Cretaceous time. (A) Early Cretaceous map (~140 Ma). (B) Late Cretaceous map (~90 Ma).

during the past 40 years; yet many of the complexities of the evolution of this orogenic system remain poorly understood. A synopsis of the initial collision and post-collisional events is summarized below.

Fig. 18. Paleogeographic reconstruction showing the position of India and other Gondwana continents during the Early Eocene (~50 Ma) when India made the initial collision with Asia on its northward journey with the closure of the Neotethys. The Kohistan–Ladakh arc made the first contact with Asia. For abbreviations, see Fig. 4.

9.1. India–Asia collision and the evolution of the Indus–Tsangpo Suture Zone (~50 Ma)

The most spectacular example of continental collision event is the convergence of India with Asia that occurred in Early Eocene (~50 Ma) (Molnar and Tapponnier, 1977; Allègre et al., 1984). This collision resulted in the closure of Neotethys and ended the rapid northward flight of India (Fig. 18). The collision between India and Asia took place along three suture zones: the Indus, Shyok, and Tsangpo (Fig. 11). The southern margin of Tibet maintained relatively stable northern hemisphere paleolatitudes during the Early Cenozoic (Achache et al., 1984). Since Asia was more or less stationary, the Neotethyan seafloor subducted along a northward dipping Shyok–Tsangpo Suture Zone along the south shore of Lhasa block, which ultimately led to the formation of the Gangdese batholith in the Transhimalayan belt (Allègre et al., 1984). It has been estimated that at least 4000 to 7000 km of the Neotethyan oceanic lithosphere was subducted beneath the southern margin of Asia during the Cretaceous and early Cenozoic (Van der Voo et al., 1999).

Does the Indus–Tsangpo or the Shyok–Tsangpo Suture Zone mark the site of the final India–Asia collision? The relative and absolute timing of the formation of these sutures—the Indus, Shyok, and Tsangpo—has been discussed and debated for decades. Much of the debates center on the sequence and timing of the collision of the Kohistan–Ladakh Arc with India and Asia respectively. If the KL arc collided with India first around 85 Ma along the Indus Suture, as discussed earlier (Reuber, 1986; Allègre, 1988; Van der Voo et al., 1999; Ali and Aitchison, 2008; Jagoutz et al., 2009a, 2009b; Chatterjee and Scotese, 2010; Burg, 2011), the Shyok Suture becomes the northern branch of the Indus–Tsangpo Suture Zone, that is to say, the initial locus for the collision (Jagoutz et al., 2009a, 2009b) (Fig. 11). The Shyok Suture is generally considered to be the collision zone between the Kohistan–Ladakh Arc and the Karakoram Mountains that formed during the collision of India with Asia in Eocene (around ~50 Ma). In the south, the KL Arc is attached to the Indian continent by the Indus Suture Zone. However, the Shyok Suture Zone has no lateral equivalent east of the Karakoram fault zone. The subduction at the Shyok–Tsangpo Suture (STS) continued as India came into contact with Asia around ~50 Ma.

In our view, the Shyok–Tsangpo Suture Zone marks the initial site of the India–Eurasia collision, which later shifted to the Indus–Tsangpo Suture Zone with intracontinental subduction along this reactivated suture. Geological evidence from Karakoram suggests that the Shyok subduction zone was locked after the closure of the Neotethys so the leading edge of continental parts of India began to subduct along the reactivated Indus–Tsangpo Suture Zone during the post-collisional phase (Burg, 2011). The orogenic deformation front has moved southward from the initial collision contact at the Shyok Suture to the Indus Suture. The Indus–Tsangpo Suture Zone defines the areas of final collision between the Indian plate and the Tibetan Lhasa block, and marks the zone along which the Neotethys Ocean was consumed by subduction processes. In the west, with the closure of the Neotethys, the Kohistan–Ladakh arc was carried to the Shyok Trench, and was added to the overriding Lhasa block (Burg, 2011). With the closure of the Neotethys, the buoyant Indian plate could not be subducted. As a result subduction slowed, and so did the associated igneous activity along the southern margin of the Tibet. As India collided with Tibet, these two continents were welded together along a zone marking the former site of subduction.

The Indus–Tsangpo Suture (ITS) zone reveals more complexity than previously recognized and contains two superimposed subduction events: the earlier event (~85 Ma) marks the origin of the Kohistan–Ladakh Arc; the latter tectonic event defines the zone of collision between the Indian and Eurasian plates (~50 Ma). The ITS can be traced discontinuously for over 3000 km from Kohistan in Pakistan in the west to the NE frontier region of India–Myanmar (Gansser, 1964, 1966, 1980). It crops out in the upper valleys of the Indus and Tsangpo (Brahmaputra) rivers, and is composed of deep-sea and flysch sedimentary rocks, ultrabasic and submarine volcanic rocks, and plutonic intrusions (Fig. 11). The structure of the ITS is complex and contains three distinct components from north to south: the Chilas and Dras volcanics, the relicts from the Kohistan–Ladakh arc; the ophiolite mélange from the Neotethys oceanic crust; and the 6-km-thick Indus molasses of the Siwalik group along the foredeep of the rising Himalaya, which are post-collisional continental clastic sequences derived mainly from the Ladakh batholith (Dezes, 1999).

The age of the Tsangpo Suture remains uncertain, with recent estimates ranging from 70 to 35 Ma, while the age of the Indus Suture is well constrained to ~50 Ma, and coeval with the end of marine sedimentation south of the Tsangpo Suture (Jagoutz et al., 2009a, 2009b). This corresponds to the marked slowdown in the northward migration of the Indian plate (5 cm/year) because of buoyancy resistance at this time (Klootwjik et al., 1985). The timing of the closure of the Neotethys along the Indus–Tsangpo Suture and the India–Asia collision is well constrained by paleomagnetic, sedimentologic, and paleontologic data at ~50 Ma (Molnar and Tapponnier, 1977; Allègre et al., 1984; Molnar, 1986). The rapid deceleration of the Indian plate from 20 cm/year to 5 cm/year is normally suggested as the initial timing for the onset of the India–Asia collision due to combination of orogeny-related increased trench resistivity and decreased slab pull due to continental subduction (Copley et al., 2010; Douwe et al., 2011). As the Neotethyan Ocean contracted, marine transgression was widespread. This is witnessed by the presence of Eocene Nummulitic Limestone (~54 Ma) far south of the Neotethys (Powell and Conaghan, 1973).

The closure of the Neotethys is marked by the change from marine to terrestrial conditions displayed by sediments in the ITS zone between 54 and 50 Ma. The northward movement and counter-clockwise rotation of the Indian plate slowly closed the Neotethys around 50 Ma. In Ladakh the youngest marine sediments are Paleocene to Lower Eocene limestones, which are succeeded by continental red sandstones, suggesting that the Neotethys Ocean must have closed by 50 Ma with southward obduction of the Indus–Tsangpo Suture Zone (Searle, 1991).

Vertebrate fossils provide an independent clue to the initial timing for the India–Asia collision event in Early Eocene (~50 Ma) time, when India had established subaerial contact with the Asian mainland leading to a dispersal corridor between the two landmasses. A great faunal interchange took place between India and Asia during the initial collision and coincided with the Paleocene–Eocene Thermal Maximum (PETM). This is reflected by the start of terrestrial deposition in several regions of Indo-Pakistan including the Cambay Formation of Gujarat, coal-bearing Ghazij Formation of Baluchistan, and Subathu Formation in Jammu and Kashmir region, which have yielded prolific vertebrate fossils. These sediments were derived from a mixture of sedimentary, volcanic and ophiolite rocks from the Indus–Tsangpo Suture during the Early Eocene indicating the timing of the collision and development of foreland basin (Najman and Garzanti, 2000). A biotic corridor between India and Asia was established that enabled hordes of Asian vertebrates (such as turtles, crocodiles, and a variety of marsupials, placental mammals) to sweep into India in the Early and Middle Eocene time. Similarly, other group of vertebrates of Gondwanan origin (such as squamates and crocodilian reptiles, and marsupial and placental mammals) migrated to Asia from India during this time (Chatterjee and Scotese, 2010). The substantial faunal exchange of Eocene vertebrates between Asia and India supports the timing of the initial collision event.

9.2. Himalaya: post-collisional tectonics (~50 Ma to Holocene)

The plate reconstruction at the time of collision (~50 Ma) places the Indian subcontinent ~2500 km south of its current position (Besse et al., 1984). Three mechanisms have been suggested to explain the 2500 km missing "gap" between India and Asia: (1) subduction of the continental Indian plate (Greater India) below the Tibetan Plateau; (2) crustal shortening of the leading edge of the Indian plate by thrusting and folding with the rise of the Himalaya; and (3) extrusion tectonics in east-central Asia, where Indochina and China blocks were squeezed eastwards (Gansser, 1966; Molnar and Tapponnier, 1977; DeCelles et al., 2001).

Continental collision between India and Asia, from Eocene to Holocene, resulted in profound crustal shortening and thickening that produced the Himalaya and Tibetan Plateau. The Himalayan orogen is the type example of continent–continent collision (Dewey and Bird, 1970; Molnar and Tapponnier, 1977; Molnar, 1986). The Himalaya, the most stunning mountain range in the world, forms a spectacular arc of 2500 km along the leading margin of the Indian plate and are bounded by the Nanga Parbat syntaxis in the northwest and the Namche Barwa syntaxis in the northeast, with an average width of 250 km (Fig. 11). The Himalaya was uplifted, folded, and complexly thrusted in response to the India–Asia collision during the Cenozoic. It comprises a series of lithologic and tectonic terranes that run parallel to the mountain belt (Gansser, 1964, 1966). From north to south they are as follows: (1) the Trans-Himalayan batholiths; (2) the Indus–Tsangpo Suture Zone; (3) the Tethyan (Tibetan Himalaya); (4) the Higher (Greater) Himalaya; (5) the Lesser (Lower) Himalaya; and (6) the Sub-Himalaya (Fig. 19).

After the Eocene closure of Neotethys, deformation spread southward across the Tibetan–Tethys Zone into the Higher Himalaya. The main Himalayan range has not developed from the Neotethyan geosyncline, but rather is of intracratonic origin. It is only in the Tibetan Himalaya that Neotethyan sediments are preserved. The Tibetan Himalaya is about 40 km wide at an altitude of 4000 m, consisting of fossiliferous shallow-water marine sediments of Proterozoic to Eocene age. Due to continuing convergence of India with Asia, the northern edge of the Indian plate was fractured, thrusted, and uplifted, resulting in the formation of the Himalaya and the associated

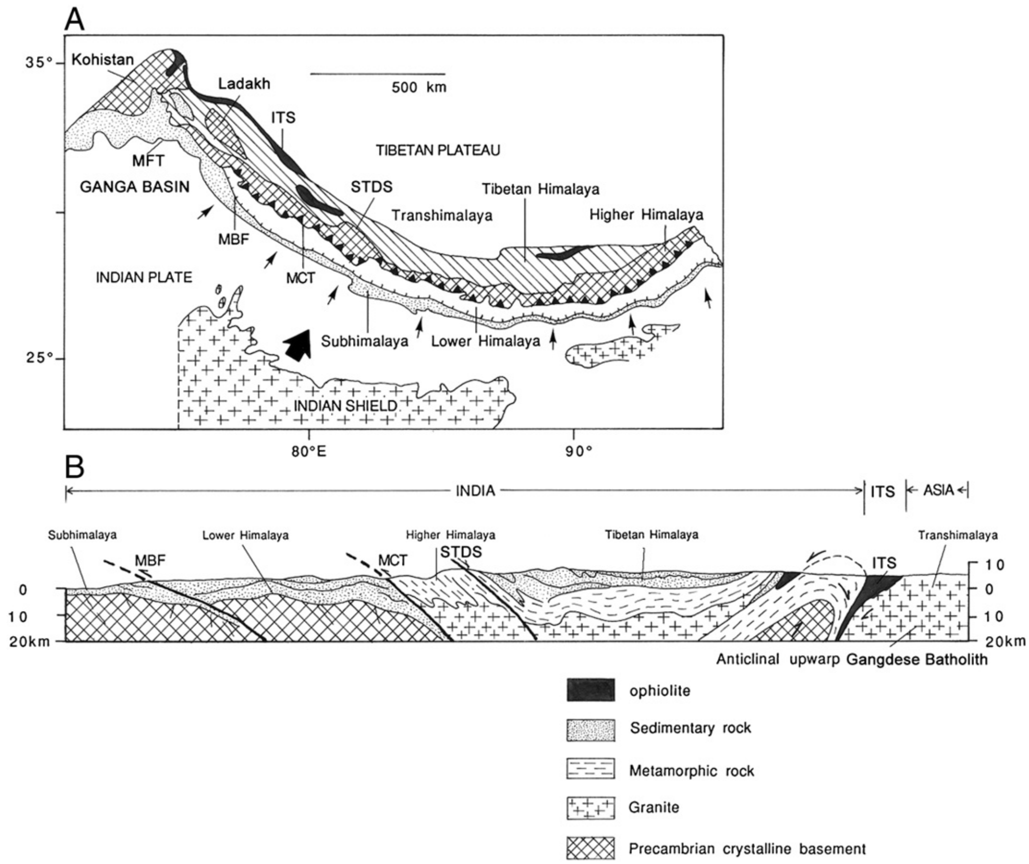

Fig. 19. Himalayan orogen. (A) Morphotectonic zones of the Himalayan arc, showing its asymmetric plan pattern in relation to India and associated suture and thrust zones from north to south: ITS, Indus–Tsangpo Suture; STDS, South Tibet Detachment System; MCT, Main Central Thrust; MBF, Main Boundary Fault; MFT, Main Frontal Thrust. (B) Schematic cross-section of the Himalaya showing different morphotectonic zones and thrusts.
After Chatterjee (1992).

structure. The higher, lesser, and Sub-Himalayas are thus slices of the old Indian shield that have stacked over each other by transfer along a series of southward verging thrusts.

The post-collisional tectonics is best expressed along the Higher Himalaya region (or Central Crystalline axis), which is about 140 km wide, forming the backbone of the Himalayan orogen. It consists of snowcapped peaks comprising all 10 of world's highest peaks including Mount Everest (8840 m) and K2 (8611 m). The Higher Himalaya consists of Precambrian gneiss overlain by Early Paleozoic and Mesozoic sediments of Tethyan origin, which were originally located on the northern margin of India (Gansser, 1964). The contact between the Tethys Himalaya and the Higher Himalaya is marked by a series of north-dipping faults and shear zones called the South Tibetan Detachment System (STDS), which probably formed during the Miocene time and a played critical role in the Neogene evolution of the Himalayan orogen (Burchfiel et al., 1992). The STDS extends ~2500 km along strike from Zanskar in NW India to Arunachal Pradesh in NE India.

South of the Higher Himalaya, the Lesser or Lower Himalaya stretches in parallel with an average height of 3900 m to 4500 m, and a width ranging from 32 to 80 km. The Lesser Himalaya consists mainly of Late Proterozoic to Early Cambrian detrital sediments derived from the passive Indian margin intercalated with some granites and acid volcanics of Early Proterozoic age (Frank et al., 1977). These sediments were thrust over along the Main Boundary fault (MBT) and often appear as tectonic windows (Gansser, 1964, 1966) (Figs. 19, 20).

South of the Lesser Himalaya lies the Sub-Himalaya that forms the foothills of the Himalayan range with an average height of 900 m to 1200 m, and width ranging from 8 to 80 km. The sediments contain the Eocene Sabathu Limestone, followed by the Miocene to Pleistocene molasse deposits, known as the Muree and Siwalik formations, which are rich in vertebrate fossils. Sub-Himalayan rocks have been overthrust by the Lesser Himalaya along the Main Boundary fault (MBF) that developed during the Pliocene time. In turn, the Sub-Himalaya is thrust along the Main Frontal Thrust (MFT) over the Holocene alluvium of the Ganga Basin deposited by the Himalayan river system (Fig. 20A).

9.3. Tectonic evolution of the Himalaya

The tectonic evolution of the Himalaya has been reconstructed through painstaking fieldwork spanning a century by a group of international earth scientists, integrating mapping, structural geology, plate tectonics, and aided by satellite imageries. However, in recent times, tomographic imaging has provided a new dimension to the interpretation of the subsurface structure of the Himalaya–Tibetan orogen that reveals a cross-sectional view of the crust and the upper mantle. Seismic tomography is an imaging technique that uses seismic waves generated by earthquakes and explosions to create computer-generated 3-D images of Earth's interior. Since the Himalaya–Tibetan orogen is seismically active, tomography has provided an important tool in interpreting the subsurface structure.

Two models have been proposed to explain the tectonics of the Himalaya: (1) a traditional thrust belt model that typically develops during plate collisions in which folding and faulting in the upper crust accommodate deformation and localized ductile shear in the lower crust (Gansser, 1964, 1966; Dewey and Bird, 1970; Powell and Conaghan, 1973; Molnar and Tapponnier, 1977; DeCelles et al., 2001); and (2) a channel-flow model revealed from tomographic imaging of the mantle under India and Tibet in which the subducting Indian plate underneath Tibet undergoes partial melting and flows southward, directly in the

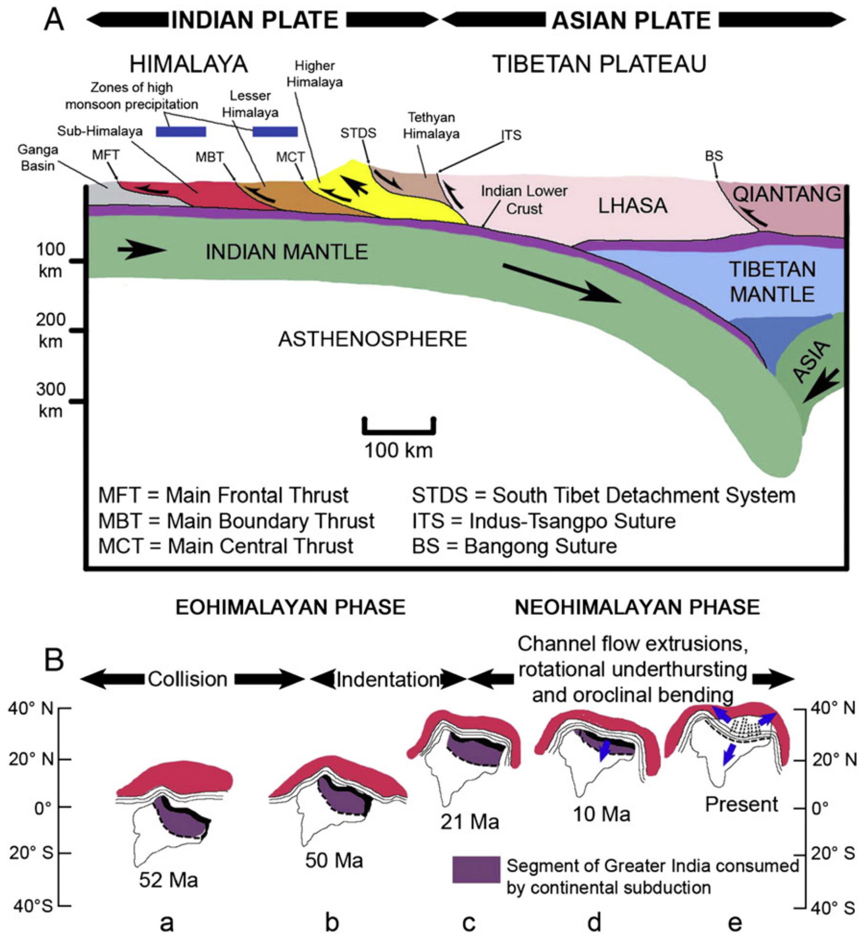

Fig. 20. (A) Tectonic interpretation of the Himalayan orogeny using high-resolution seismological and tomographic data showing north-vergent thrust systems in the Himalaya such as MFT, MBT, and MCT producing a series of continental slices and moving southward in relation to Indian mantle; the lower part of the Indian lithosphere along the Main Frontal Thrust underplates the Himalaya and Tibet up to 31°N, and almost extends to the middle part of the Lhasa block. However, the faulting at STDS is anomalous; it is a normal fault dipping northward below the Tibet and undergoes melting. Both MCT and STDS facilitated the channel flow extrusion of the Higher Himalaya; (B) five main stages in the evolution of the India–Tibet collision.
Panel A was simplified from Hodges (2006b) and Nabalek et al. (2009). Panel B was modified from Klootwjik et al. (1985).

direction of least resistance, toward the Himalayan mountain front (Beaumont et al., 2001; Hodges, 2006a, 2006b). Here we discuss the thrust belt model first, followed by a discussion of the channel flow model.

The collision of the Indian plate with Asia is the reason for the rise of the Himalaya and the uplift of the Tibetan Plateau. Several models have been proposed for the evolution of the Himalaya that are based on collision tectonics (Gansser, 1964, 1966; Le Fort, 1975; Molnar and Tapponnier, 1977; Patriat and Achache, 1984). Let us start from the Indus–Tsangpo Suture, the collision zone between India and Asia. South of the ITS, virtually all of the sedimentary rock of the Tethys Himalaya appears to have been part of the Indian subcontinent or its northern continental margin from the time of its deposition until the collision with Asia. Continued convergence of the Indian plate at 5 cm/year since the Eocene has led to the progressive development of a series of thrusts in the Himalaya accompanied by some 2500 km of crustal shortening (Patriat and Achache, 1984). There seems to be a steady southward progression of thrust faulting, beginning at the Indus–Tsangpo Suture (ITS) when collision occurred in Early Eocene time. Subsequently, the Main Central Thrust (MCT), the Main Boundary Fault (MBT), and the Main Frontal Thrust (MFT) occurred farther and farther south of the ITS (Allègre et al., 1984). All these thrusts are subparallel and form a NNE–SSW arcuate trend, pushing up the Himalaya in front of each of them. North–south vertical cross sections through Himalaya show that the MCT, the MBF, and the MFT fault systems merge at a depth to become the Himalayan Sole Trust fault (Fig. 20A). Crustal material above these faults is moving southward relative to the northward-subducting lithosphere of the Indian plate; it may mark the top of the fluid lower crustal channel below Tibet (Hodges, 2006b).

It is generally believed that thrusting within the Himalaya is a relatively late stage response in overall accommodation of India's collision with Asia during Neogene time (Gansser, 1964, 1966). Extensive crustal shortening took place along the entire 2400 km long northern edge of the Himalaya, along these thrust planes, which decrease in age southwards (Gansser, 1964; Powell and Conaghan, 1973; Le Fort, 1975; Molnar and Tapponnier, 1977). Thrust faulting and major uplift of the Himalaya is an obvious manifestation of the India–Asia collision. In addition to thrusting, the NNE motion of the Indian plate, accompanied by counterclockwise rotation, may explain the pronounced asymmetrical pattern plan and the apparent oroclinal bending of the Himalayan arc during the last 10 Myr (Klootwjik et al., 1985) (Fig. 19).

Klootwjik et al. (1985) recognized five stages of collision chronology during the evolution of the Himalaya, which are slightly modified here with new data and addition of two stages (Hodges, 2006a): (1) initial collision in Early Eocene (~50 Ma); (2) complete of suturing in Early Eocene (~50 Ma); (3) sediments on the leading margin of the Indian plate squeezed up and subduction ceases along the ITS during Middle Eocene–Late Oligocene with the development of the Tethyan Himalaya; (4) onset of underthrusting of the continental Indian plate in Early Miocene (~21 Ma) along the Main Central Thrust with synchronous development of South Tibet Detachment System between Higher and Tethys Himalaya, resulting in channel flow extrusion of Higher Himalaya; (5) continued underthrusting of India along the MCT and indentation of Tibet as India rotated counterclockwise in the Late Miocene (~10 Ma) in relation to Asia, causing southward extrusion of Indochina; and (6) oroclinal bending of the Himalayan arc in the last 10 Myr in response to continued convergence, and the development of the Main Boundary Fault. To this may be added the most recent stage: (7) the morphotectonic rise of the Himalaya during the last 2–3 Ma (Gansser, 1964) (Fig. 20B).

Seismic tomography provides a snapshot of the 3-D cross-sectional structure of the crust and upper mantle beneath the Himalaya and the southern Tibetan Plateau as expressed by its seismic properties and has been combined with plate tectonic models to reconstruct the subsurface structure at the India–Asia collision zone. The tomographic image indicates that the Indian lithospheric slab has been subducted horizontally beneath nearly the entire Tibetan Plateau to depths of 165–260 km (Kind et al., 2002; Zhou and Murphy, 2005; Li et al., 2008; Nabalek et al., 2009; Replumaz et al., 2010) (Fig. 20A). The temporal progression of deformational episodes of the Himalaya during the past 50 million years soon after the India–Asia collision is synthesized below.

In post-collisional stages, two broad phases of deformation in the Himalayan orogen can be identified: Eohimalayan phase (Middle Eocene–Late Oligocene) and Neohimalayan phase (Early Miocene–Holocene) (Hodges, 2000). In the Eohimalayan phase, the Indian plate collided with Tibet, but India was too buoyant to be subducted into the mantle. The collision slowed India's northward motion from 20 cm/year to 5 cm/year, but the indenter continued to drive into Asia (Tapponnier et al., 1982). With continuing convergence, an accretionary wedge developed on the leading edge of the Indian plate, which was squeezed up along the ITS zone, grew upward, folded, imbricated, and weakly metamorphosed to form the Tethys Himalaya. The Tethys Himalaya is about a 100-km-wide synclinorium that contains almost complete fossiliferous marine sequences from the Proterozoic to Eocene, which were deposited on the shelf and slope of the Indian continental margin, and squeezed during the collision (Gansser, 1964). The Tethys Himalaya would eventually break away from the rest of the Indian craton by a huge normal fault, the South Tibet Detachment System (STDS) during the Miocene.

The Neohimalayan phase spanning from Early Miocene to Holocene is the most active mountain building event of the Himalaya with the development of a series of south-vergent shortening structures in the form of thrust systems that separate the Higher Himalayan, Lesser Himalayan, and Sub-Himalayan zones from one another (Gansser, 1964; Hodges, 2000). As the collision continued, India's northward motion was taken up along a new thrust fault, the Main Central Thrust (MCT) south of the ITS, where a slice of Indian crust and shelf sediments began to wedge beneath Tibet and stacked on the oncoming subcontinent (Molnar and Tapponnier, 1977; Molnar et al., 1993). Movement along this fault thickened the Tibetan crust, deformed the accretionary wedge of the Tethyan Himalaya, and created mountainous terrane of the Higher Himalaya, which was formed from overthrust slices of the old Indian craton, stacked one atop another. The widespread folding and thrusting of older rock on top of younger rock that occurred in making the Himalaya attest to horizontal compression and thickening of the crust.

Continued convergence led to the separation of the Tibetan Himalaya from the northern margin of Indian craton along the South Tibetan Detachment Thrust (STDS). The genesis of the STDS is highly controversial because unlike other thrust fault systems in the Himalaya, the STDS is normal, moved northward relative to the rocks below in opposite direction, and is the stimulus for the "channel flow" theory. Such fault systems, which are known as detachment systems, are common in extensional tectonics where the crust spreads and thins such as the Basin and Range province of North America but unusual in collision zone tectonics (Bird, 1991; Hodges, 2006a). As the Tibetan Himalaya consisting of sediments and underlying continental crust broke away from India toward the north, the remaining Indian craton then slid beneath the margin of Asia for at least 100 km along a huge thrust fault, the Main Central Thrust (MCT).

The presence of South Tibetan Detachment System (STDS) is anomalous in the collision zone of the Himalaya. The basic architecture of the Himalayan orogen is dominated by a series of south vergent thrust fault systems such as MCT, MBT, and MFT as expected in a collision zone (Gansser, 1966). However, STDS is a normal fault, and its relationship with MCT is complex (Burchfiel et al., 1992) and led to the development of the channel flow-extrusion model in Himalayan orogeny (Bird, 1991; Beaumont et al., 2001; Hodges, 2006a, 2006b). The average thickness of continental crust is about 30 km, but Tibetan crust is about 70 km thick with a deeper root in the underlying hot, dense mantle as a result of subhorizontal subduction and heating of the Indian lithosphere that is separated from the Tibetan lithosphere by a thin channel of molten material (Zhou and Murphy, 2005). Various geophysical observations have been interpreted as evidence for a channel of weak, partially molten, middle crust beneath southern Tibet sandwiched between two rigid layers that flow southward, directly toward the Himalayan mountain front. Evidence for the existence of such melts comes from high geothermal gradient and INDEPTH seismic data (Harris, 2007). According to channel flow model, as the Indian plate subducted horizontally below Tibet during Early Miocene along the MCT, the lower crust of the overriding southern Tibet was sufficiently hot and thick to enable lateral flow. To the north and east this flow resulted in the expansion of the Tibetan Plateau. To the south, the lower crustal channel flowed directly toward the Himalayan mountain front in the Higher Himalayan sector pushing it to the surface. This is possible because two shear zones with opposite sense of movement border the Higher Himalaya: in the north by the STDS normal faulting, and in the south by the MCT thrust faulting. Both fault systems dip shallowly northward, such that the Higher Himalayan sequence forms an inclined channel bound on top and at the base by the two fault systems (Fig. 20A).

The STDS and MCT shear zones were simultaneously active, and together they accommodated the southward extrusion and rapid uplift of the Greater Himalaya. Because both STDS and MCT were coeval and active during Early Miocene (22–16 Ma), and because the Higher Himalaya contains high-grade crystalline sheared rocks, it has been interpreted as the surface manifestation of viscous channel flowing southward through the lower/middle crust of Tibet (Godin et al., 2006). In the channel flow model, the increased erosion of the high-elevation of the Higher Himalaya creates the least resistant pathway to the surface for the flowing crustal channel below the orogenic system. As more erosion occurs, the area becomes less and less resistant to the flow of crustal material to the surface (Fig. 20A).

Crustal compression, folding, and thrusting continued throughout the Neogene Period, with widespread metamorphism and tectonism along the Higher Himalaya. The paleomagnetic record suggests that in the Late Miocene India has continued to impinge into Tibet with a significant counterclockwise rotation (Gansser, 1964; Carey, 1976; Patriat and Achache, 1984; Klootwjik et al., 1985). The rotation about the Nanga Parbat syntaxis has produced spectacular Punjab orocline in the Sulaiman–Salt Range on the western Himalayan region (Fig. 11). Indian continental crust began to underthrust Tibet as convergence continued. The Nanga Parbat syntaxis formed a pivot as India moved anticlockwise while Greater India subducted along the Indus–Tsangpo Suture. Rotational underthrusting and oroclinal bending led to the disappearance of Greater India by continental subduction. Most likely the collision took place at different times along the ITS from west to east like the closing of a zipper because of the anticlockwise motion (>20°) of the Indian plate (Fig. 20B).

The fourth phase of movement of India started in the Late Miocene (~10 Ma). Continuing convergence of India resulted in the oroclinal bending of the Himalaya, and the compression was accommodated along a weaker, more southerly zone of Main Boundary thrust (MBT) to form the Sub-Himalayan molassic sediments. The Sub-Himalaya forms the foothills of the Himalayan Range and essentially consists of Miocene to Pleistocene molasse deposits derived by erosion of the Himalaya. The erosion history of the Himalayan belt is reflected in a nearly continuous deposition of sediments in a foreland trough stretching from northeast India to northwest Pakistan, and in distal counterpart, the Indus and Bengal fans (Burbank, 1996). These sediments, known as the Siwalik Group, preserve a continuous record of deposition by rivers ancestral to the modern Indus, Ganga, and Brahmaputra. The Siwalik sediments consist of conglomerates, sandstones, and shales with a thickness more than 5000 m, ranging in age from Middle Miocene (~16 Ma) to Lower Pleistocene (~5 Ma) and provide a record of the orographic evolution of the Himalaya through time. The Siwalik Group was incorporated in the Sub-Himalayan thrust belt and is divided into three informal members. Lower, Middle, and Upper.

Fossil vertebrates from the Siwalik Group provide a critical window of ecology, environment, and vertebrate life along the foothills of the Himalaya. The Siwalik vertebrates are one of the longest and richest sequences of terrestrial vertebrate faunas known. They are famous for spectacular mammalian vertebrates including tiny rodents to early elephants and Asian apes such as *Ramapithecus*, *Sivapithecus*, and *Gigantopithecus*. Typically, vertebrate fossils occur as dense aggregates of disarticulated elements, hundreds or even thousands in number, filling small channels within braided river systems.

Finally, with continuing orogeny, the Sub-Himalaya is thrust over the Holocene Ganga Basin along the Main Frontal Thrust (MFT). The underthrusting of the Indian lithosphere along the Main Frontal Thrust below Tibet indicates the initiation of the latest zone of weakness to accommodate the continuing convergence of India (Fig. 20A). The last phase of the Himalayan orogeny is the morphotectonic rise that began with mostly vertical uplift probably due to isostatic adjustment (Gansser, 1964). The Himalaya is undergoing rapid uplift and is consequently experiencing rapid erosion, with a thick terrigenous sequence deposited in the Sub-Himalaya. Wang et al. (1982) estimated differential uplift between the Tibet Plateau and the Himalaya—the annual rate of uplift of the Himalaya is 0.5 mm to 4 mm/year, greater than that of Tibet. The difference in uplift rates indicates that the Himalaya was lower in elevation than Tibet some 2 Ma. This is evident as many antecedent rivers flow southward across the main ridge of the Himalaya. Recent estimates of Himalayan uplift derived from the pattern of fluvial incision vary in different sectors of the Himalaya, ranging from 4 to 8 mm/year in the Lesser Himalaya to 10–15 mm/year in the Subhimalaya, where the rate of erosion is found to closely mimic uplift (Láve and Avouac, 2001).

The relentless push of India toward Asia is baffling. What is the force behind this continued convergence of the Indian plate? The subduction of the Indian plate, the slab pull, is generally considered as the main force for the continuing northward motion of the Indian plate (Capitano et al., 2010). Subduction might have played a significant role in underthrusting and convergence of the Indian plate. As the Indian and Eurasian plates collide, the Indian upper crust is scrapped off at the Himalayan front (Fig. 20A); the lower crust slides under the Tibetan crust along the MFT, while the upper mantle peels away from the crust and drops down in a diffuse manner. High resolution seismological data reveal that the lower part of the Indian lithosphere along the Main Frontal Thrust underplates the Himalaya and Tibet up to 31°N and almost extends to the middle part of the Lhasa block Nabalek et al., 2009). Such a dense continental slab with higher density could subduct under Tibet and provides a significant driving force for India–Asia convergence. The slab pull is reinforced by the ridge-push from the spreading of the Central Indian Ridge and Carlsberg Ridge (Allègre et al., 1984).

9.4. Tectonic evolution of the Tibetan Plateau

The Tibetan Plateau, bounded on north by the Kunlun Shan and in the south by the Himalaya, is an immense upland plateau averaging more than 5 km in elevation. The elevation history of the world's highest plateau offers critical insights into the geodynamics of continental collision. The Tibetan Plateau is a collage of a number of exotic continental terranes that were successively accreted to Asia in Mesozoic time prior to its collision with India (Allègre, 1988). Paleomagnetic data indicate that these continental terranes were part of Gondwana during the Paleozoic. The suture zones between different terranes in the Tibetan Plateau are marked by discontinuous exposures of ophiolites and ophiolitic mélanges. The Indus–Tsangpo Suture Zone separates the southernmost terrane of the Tibetan Plateau (Lhasa terrane) from the Himalaya and the Indian plate. Farther north of the ITS, three more sutures can be seen: the Bangong–Nuijang Suture separating Lhasa from Qiangtang terrane, the Jinshajiang Suture separating Qiangtang from Songpan–Ganze terrane, and Kunlun Suture separating Songpan–Ganze from Kunlun terrane (Fig. 11).

The evolution of Tibetan Plateau involves subduction of Indian lithosphere, thickening of the Tibetan crust, and eastward extension of Tibetan lithosphere by indention tectonics (Tapponnier et al., 1982; Royden et al., 2008). The development of high topography and thickened crust in the Tibetan region began during subduction of the Neotethys Ocean floor northward beneath Asia with Andean-type magmatism represented by the Gangdese Batholith (Fig. 19B). During the post-collisional orogeny with the closure of Neotethys, as the Indian continental plate began to subduct beneath Asia, crustal shortening formed in central and southern Tibet with the development of a series of south-vergent thrust faults or subduction zones. Uplift of the plateau accelerated in the Early Miocene and it probably reached its present elevation by about 8 Ma (Molnar, 1989). Continental subduction of the Indian plate beneath Asia has played a key role in the tectonic evolution of Tibetan Plateau, but the way in which it grew vertically and laterally has been a matter of debate (Tapponnier et al., 1986). Large negative Bougerer anomalies, coupled with seismic studies support the long standing hypothesis that the Tibetan Plateau is underlain by a crust averaging 70 km thick, or twice the normal thickness of the continental crust (Gansser, 1964; Powell and Conaghan, 1973; Molnar, 1989; Kind et al., 2002; Nabalek et al., 2009).

The great height of Tibet is a consequence of underthrusting of India beneath Tibet. Large scale underthrusting of Greater India beneath Asia has been postulated to explain double thickness of Tibetan Plateau on geological and geophysical grounds (Gansser, 1964; Powell and Conaghan, 1973; Powell, 1986). Patriat and Achache (1984) suggested 400 km, Klootwjik et al. (1985) at least 650 km, and Powell (1986) more than 500 km of crustal shortening from paleomagnetic evidence because of underthrusting. Recent investigations of seismic and tomographic images support underthrusting of Indian continental lithosphere beneath Tibet subhorizontally about 500 km about the middle of the plateau, sinking down to the Earth's mantle to a depth of at least 200 km (Kind et al., 2002; Nabalek et al., 2009).

During the Eocene/Oligocene time (~40 Ma), the advancing Indian plate, acting as a rigid indenter, forced Tibet out of its way (Molnar and Tapponnier, 1977; Tapponnier et al., 1982, 1986, 1990). In the past 10–15 Myr, as the Himalayan–Tibetan Plateau began to rise, lateral extrusion gradually became the dominant mechanism to accommodate the India–Asia collision and crustal shortening. Major strike-slip faults seem to have played a leading role in accommodating India's continued penetration as an indenter into Asia. The majestic Tibetan Plateau, which rises about 5 km high, and has earned the title "roof of the world," plays a crucial role in absorbing the continuing push from India into Asia by sliding to the east and out of India's path. Eastern Tibet is traditionally interpreted as being part of a broad accommodation zone. Avouc and Tapponnier (1993) suggested that as much as 50% of the convergence between India and Asia has been absorbed by extrusion of south-eastern Asia. Others disagree about the role of lateral extrusion; they believed that lateral extrusion has not contributed more than 250 km shortening to the India–Asia convergence (Dewey et al., 1989).

After collision (~50 Ma), the crust was shortened in western and central Tibet along the Gangdese Thrust system, while large fragments of lithosphere moved from the collision zone toward western Pacific and Indonesia. On the north side of the Tibetan Plateau, where the continental collision reaches the farthest, the Tibetan Plateau is compressing the adjacent blocks northward and eastward resulting in a series of major active faults. Molnar and Tapponnier (1977) recognized various east–west trending strike-slip faults such as Altyn Tagh and Red River faults north of the Himalaya in Central China from earthquake seismology and satellite imagery (Fig. 21). Much of the collisional shortening, which is not absorbed in the Himalaya or underthrusting beneath Tibet, apparently is converted into horizontal movements along these faults, predominantly to the east.

These left-lateral faults operated sequentially at two stages during the Cenozoic to absorb the push from India (Tapponnier et al., 1986, 1990). In the first stage, Indochina escaped along the Red River Fault, which can be traced from southeastern Tibet through Yunnan to the South China Sea. The Asian blocks west of 70°E remained more or less undeformed but the Indochina block pushed to the east at least 500 km southeastwards relative to South China during this stage. Extrusion of the Indochina block during the Indo-China collision has long been speculated on the basis of laboratory experiments with plasticine (Tapponnier et al., 1982).

The movement of the Red River Fault began soon after the initial collision (50 Ma), but stopped around 17 Ma. Cenozoic structures in northern Tibet are dominated by the Altyn Tagh Fault system that bounds more than 1200 km and bounds the rather rigid Tarim basin to the northwest and the more deformable Tibetan Plateau to the southeast. India has been converging with Tibet since the cessation of motion along the Red River Fault. The second strike-slip movement, measuring 3 cm/year, began along the Altyn Tagh Fault approximately 17 Ma. As a

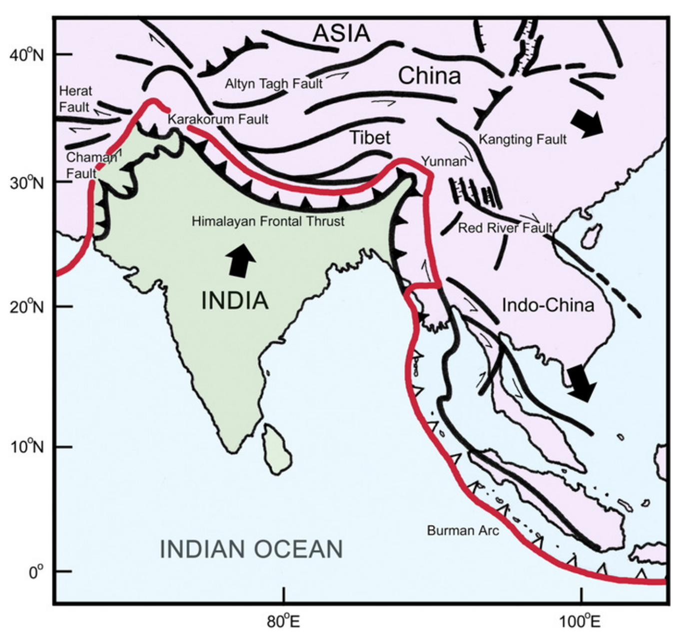

Fig. 21. Schematic map illustrating extrusion tectonics in Tibet and Indochina by two great strike-slip faults, the Red River and Altyn Tagh that allowed Indochina and Tibet to slide eastward; the complex plate boundary of India is demarcated by a red line.
After Tapponnier et al. (1986).

result, the Tibetan Plateau is slowly being pushed to the east toward the Pacific. It has been estimated that about a 270 km north–south shortening has been accommodated along the Altyn Tagh Fault, causing Tibet to slip out dextrally along the fault line.

Because the crust of Tibetan Plateau is twice the normal thickness of the continental crust, it has a deep root that floats on underlying hot and dense mantle. The bottom layer of lightweight crust is unusually hot and soft and flows easily. Channel flow model has been used to explain the orogeny and uplift of the Tibetan Plateau (Beaumont et al., 2001; Hodges, 2006a). As discussed earlier, the middle and lower crust beneath the Tibetan Plateau might be partially molten, which may explain why the plateau is so flat. The Tibetan fluid layer of the middle crust is sufficiently weak to support large variations of surface topography and it simply flows laterally under the influence of gravity (Bird, 1991; Hodges, 2006a).

An important prediction of the channel flow theory is that the uplift of the Tibetan Plateau would develop initially during the Early Miocene time proximal to the ITS zone. Since then, the eastern margin of Tibetan Plateau is marked by elevation and crustal thickness gradients and has grown eastward. Most likely, the channel flow had spread laterally to the north and east, and resulted in the topographic rise and expansion of the Tibetan Plateau. Unlike Higher Himalaya, there is little evidence of this channel flow in the surface structural geology of Tibet (Beaumont et al., 2001; Hodges, 2006a). However, surface wave tomography and receiver function analysis from the eastern and south central plateau of Tibet supports the presence of the middle and lower crustal network of hot channel flow (Kind et al., 2002; Royden et al., 2008; Nabalek et al., 2009).

A recent model relates the uplift, growth, and crustal shortening of the Tibetan Plateau to a plate-like behavior including sequential intracontinental subduction zones along the three main suture zones, namely the Bangong–Nuijang Suture, Jinshaijang Suture, and Kunlun Suture (Fig. 11) (Tapponnier et al., 2001). These south-vergent subduction zones in Tibet and north-vergent thrust zones of the Himalaya occur on either side of the Indus–Tsangpo Suture Zone, both represent post-collisional convergent deformation (Fig. 20B). In Tibet, they decrease in age northward ranging from Eocene to Pliocene and may accommodate the crustal shortening after the collision. These slabs of Asian mantle subducted en echelon under ranges north of the Himalaya. Subduction was oblique and accompanied by lateral extrusion of SE Asia, facilitated by the lateral flow of the middle and lower fluid crust of Tibet.

10. Cenozoic climate

From Paleocene to Early Eocene the Indian plate moved northward and entered the equatorial humid belt. The continuing erosion of the Deccan Traps removed much of the CO_2 from the atmosphere, thus starting a cooling trend over the Eocene–Oligocene boundary (Kent and Muttoni, 2008). During the Oligocene (~30 Ma), the Indian subcontinent was very close to its present day position. The Tethys Himalaya rose, and the center of the subcontinent was located at 10°N. The southern part of the subcontinent was in the Equatorial Rainy Belt whereas the northern part was extended into the Subtropical Arid Belt (Fig. 22A). Annual mean temperature ranged between 27 °C (southern part) and −5 °C (northern part) (Scotese et al., 2007, 2008, 2009, 2011; Goswami, 2011). Though the northern areas along the line of collision were experiencing a drier climatic condition, the peninsular part was having significant precipitation of about 10 cm/month to 26 cm/month. It is important to mention that the southern tip and the mid-section of eastern Indian peninsula were relatively dry (6–8 cm/month). The collision of India with Asia and the resulting disappearance of Tethys may have provoked strong global warming.

The uplift of the Himalayan–Tibetan Plateau with an average elevation of 5 km above the snow line forms a climatic divide between

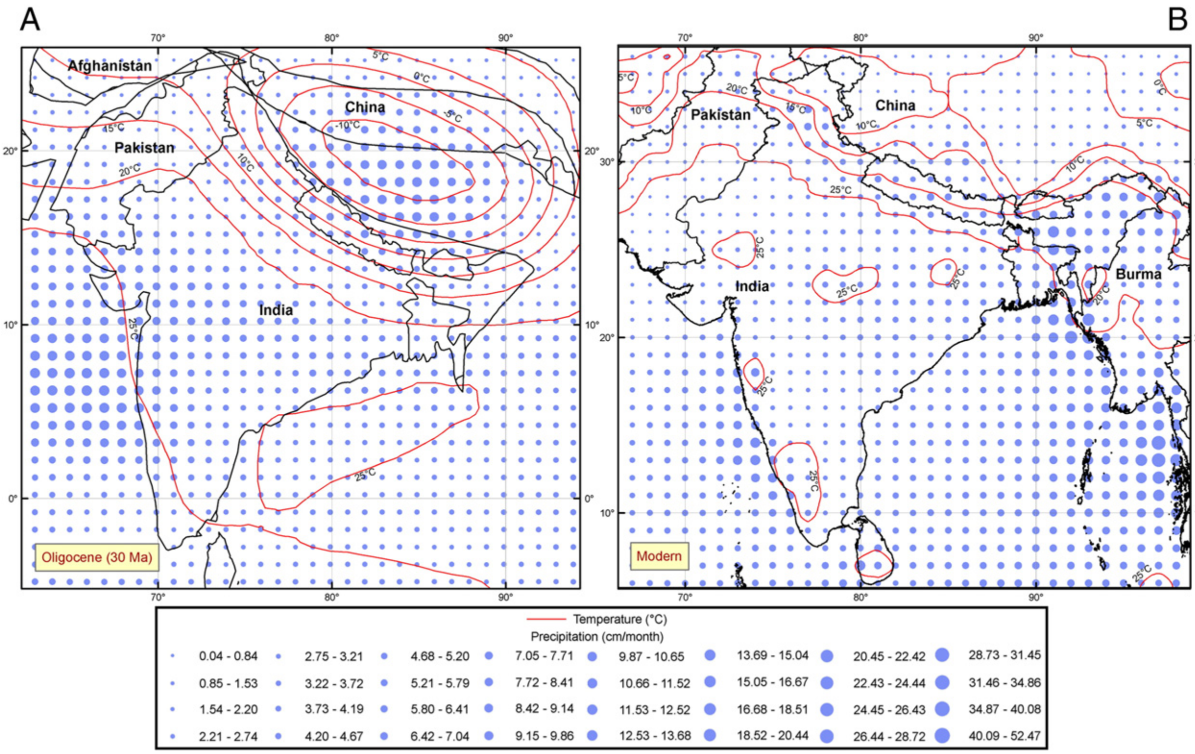

Fig. 22. (A and B) Mean annual precipitation and mean annual temperature maps of Indian plate during the Oligocene and Holocene periods.

India and Central Asia. The climatic effects of Himalaya–Tibet orographic barrier are multifaceted and complex. In addition to providing a barrier to atmospheric flow, mountains and plateaus produce rain shadows on their leeward slope and can have important effects on albedo and energy balance. The Himalaya and Tibetan Plateau may have been an important contributor to long-term climatic change for Cenozoic cooling and Northern Hemispheric glaciation (Hay et al., 2002). Atmospheric circulation in the absence of the Himalayan–Tibetan orogen—the "Third Pole"—would be very different from what it is today: the modern monsoon climate would not exist.

The fast-rising Himalayan–Tibetan Plateau must have become high to disrupt global circulation of the westerly winds, so that climates thousands of kilometers east and west are affected (Molnar, 1989). Air moving from west to east, in general circulation of the atmosphere, go over smoothly if there is no topographic barrier. Before uplift of the Himalaya–Tibetan Plateau in the Paleocene, westerly winds would pass directly across the India–Asia landscape, the axis of flow shifting seasonally north and south. As the Himalayan–Tibetan Plateau uplifted, the westerly flow was diverted around the high plateau, a winter high-pressure system develops over the snow-covered plateau and cold, dry winds blow clockwise off the upland region. The uplift not only disrupted the west-to-east circulation, it brought the low-pressure area southward over India. The uplift of the Himalayan and Tibetan Plateau at an elevation averaging ~5 km is the primary factor in the present monsoon circulation.

India's continuing convergence to Asia was a major driver for the uplift history of the Himalayan–Tibetan orogen and the onset of Asian monsoon during the late Cenozoic. Uplift of this region began about 50 Ma during the initial collision, and intensified during the Early Miocene (~23 Ma) during the inception of the Asian monsoon that reached a peak 10 Ma, and has subsequently slowed down. Rapid uplift of the surface topography of the Higher Himalaya coincides with a substantial increase in precipitation and erosion of rising mountains as revealed from deep-sea sediments of the Bay of Bengal and the Arabian Sea (Clift et al., 2008).

The channel flow extrusion of the Higher Himalaya in the Early Miocene has been linked to the onset of monsoon (Hodges, 2006b). The summer monsoons strike the Higher Himalaya with abundant rainfall, causing erosion, enabling the fluid lower crust of Tibet to extrude toward the range front. The Higher Himalaya probably reached the maximum height possible because erosion accelerates with height. A feedback loop exists where more extrusion creates higher topography in the Himalayan front that in turn allows more precipitation and erosion to occur, maintaining a steep range front.

The South Asian Monsoon system is one of the most important and influential of the Earth's major climate systems. A monsoon-type climate with warm and wet summer and cold and dry winter might have commenced during the morphotectonic rise of the Himalaya that formed a climatic barrier between India and Asia. The onset of the monsoons may have been triggered when the Tibetan Plateau reached a threshold height of 2 to 3 km about 8 Ma (Clift et al., 2008). The rise of the Himalaya and Tibetan Plateau enhanced both the winter and summer Asian monsoons and gave rise to a drying trend in Central Asia.

Monsoons are strong onshore and offshore winds that are caused by the difference in heat capacity between land and water. In winter the Himalaya and Tibetan Plateau become cold, and they cool the air above them. This cool, dense air descends and flows seaward as winter monsoons displacing warmer air above the ocean. In summer the mountains and plateaus warm up, and the air above them rises. The rising air is replaced by strong winds, the summer monsoons, from the Indian Ocean. Sweeping northward from the Bay of Bengal, the Indian monsoon strikes the eastern Himalaya with full force causing tropical depression and torrential rain.

The cause of the Pleistocene Ice Age remains controversial. It is generally believed that during the morphotectonic phase, Himalaya acquired the present relief during the Pleistocene Period (~2.6 to 0.01 Ma) (Gansser, 1964). As the Himalaya jutted skyward, CO_2, an important greenhouse gas, was removed from the atmosphere by the carbon cycle. As a result, climatic conditions grew cold, spawning a period of glaciation along the mountain ranges, resulting in albedo-induced heat losses (Hay et al., 2002; Kuhle, 2002). Both high altitude and availability of moisture from the monsoon favored the initiation of glaciation in the Himalayan–Tibetan Plateau. The glaciers spread far and wide in the Potwar, Kashmir, Ladakh, and Tibetan regions. New data show that the glaciations were not synchronous throughout the region. In some areas they were most extensive during 60–32 Ka and in other places during the global maximum at 18–20 Ka (Benn and Owens, 1998). After the end of the Pleistocene ice age around 20 Ka, the southwest monsoon weakened its intensity, bringing about climate change. Heavy rains gave away to spells of cold-dry conditions. As the temperature became warmer, there was oscillation of dry–cold and wet–warm climates. The melting of the ice sheet covering the Himalayan ranges and Tibetan Plateau might have contributed a large amount of melt water to the oceans and a significant rise of the eustatic sea level. The Himalaya and Tibet are the most glaciated regions outside the polar regions. These glaciers are the source of innumerable rivers that flow across the Indus–Ganga Plain and the Central and East China.

The uplift of the Himalaya continues today, as evident from active seismicity along the major boundary-thrust faults and dislocation and deformation of Pleistocene and Holocene landforms and deposits. The very fast uplift in the northwestern part of the Himalaya is manifest in rapid erosion, about 2–12 mm/year, which is determined on the basis of different bedrock uplift and incision made by the Indus River (Burbank, 1996). The Himalaya is so high and the monsoon rainfall upon it is so immense that erosion is very extensive. In the Indian plains, as the intensity of the monsoon rains increased, the rapid uplift of the Himalaya induced rapid erosion of the mountains, and the vast amount of sediments were carried downward by two glacier-fed rivers, the Indus on the west, and the Ganga–Brahmaputra in the east. These two rivers dumped sediments to the Indian Ocean that created two great deep-sea fans, the Bengal and the Indus.

The Himalaya forms a formidable rain shadow to the monsoon, preventing moisture from reaching to Tibetan Plateau, which helped create the Gobi and Mongolian deserts. This drying trend created enhanced aridity in the Asia interior and was manifested in China about 2.5 million years ago in the form of extensive eolian deposits. Fine dust, eroded by strong winds from desert basins north of the Himalaya, began accumulating widely over central and eastern China (Zhisheng et al., 2011).

Today, the majority of the peninsular India falls within the Equatorial Rainy Belt, whereas the central and western part is in the Subtropical Arid Belt (Fig. 22B). The Indian plate at its present location, with the center at 22°N latitude, extends through three broad latitudinal climatic belts (Figs. 22B, 23). The northern part is representative of the Warm Temperate Belt climate.

The modern climatic condition of the Indian Subcontinent is greatly influenced by the monsoon. The western part of the Indian peninsula (up to 25 cm/month) and eastern India (up to 30 cm/month) has the maximum annual precipitation (Legates and Willmott, 1990a). The central part has 5–10 cm/month annual precipitation whereas the northern part gets up to 15 cm/month (Fig. 23). Except for the northern part of the Indian subcontinent, the mean annual temperature (Legates and Willmott, 1990b) ranges between 29 °C and 23 °C. In the northern part the annual mean temperature ranges 15 °C to 20 °C.

11. Conclusion

The tectonic evolution of the Indian plate, which represents one of the most remarkable journeys of all the continents (about 9000 km in 160 million years), is largely the story of the breakup of Gondwana through time and its subsequent collision with the Kohistan–Ladakh Arc and Asia. The tectonic history of India plays a central role in the dispersal of Gondwana. Though complex, this story can now be told using evidence from linear magnetic anomalies, paleomagnetism and hot spot tracks. Information from different fields of continental geology and deep sea drilling also provide useful information to constrain India's journey. Both the eastern and western coasts of India are

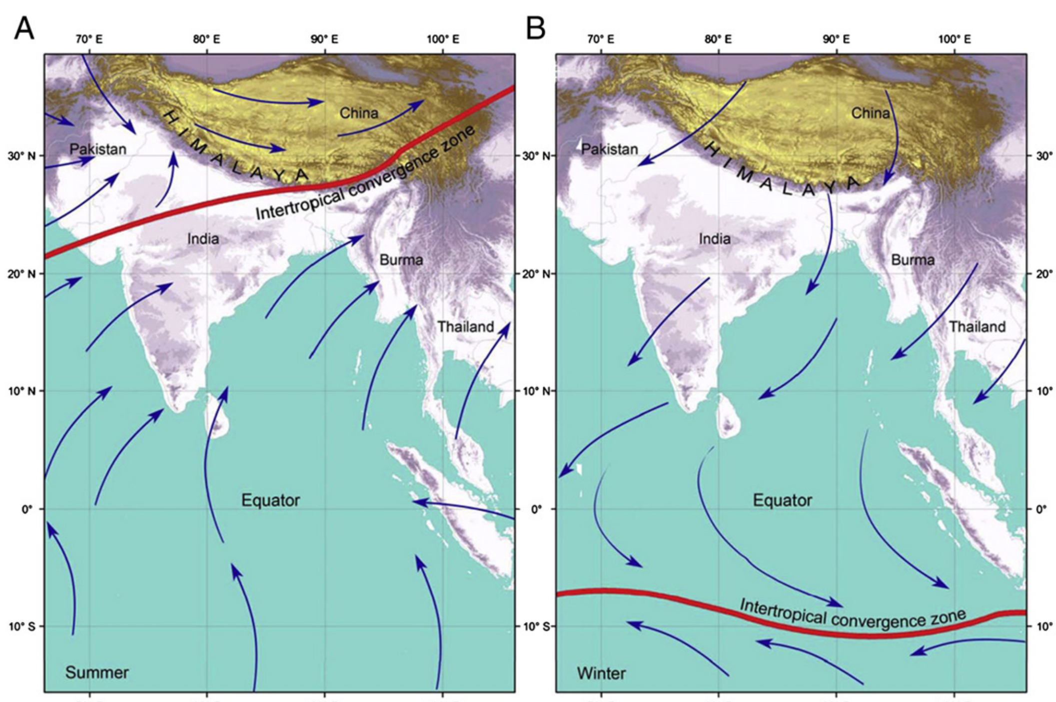

Fig. 23. Linking rise of the Himalaya–Tibetan Plateau and the birth of the monsoon. India's climate is dominated by monsoon. Monsoons blow from the land toward sea in winter, and from the sea toward land in the summer. (A) During the summer, the winds carry moisture from the Indian Ocean and bring heavy rains from June to September. (B) During the winter the monsoon winds blow from the northeast and carry little moisture. The temperature is high because the Himalaya forms a barrier that prevents cold air from passing into the subcontinent.

rifted margins, characterized by sub-aerial volcanic rocks. The age of flood basalts along the continental rift margins help to constrain the timing of major tectonic events.

The present day configuration of passive margins of the Indian peninsula is the consequence of seven episodes of continental flood volcanism and sequential rifting events since the Jurassic. By reviewing spatial association of flood basalts, mantle plumes, and rifting events of the Indian plate through time and space and reconstructing the thermal history of the evolving continental margins, these events can be summarized as:

a) Separation of East Gondwana from West Gondwana during Middle Jurassic time (Karoo–Ferrar flood basalts, Bouvet plume, 182 Ma),
b) India separated from Antarctica–Australia during the Late Cretaceous (Kerguelen–Rajmahal basalts, Kerguelen plume, 118 Ma),
c) India separated from Madagascar during the Late Cretaceous (Morondava–St. May basalts, Marion plume, 88 Ma),
d) An island arc, the Kohistan–Ladakh Arc, formed on the northern side of the Indothethys and became trapped between India and Asia. This allowed for two stages of deformation. The first consisted of a collision of the arc with India, followed by the collision with Asia. India rapidly drifted northward as an island continent across the Indotethys and collided with the Kohistan–Ladakh arc in Late Cretaceous (~85 Ma),
e) The eruption of the Deccan traps as the Reunion plume breaches the lithosphere (65 Ma),
f) The breakup of the western margin of the Indian subcontinent continued in the Late Cretaceous as the Seychelles separated from Laxmi Ridge and the western Indian margin (70–65 Ma),
g) Eohimalayan phase: India–Asia collision began in Early Eocene along the Shyok–Tsangpo Suture Zone, which later transferred to the Indus–Tsangpo Suture Zone with the closure of the Neotethys (~50 Ma).
h) Neohimalayan phase: the architecture of the Himalayan–Tibetan orogen is dominated by deformational structures developed in the Neogene Period (<23 Ma), dominated by the channel flow extrusion of the higher Himalaya with the onset of Asian monsoon. India–Asia collision and intracontinental deformation resulted in large crustal shortening and uplift of the Himalaya and Tibetan Plateau above the snow line that might have triggered the Pleistocene glaciation in Northern Hemisphere.

During its tectonic evolution from Gondwana to Asia, India became smaller and smaller, trimming its rifted continental margins, and leaving behind several continental terrances such as Madagascar, Seychelles, and Laxmi Ridge, stranded in oceanic crust as India drifted away from Africa opening the Indian Ocean. With the complete subduction of Neotethys at the Shyok–Tsangpo Suture, the northern one-third part of the Indian plate was subducted beneath Asia during the initial phases of collision (50 Ma), and now lies buried beneath the Tibetan Plateau.

Starting northward at a rate of 3–5 cm/year during the Mesozoic, the Indian plate attained a sudden acceleration of 20 cm/year from Late Cretaceous (~67 Ma) at the KT boundary, which has been a major tectonic puzzle. This accelerated northward migration of the Indian plate during the Late Cretaceous–Early Eocene makes India unique among the fragments of Gondwana. India slowed down considerably 5 cm/year at Early Eocene (~50 Ma) during its convergence with Asia and the closure of the Neotethys. India–Asia collision is marked by indentation, channel flow extrusion of Higher Himalaya, rotational underthrusting, and oroclinal bending of the Himalaya. With the rise of the Himalaya and Tibetan Plateau, the onset of the Asian monsoon began dramatically, changing the climatic regime of Asia. During its long northward journey, the climatic and biotic regime of the Indian plate changed gradually with the onset of monsoon and Pleistocene glaciation during the Neohimalayan phase.

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Appendix A. Supplementary material

Supplementary materials related to this article can be found online at 10.1016/j.gr.2012.07.001.

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