CHAPTER 2
Chapter - 2

GEOLOGICAL FRAMEWORK

2.1 Introduction

The continental margins off west coast of India, Seychelles and east coast of Madagascar are believed to have been evolved by rifting and successive drifting of India, Seychelles and Madagascar. The present day geological scenario of the deep offshore regions around these continental margins is a consequence of this rift, drift and subsequent tectonic and sedimentary evolution. Various studies carried out so far have revealed the existing regional structural and sedimentary features of these areas and allowed proposing broad models of their evolution. In order to view the results of the present study in the background of these existing geological knowledge, relevant available information about structure and tectonics of various features of the areas in the deep offshore regions adjacent to west coast of India as well its conjugate Madagascar and Seychelles, have been collated in this chapter. Further, as the western continental margin of India was classified as a passive continental margin (Biswas, 1982, 1987), therefore in order to appreciate the passive continental margin setting, a brief account of the concept of continental margin in general and evolution of the passive continental margin in particular have also been presented in this chapter.

2.2 Tectonic elements of the western continental margin of India and the adjoining land and oceanic areas

Marine geoscientific investigations on the continental shelf, slope and deep offshore regions off west coast of India have so far revealed the existence of several physiographic, sedimentary and basement features. Many parts of the continental shelf have been investigated in detail in connection with hydrocarbon exploration but most of these results are not available in public domain, whereas, not many studies were carried out in the deep offshore regions. In following paragraphs the information about the regional features of this area have been presented.
2.2.1 Features on the western part of the Indian mainland

The Indian subcontinent is made up of three major physiographic provinces (Fig. 2.1), the Himalayas, the Indo-Gangetic Plain and the Peninsular Shield, all of which are characterized by distinctive geomorphological, structural, stratigraphic and deep crustal features (Mahadevan, 1994). The Peninsular Shield comprises of Precambrian terrains of South India and the Deccan Trap volcanic province. The Precambrian terrains in South India are further divided into Southern Granulite Terrain (SGT), Eastern Ghat Mobile Belt (EGMB), Western Dharwar Craton (WDC) and Eastern Dharwar Craton (EDC). Since the study area is adjacent to the western part of India, emphasis for description is given to the Southern Granulite Terrain, Western Dharwar Craton and Deccan Trap, which form a major part of the western Indian mainland.

The Southern Granulite Terrain (SGT) of India (Fig. 2.1) is one of the few terrains in the world that has preserved Archean crust with extensive granulites, believed to be of lower-crustal origin (Anand and Rajaram, 2003). The SGT primarily consists of several granulite blocks, which are dissected by a number of shear zones. The most prominent shear zones are the Achankovil Shear Zone, the Palghat-Cauvery Shear Zone and the Moyar Shear Zone (Kroner and Brown, 2005). The Achankovil Shear Zone is a 10-20 km wide and 100 km long NW-SE trending dextral shear zone bounding the Madurai Block in the north and Kerala-Khondalite Belt (KKB) in the south. Some authors considered that the Palghat-Cauvery Shear Zone and Moyar Shear Zone together form a 100-150 km wide shear zone system of which the Palghat-Cauvery Shear Zone is the southern one. These shear zones are important in the context of India-Madagascar pre-drift juxtaposition, as they are often considered to be conjugate of several comparable shear zones in the Madagascar. The Western Dharwar Craton (WDC), which is composed of Dharwar type schist belt, is bounded (Fig. 2.1) to the south by the Southern Granulite Terrain (SGT), to the east by the Eastern Dharwar Craton (EDC) and to the northwest by the Deccan Trap flows. The most characteristic feature of the Archean cover sequence of the Dharwar craton is their arcuate NS trend with convexity towards east. A major portion of the deeper geology of western India lies hidden under the Deccan Traps (Fig. 2.1), which is one of the largest known continental flood basalt (CFB) provinces on the Earth.
Fig. 2.1 Detailed geological map of India (modified after GSI, 1999 as reproduced in Naqvi, 2005). SGT: Southern Granulite Terrain; EGMB: Eastern Ghat Mobile Belt; WDC: Western Dharwar Craton; EDC: Eastern Dharwar Craton.
The Deccan Trap CFB extend over nearly 500,000 sq. km. and lie flat as horizontal sheets and are believed to have erupted sub-aerially through fissures in the Earth's crust (Kumar, 1986). A large thermal anomaly, generated by a deep mantle plume (Reunion Hotspot) is commonly postulated to explain the genesis of the Deccan Trap CFB event (Morgan, 1972, 1981; Courtillot et al., 1986; White and McKenzie, 1989; Duncan, 1990; Baksi and Farrar, 1991; Basu et al., 1993). The timing and duration of this volcanic event over the Indian peninsula have been a subject of considerable debate (Alvarez et al., 1980; Baksi, 1988, 1994; Courtillot et al., 1988; Duncan and Pyle, 1988; Rampino and Stothers, 1988; Acton and Gordon, 1989; Vandamme et al., 1991; Negi et al., 1993; Basu et al., 1993) and various postulated ages of their emplacement range between 62-72 Ma. However, views appear to converge on a short duration (1my) peak volcanism around 65 Ma (Courtillot et al., 1986; Vandamme et al., 1991; Hofmann et al., 2000).

Three orogenic trends predominate the Precambrian basement grains of the western part of the Indian peninsula. These are the Dharwar trend, the Delhi-Aravalli fold trend and the Satpura trend (Fig. 2.2). The generally NNW oriented Dharwar trend is the dominant basement grain in the western peninsular India (Biswas, 1982). It is expected that Dharwar trend continue below much of the area now covered by Deccan Trap CFB (Das and Ray, 1976; Krishnan, 1968; Biswas, 1982; Gombos et al., 1995). The Aravalli Delhi Fold trend is a major segment of the NW Indian shield and consists of intensely folded, deformed and metamorphosed rocks of Proterozoic period that are intruded by various granites, basic and ultra-basic rocks (Tewari and Rao, 2003). This NE-SW oriented trend bifurcates into two branches at its southward end; the Delhi trend swings E-W into the Kutch region and NE-SW trending Aravalli trend continues across the Cambay graben into the Kathiawar Peninsula. The ENE-WSW Satpura trend along the Narmada Rift is the other dominant structural fabric of the western India (Biswas, 1982). These three major trends, resulting from orogenesis dating back more than 2000 Ma, were the zones of deformed and weakened crust along which later Phanerozoic rifting was facilitated (Biswas, 1982, 1987; Gombos et al., 1995). Kutch, Cambay and Narmada (Fig. 2.2) are three pericontinental rift basins on the western margin of India, which evolved during different phases of
Fig. 2.2. Major structural and tectonic trends within the study area and adjacent regions. Thin dotted lines are selected bathymetry contours (in metres). Solid annotated triangles are DSDP drill sites with site numbers. Inferred magnetic lineations: L1-L4 in the Laxmi Basin and MAL1-MAL4 in the offshore Indus Basin. Identified magnetic isochrons are labelled with anomaly number and corresponding age in Ma [eg. 24n3 (53.12)]. FZ: Fracture Zone; pf: pseudofaults; R: Raman Seamount; P: Panikkar Seamount; W: Wadia Guyot; BH: Bombay High; LB: Laxmi Basin; LR: Laxmi Ridge; OIB: Offshore Indus Basin; ASZ: Achankovil Shear Zone; PCSZ: Palghat-Cauvery Shear Zone; MSZ: Moyar Shear Zone. Compiled from Biswas (1982), Singh and Lal (1993), Bhattacharya and Chaubey (2001), Bhattacharya et al. (1994a, b), Malod et al. (1997), Chaubey et al. (2002a) and Meissner et al. (2002).
India's geotectonic history during its break-up from Gondwanaland, its northward drift and final collision with Eurasia (Biswas, 1982). These basins were formed by rifting along Precambrian tectonic trends. Interplay of three major Precambrian tectonic trends of western India, Dharwar (NNW-SSE), Aravalli-Delhi (NE-SW) and Satpura (ENE-WSW) controlled the tectonic style of these basins. The geological history of the basins indicates that these basins were formed by sequential reactivation of primordial faults. The Kutch Basin opened up first in the Early Jurassic (rifting was initiated in Late Triassic) along the Delhi trend followed by the Cambay basin in the Early Cretaceous along the Dharwar trend and the Narmada Basin in Late Cretaceous time along the Satpura trend.

One of the major geomorphic features in the western peninsular India adjacent to the study area is the Western Ghats escarpment, which forms the western edge of the mountainous Sahyadri Range. This spectacular Western Ghat escarpment has more than 700 m drop at places (Valdiya, 2001). This feature appears to be interesting in connection with the evolution of the western continental margin because some researchers (Widdowson, 1997; Widdowson and Gunnell, 1999) believe that this feature was formed at the time of rifting of Seychelles from India (~65 Ma) and subsequently retreated by about 120-180 km.

2.2.2 Western continental shelf of India

The western continental shelf of India (Fig.2.2) is considered to be limited by 200 m isobath. Towards north this shelf is relatively wider, being more than 300 km in the areas north off Mumbai coast, whereas towards south this width gradually narrows down to about 50 km off Trivandrum. In contrast to this, the continental slope is narrow in the north but widens towards south (Biswas, 1989).

A series of narrow regional and local horsts and graben structures formed by longitudinal extension faults characterizes the basement trends of the shelf area. The style of faulting is controlled by three major orogenic trends of western part of the Indian mainland. From Kerala offshore to Bombay offshore, the Dharwarian trend (NNW-SSE) predominates, to the north, in the Gulf of Cambay region – the Satpura trend (ENE-WSW) dominates the structural style while
further north in Kutch–Saurashtra region the Aravalli–Delhi trend (NE-SW) is predominant (Biswas, 1989).

The western continental shelf sedimentary basins appear to have been divided into several sub-basins delimited by transverse basement arches or fault bounded highs (Ramaswamy and Rao, 1980; Biswas and Singh, 1988; Biswas, 1989; Nair et al., 1992; Singh and Lal, 1993; Raju et al., 1999). These basins (Fig. 2.2) are Kutch Basin, Surat Basin, Ratnagiri Basin, Konkan Basin and Kerala Basin. The southwesterly plunging Saurashtra Arch separates the Kutch Basin from the Surat Basin. West plunging Bombay Arch occurs between the Surat and Ratnagiri basins. The Konkan Basin is separated from the Ratnagiri Basin by the southwesterly plunging Vengurla Arch and from the southernmost Kerala Basin by southwesterly plunging Tellicherry Arch. The southern extension of the NNW-SSE trending Cambay graben in the offshore area together with a series of parallel horst-graben features cut across the Bombay Arch isolating the Bombay High.

2.2.3 Laxmi Ridge

Laxmi Ridge (Fig. 2.2) is a prominent physiographic feature located in the deep offshore region off west coast of India. This physiographic feature was first observed and named by Naini and Talwani (1977), and subsequently studied by various researchers (Naini and Talwani, 1982; Kolla and Coumes, 1990; Droz and Bellaiche, 1991; Shaynurov and Terekhov, 1991, Miles and Roest, 1993; Pandey et al., 1995; Malod et al., 1997; Chaubey et al., 1998; Miles et al., 1998; Talwani and Reif, 1998, Singh, 1999, 2002; Lane et al., 2003, 2005; Collier et al., 2004a, b, Mishra et al., 2004; Bansal et al., 2005; Krishna et al., 2006). This is an aseismic basement high feature, mostly buried under sediment cover. The average water depth over the ridge is about 2.8 km and has a basement relief of about 2 km (Naini and Talwani, 1982; Droz and Bellaiche, 1991). Even though a positive basement feature, this ridge is associated with a characteristic broad negative free-air gravity anomaly (~50 mgal). It is expressed as NW–SE trending topographic high in the southerly end, while its topographic expression is not discernible northward beyond 18°30’N. However, based on associated characteristic gravity low and the adjacent magnetic anomalies, it was deduced that around 65°30’E this ridge turns WNW-ENE and extends westwards at least
up to 63°40'E (Miles and Roest, 1993). The southward extension of the physiographic expression of the NW-SE trending most prominent segment of the Laxmi Ridge appears to terminate abruptly against an oceanic crust containing east-west trending magnetic lineations (Bhattacharya and Chaubey, 2001).

In absence of direct evidences like drill well information, various authors have inferred nature of the basement of the Laxmi Ridge based on geophysical data. Some researchers (Naini and Talwani, 1982; Kolla and Coumes, 1990; Miles and Roest, 1993) considered this physiographic feature as a continental sliver, forming the boundary between rifted transitional type of crust lying landward (north and east) and the oceanic crust of the Arabian Basin in the south and west. On the other hand consideration of some other studies (Bhattacharya et al., 1994b; Malod et al., 1997) appear to suggest this ridge is flanked on both sides by oceanic crust. Based on the gravity modeling and plate tectonic reconstructions, Todal and Eldholm (1998) opined that the Laxmi Ridge is a marginal high complex, comprising both continental and oceanic crust, where inner part of the ridge is underlain by faulted continental blocks. Talwani and Reif (1998) have modeled the Laxmi Ridge as a continental fragment lying between the oceanic crust of Arabian Basin in the west and the oceanic crust of the Laxmi Basin in the east. Based on the modeling of velocity structure of the Laxmi Ridge using the recently acquired seismic reflection and refraction data, Collier et al. (2004a, b) observed that the Laxmi Ridge is associated with bright reflectivity similar to the confirmed continental crust in the continental rise region off Pakistan. Using the same data, Lane et al. (2005) inferred that the Laxmi Ridge is underlain by magmatic underplating. Bansal et al. (2005), based on their admittance analysis of gravity data, interpreted the Laxmi Ridge as a fragment of continental crust.

2.2.4 Laxmi Basin

Naini and Talwani (1982) divided the deep offshore regions off western continental margin of India into two provinces viz., the Eastern Basin and the Western Basin. The dividing line between these two provinces was considered to coincide approximately along the western limit of the Laxmi Ridge and the Laccadive Plateau. Bhattacharya et al. (1994b) considered a part of this Eastern Basin as a distinct entity and named that part as the “Laxmi Basin”. The Laxmi
Basin (Fig. 2.2) was considered to be bounded in the west by the Laxmi Ridge, in the south by the northern extremity of the Laccadive Plateau, and in the east by the continental slope of India respectively. The northern limit of the basin is considered to be limited approximately along 21°N, where the bathymetric contours of the adjacent slope region abruptly change to westerly trend. It may be noted that the deep-sea basin region eastward of the Laxmi Ridge and the Laccadive Plateau was referred by Biswas and Singh (1988) as the "Laxmi-Laccadive depression." In this context the Laxmi Basin represents a part of the Laxmi-Laccadive depression. Approximately along the axial part of this basin, a NNW-SSE trending seamount chain is present (Bhattacharya et al., 1994a). Apart from these seamounts, the water depths in the Laxmi Basin area range between 3000 and 3750 m and seafloor gently dips southwestward (Bhattacharya et al., 1994a). In this region, the sediment thickness is minimum over the seamounts and the adjacent Laxmi Ridge, whereas in rest of the basin, it attains a maximum thickness of about 2.0 km (Naini, 1980).

The seismic refraction and gravity studies (Naini, 1980; Naini and Talwani, 1982) in this region suggest that; i) the crustal thickness is about 17 km, which implies that the underlying crust is thicker than normal oceanic crust and nearly half that of a standard continental crust and ii) the basin is characterized by a low amplitude (~20 mgal) short wavelength (~60 km) free-air gravity low superimposed on a long wavelength (~350 km) gravity high. Based on a study of closely spaced magnetic and gravity profiles, Bhattacharya et al. (1994b) mapped the existence of well-correlatable NNW-SSE trending linear magnetic anomalies in this basin. These linear magnetic anomalies were reported to be symmetric about a central negative magnetic anomaly and the axis of symmetry coincides with a characteristic short-wavelength free-air gravity low. The magnetic lineations are contiguous and parallel to the adjacent segment of the Laxmi Ridge in the west and the continental shelf in the east. It was inferred (Bhattacharya et al., 1994b) that the Laxmi Basin magnetic lineations (Fig.2.2) record a two-limbed seafloor spreading anomaly sequence, probably representing anomalies 33n (~79 Ma, Late Cretaceous) through 28n (~62 Ma, Late Paleocene). In the axial part of the Laxmi Basin, a well-defined basement peak was observed by Naini (1980) in one of the seismic profiles. This feature is about 30 km wide and rises
by about 3.0 sec TWT above the adjacent basement. Basement rises were also observed (Naini, 1980) in the neighbouring seismic profiles, but they are much subdued and sub-surface. Together these features were considered (Naini, 1980) to represent an axial basement high zone in the Laxmi Basin. Based on analysis of additional seismic profiles in the nearby areas, Rao et al. (1992) inferred that this basement high zone represents a 360 km long linear feature and named it as the Panikkar Ridge. This basement high zone approximately coincides with the reported axis of symmetry of two-limbed seafloor spreading magnetic anomalies, which Bhattacharya et al. (1994b) inferred to represent an extinct spreading centre.

Difference of opinion exists regarding the nature of the crust underlying the Laxmi Basin. Based on semi-continental crustal thickness and lack of identifiable seafloor spreading type magnetic anomalies, Naini and Talwani (1982) believed that the crust underlying this region is transitional in nature. Based on identification of hyperbolic reflection pattern, typical of an oceanic crust, in the multichannel seismic reflection data from deep offshore regions off Ratnagiri coast, Biswas (1989) and Biswas and Singh (1988) favoured an oceanic nature of the basement in this area. Based on observation of structural and tectonic grains parallel to the ancient Precambrian structural grain of the adjacent western part of the Indian subcontinent, Kolla and Coumes (1990) inferred that the Laxmi Basin area represents rifted transitional crust. As mentioned earlier, based on identification of seafloor spreading type magnetic anomalies, Bhattacharya et al. (1994b) opined that the Laxmi Basin is underlain by oceanic crust formed as a result of a now extinct two-limbed seafloor spreading. Based on interpretation of gravity data it was opined that this area is an underplated normal oceanic crust (Pandey et al., 1995; Singh, 1999). Studying ship-borne and satellite gravity and magnetic data, Miles et al. (1998) concluded that the crust in the Laxmi Basin is rifted and underplated continental crust. It may be mentioned here, that areas north and northwest of the Laxmi Basin, i.e. in the areas northward of Laxmi Ridge, Malod et al. (1997) inferred the existence of oceanic type of crust, which evolved during chron 29r-29n (about 66-64 Ma, early Paleocene). Mainly based on gravity and magnetic data and paleogeographic reconstructions, Todal and Eldholm (1998) opined that the crust underlying the Laxmi Basin is of continental
in nature. Talwani and Reif (1998) on the other hand favoured oceanic nature of the crust underlying the Laxmi Basin. Studies of Bernard and Munschy (2000) for understanding the structural scheme and evolution of Mascarene Basin, suggested that the Laxmi Basin need to be underlain by an oceanic crust matching in width with the missing oceanic crust in the northwest Mascarene Basin. Based on the gravity and magnetic modeling, Krishna et al. (2006) opined that the Laxmi Basin is underlain by continental crust, which subsequently got modified by extensive stretching and volcanic outpourings of the Reunion hotspot. According to them, the magnetic anomalies in the Laxmi Basin, that Bhattacharya et al. (1994b) interpreted as seafloor spreading magnetic anomalies, could best be explained as volcanic intrusives within the stretched continental crust.

2.2.5 Offshore Indus Basin

The Offshore Indus Basin (Fig. 2.2) is located in the upper Indus Fan region (Miles et al., 1998) and is bounded by the east-west trending buried Laxmi Ridge to the south, the Murray Ridge and the Owen Fracture zone in the northwest and the continental slope of India and Pakistan in the northeast. Some researchers (Malod et al., 1997; Collier et al., 2004a, b) used the nomenclature of ‘Gop Rift’ for the Offshore Indus Basin region. The southern boundary of the basin lies approximately along 3000 m isobath, where the basin merges with the northern boundary of the Laxmi Basin. The water depths in the basin range from 1400-1600 m at the foot of the continental slope to ~3400 m near the E-W trending buried segment of the Laxmi Ridge (Bhattacharya and Chaubey, 2001). The maximum sediment thickness in the basin is about 6 sec two-way travel time (TWT), of which the fan type sediments may exceed 3 sec TWT. These sediments are interpreted to have been deposited since the Cretaceous period, but the fan type sequences were probably deposited since Middle Oligocene to Early Miocene (Kolla and Coumes, 1987). The tectonic structure of the basement comprised of E-W trending horst and graben with several NE-SW basement fault. Malod et al. (1997) reported the presence of a basement high feature, called Palitana Ridge, in the axial part of the Offshore Indus Basin. The genesis of this feature has been attributed by Malod et al. (1997) to the uplifting during the
Miocene reactivation. This feature has been well demarcated in the most recent available seismic reflection section of Collier et al. (2004a, b).

As in the case of the Laxmi Basin, difference in opinion exists about the nature of the crust in the Offshore Indus Basin also. Naini and Talwani (1982) had observed some of the linear magnetic anomalies in this region; however, they were not successful to model these anomalies in terms of seafloor spreading. Malod et al. (1997) interpreted the Offshore Indus Basin to have been underlain by oceanic crust formed as a result of two-limbed seafloor spreading between the Laxmi Ridge and the adjacent continental slope of Pakistan. They identified these magnetic lineations as 29r-29n (~66-64 Ma, Early Paleocene), and made attempt to correlate this result with the inferred seafloor spreading magnetic anomalies in the Laxmi Basin. Based on the gravity and magnetic modeling exercises, Miles et al. (1998) interpreted that the Offshore Indus Basin is underlain by underplated and thinned continental crust, where the magnetic anomalies can be explained in terms of intrusives within the thinned continental crust instead of seafloor spreading type magnetic anomalies. Krishna et al. (2006) negated the possibility of an oceanic crust underlying the Laxmi Basin region; however, they supported the views that the Offshore Indus Basin region is underlain by oceanic crust.

2.2.6 Laccadive Basin

The Laccadive Basin – a narrow triangular shaped basin – is located between the Laccadive Plateau in the west and the southwestern continental slope of India in the east (Fig. 2.2). The northern boundary of the basin lies approximately near 16°N where the northern extremity of the Laccadive Plateau apparently meets the adjacent continental slope of western India. In the south, this basin opens into the Central Indian Basin. The water depth in this basin varies from ~2000 m in the north to ~2800 m in the south (Bhattacharya and Chaubey, 2001). Based on limited magnetic and seismic reflection data Rao and Bhattacharya (1975) inferred that the underlying basement in this area is block-faulted. The seismic reflection studies (Ramaswamy and Rao, 1980; Naini and Talwani, 1982; Rao and Srivastava, 1984) suggest that the sediment thickness in this basin is about 2.5 sec (TWT) in the southern part, which gradually increases to about 3.5 sec (TWT) towards the northern part of the basin. The underlying
basement widens and deepens towards south and is characterized by several basement high features, which form an approximately NNW-SSE trending lineament which was named by Naini and Talwani (1982) as the 'Prathap Ridge'.

The Laccadive Basin is reported to be associated with broad low to subdued magnetic anomaly, and generally low free-air gravity anomaly (Rao and Bhattacharya 1977; Naini and Talwani, 1982; Rao et al., 1987; Subrahmanyam et al., 1995). However, these geophysical signatures are locally modified due to the presence of Prathap Ridge, which is associated with relative free-air gravity high. Based on gravity studies, Subrahmanyam et al. (1995) opined that the Prathap Ridge is offset along pre-existing ENE-WSW trending Precambrian fault trends extending from adjacent Indian mainland and was formed during the separation of India from Madagascar. The Prathap Ridge is mostly buried below sediments and divides the basin in two parts. Based on widely spaced seismic reflection profiles, Rao et al. (1987) inferred that the ridge is depicted as a single peak basement high in the north, and multiple peaks in the south. Based on the magnetic data and seismic reflection studies Krishna et al., (1992) interpreted that, the Prathap Ridge consists of basement having variable magnetic signatures, and formed due to Reunion Hotspot activity. By identifying rotated fault blocks representing half-grabens, which are equidistant from a central basement high (Prathap Ridge) in the Laccadive Basin, Chaubey et al. (2002b) suggest that the basin is formed as a result of failed rift and volcanism of the stretched continental regime.

2.2.7 Laccadive – Chagos Ridge

The Laccadive-Chagos Ridge is one of the most prominent physiographic and aseismic features of the Indian Ocean (Fig. 1.1). This slightly arcuate major elongated tectonic feature is considered to extend for about 2500 km between 12°S and 14°N. A considerable length of the crest of this ridge is composed of shoals, banks, and coral reefs at depths less than 1500 m. This ridge can be divided into three main segments by breaches in its topographic continuity due to several relatively deep saddle like features (Bhattacharya and Chaubey, 2001). Following Bhattacharya and Chaubey (2001) these three segments of Laccadive-Chagos Ridge are being referred from north to south as the Laccadive Plateau, the Maldive Ridge and the Chagos Bank.
Gravity data in the Laccadive Plateau region indicates that the Free-air gravity anomaly, in general, is negative and subdued. However, a belt of relative positive anomalies was observed approximately over the crestal region (Talwani and Kahle, 1975; Avraham and Bunce, 1977; Naini, 1980; Naini and Talwani, 1982). The magnetic anomalies over the eastern half of the Laccadive Plateau are reported (Naini and Talwani, 1982; Rao et al., 1987) to be subdued, whereas its western half in contrast, appears to be associated with several prominent high amplitude anomalies. Seismic reflection studies suggest that the basement in general is in the form of a broad bulge, over which at places sharp peaks are present. These peaks are devoid of sediment cover and some of them reach very close to the sea surface. In the areas north of 12°N, the basement appears to consist of smaller blocks that drop in a step-like fashion to the west (Naini, 1980; Naini and Talwani, 1982; Reddy et al., 1988).

A number of workers have postulated that the Laccadive-Chagos Ridge is an inactive and subsided part of a linear volcanic feature formed during the northward motion of the Indian plate over the Reunion Hotspot. The linearity of this ridge and the north-south age progression of the volcanic rocks along the trace of the ridge are considered as strong evidences for such a hotspot model (Francis and Shor, 1966; Dietz and Holden, 1970; Whitmarsh, 1974; Morgan, 1981; Duncan, 1981; Duncan and Hargraves, 1990; Verzhbitsky, 2003). However, a number of alternate models were also forwarded to explain the origin of this ridge. Particularly, it appears that many observations and inferences do favour a non-hotspot model of origin for the Laccadive Plateau region. For example, based on seismic refraction studies Babenko et al. (1981) inferred that the Moho in this region lies at a depth of about 18-19 km, which suggests that the thickness of the crust in the Laccadive Plateau region is higher than the normal oceanic crust. Higher than normal oceanic crust thickness was also inferred by Naini and Talwani (1982) from analysis of seismic refraction data and by Chaubey et al. (2002b) from two-dimensional modeling of gravity and magnetic data. Based on identification of several rotated fault blocks from multichannel seismic reflection data, Murty et al. (1999) inferred existence of continental ribbon (or continental fragment) structure over the Laccadive Plateau region. The lack of appreciable magnetic anomalies led Rao et al. (1987) to infer continental origin
for the Laccadive Plateau region. Based on two-dimensional modeling of magnetic data, Satyanarayana et al. (1997) inferred that the basement of the Laccadive Plateau region is volcanic in nature. Narain et al. (1968), based on seismic refraction results of Francis and Shor (1966), have opined that the Laccadive Plateau region forms a transition between oceanic crust to the west and continental crust to the east. Fisher et al. (1971) have suggested that the Laccadive-Chagos Ridge was built up over an old transform fault during India’s northward movement. By considering the refraction velocities over Laccadive-Chagos Ridge (Francis and Shor, 1966) and plate reconstruction models, McKenzie and Sclater (1971) opined that the Laccadive-Chagos Ridge was formed due to volcanism since Upper Cretaceous. Avraham and Bunce (1977) suggested that the Laccadive-Chagos Ridge is composed of structural elements of multiple origin. They suggested that the ridge consists in part of several north-south fracture zones and in parts of volcanic features formed either by leaky transform faults or by the passage of Indian plate over a hotspot.

2.3 Tectonic elements of the areas conjugate to the western continental margin of India and adjoining areas

As described in the previous sections, the western continental margin of India and the adjoining deep offshore regions were formed due to the rifting and successive drifting of India, Seychelles and Madagascar. In this framework, some of the tectonic elements off west coast of India appear to be related to separation of India and Seychelles while few others appear to be related to India–Madagascar separation. Therefore, a brief account of those tectonic elements from the conjugate areas of Seychelles and Madagascar that appears genetically related to the study area has been presented in this section.

2.3.1 Features on the eastern part of Madagascar mainland

About two-third part of Madagascar is occupied by Precambrian rocks (Boast and Nairn, 1982), which are overlain and intruded by Mesozoic and Cenozoic rocks related to the separation of Madagascar from Africa and India during the break up of Gondwanaland. The Precambrian of Madagascar can be subdivided into southern and central-northern sectors, separated by Ranotsara Shear Zone (Fig. 2.3). The southern Madagascar is divided into six tectonic units
Fig. 2.3. Major structural and tectonic trends within the Mascarene Basin and adjacent Madagascar and Seychelles regions. RSZ: Ranotsara Shear Zone; AXSZ: Axial Shear Zone; SB: Seychelles Bank; SM: Saya de Malha Bank; MI: Mauritius Island; RI: Reunion Island. Compiled from Windley et al. (1994), Bernard and Munschy (2000), and Chaubey et al. (2002a). Other details are as in Fig. 2.2.
(Windley et al., 1994) and the central-northern Madagascar is divided into five tectonic units (Collins et al., 2000a, b; Collins and Windley, 2002), which are separated from each other by a regionally significant unconformity or by shear zones. In the southern Madagascar, from west to east, these tectonic units are Vohibory belt, Ampanihy belt, Bekily belt, Betroka belt, Tranomaro belt and Dauphin–Anosyan belt. The tectonic units in the north-central Madagascar, subdivided by Collins and Windley (2002) are the Antongil Block, Antananarivo Block, Itremo Sheet, Tsaratanana Sheet and the Bemarivi Belt. The Axial Shear Zone is a high-grade zone which forms a central N-S axis to the north-central region of Madagascar. It is dominated by granulite and high amphibolite facies gneisses that commonly contain graphite.

The Ranotsara Shear Zone and the Axial Shear Zone have been considered by several authors (Katz and Premoli, 1979; Windley et al., 1994; Sacks et al., 1997) to provide the qualitative models to depict the India-Madagascar juxtaposition in Gondwanaland perspective. Windley et al. (1994) considered the sinistral Ranotsara Shear Zone and the dextral Axial Shear Zone as the continuation of Achankovil Shear Zone and Palghat-Cauvery Shear Zone in India respectively. However, based on the detailed field studies in the Achankovil shear zone area, Sacks et al. (1997) reported that the Achankovil shear zone is dextral in nature and therefore, the proposition put forward by Windley et al. (1994) cannot be accepted.

The additional geological information available from the Madagascar in context of the present study is the presence of volcanics throughout the east coast of Madagascar. Storey et al. (1995) made an attempt to obtain the age for these volcanic rocks based on the $^{40}\text{Ar} / ^{39}\text{Ar}$ method, and they provided a mean age of ~88 Ma for this volcanic province. This Cretaceous volcanic province in the east coast of Madagascar is considered to have been formed by the influence of Marion hotspot.

2.3.2 Eastern continental margin of Madagascar

The continental shelf of Madagascar (Fig. 2.3) is generally narrow, averaging about 25 km in width. At some places, all along the northeast coast, no shelf is present (Pepper and Everhart, 1963). The straight east coast is bordered
by a narrow shelf whose edge is a fault scarp to extend from Fort Dauphin to the Bay of Antongil. The scarp dips steeply to 1800 m, and the continental slope to this depth lies along the fault plane. The Bay of Antongil, which is bordered on each side by a fault trending northwestward along its shore, is probably a downfaulted block trending at an acute angle to the main coastal fault (Pepper and Everhart, 1963). This straight edge has been interpreted by several authors (Barron, 1987; Lawver et al., 1999) as an evidence of transform motion between India and Madagascar that is believed to have been taken place between 160 and 105 Ma.

2.3.3 Seychelles—Mascarene Plateau

Seychelles, a Precambrian continental fragment in the Western Indian Ocean, is located approximately between 4°S and 6°S and between 54°E and 57°E (Figure 2.3). This oval shaped Seychelles Bank, of an area of about 80000 sq. km, has an almost flat top lying at an average water depth of 50 m. The edge of the bank is well defined by steep slopes, which drops to depths exceeding 3000 m in all directions, except for the southwest and the southeast. The Seychelles Bank is connected by a 2000 m deep saddle to the Amirante Arc in the southwest, whereas to the southeast, it is separated from the remainder of Mascarene Plateau by a 1500 m deep saddle (Mart, 1988).

Although situated in an oceanic environment, the continental character of the Seychelles Bank has been confirmed by seismic refraction profiling (Francis et al., 1966) and by the determination of Precambrian ages of around 700 Ma of the granitic rocks, which make up most of the Islands. The numerous islands that protrude from the Seychelles Bank are founded on igneous rocks, which outcrop prominently on the three largest islands, Mahe, Praslin and Silhouette. The Precambrian basement of the Mahe and adjacent island groups show that the granites in these islands are commonly hornblende granite, as well as porphyritic granite and aplite, with dioritic and gabbroic xenoliths in places (Baker, 1963 quoted by Mart, 1988). The granites of Mahe and its adjoining islands (except Silhouette and North Island, which are Paleocene) are Late Precambrian and are the type usually found on major continental landmasses (Khanna and Walton, 1992). The sedimentary sequence observed on the Seychelles Bank consists of Quaternary sediments. However a thick sedimentary sequence was encountered
in several exploratory bore holes, which were drilled at the western edge of the bank. The Mesozoic series drilled in these boreholes consists of about 300 m of clastic Triassic sediments overlain by about 2000 m of Jurassic rocks. The Jurassic series is covered by 400-900 m of Cretaceous rocks, the upper part of which consists of volcanic rocks. The volcanics, in turn, are overlain by a Tertiary sedimentary sequence, which consists of approximately 1500 m of carbonates, with clastics in places (Kamen-Kaye, 1985 quoted by Mart, 1988). Seismic refraction measurements show that the Seychelles Bank has a crustal thickness of more than 30 km, and the continental crust is characterized by three layers that show seismic compressional velocities of 5.7, 6.3 and 6.8 km/sec. The region southeast of Seychelles represents a normal passive continental margin (Matthews and Davies, 1966).

The Mascarene Plateau (Fig. 2.3) is a major physiographic feature located in the area between Madagascar and the Central Indian-Carlsberg Ridge segments. This feature is connected with the Seychelles continental block through a broad 1500 m deep saddle in the north, and to the Mauritius Island, through a 2500 m deep channel in the south (Mart, 1988). This arcuate aseismic ridge consists of the Saya de Malha Bank, the Nazareth Bank, the Cargados Carajos Bank and Mauritius Island. Among these, the Saya de Malha, the Nazareth and the Cargados Carajos banks are submerged at water depths of approximately 50 m. These banks are bounded by steep scarps that drop down to depths greater than 2000 m. The Saya de Malha Bank, which is the largest among the three banks, has a 350 km wide southward dipping summit surface (Kara and Sivukha, 1990 quoted by Bhattacharya and Chaubey, 2001). The radiometric dating of the basalts of Saya de Malha Bank provided an age of 45 Ma, while Nazareth bank basalts provided an age of 31 Ma. The Mauritius Island is an eroded volcanic island built perhaps by three eruptive episodes.

2.3.4 Madagascar Ridge

Madagascar Ridge (Fig. 1.3) is a major physiographic feature, which extends south of the Madagascar Island. This N-S striking elongated feature extends between latitudes between 26°S and 36°S, with a maximum width of 750 km at 32°S latitude. This aseismic ridge separates Madagascar Basin from Mozambique Basin in such a ways that it acts as the western boundary of the
Madagascar Basin and Eastern Boundary of the Mozambique Basin. To the south, this ridge abuts at the Southwest Indian Ridge.

The bathymetric map compiled by Goslin et al. (1980) provided detailed information on the topography of the Madagascar Ridge. Based on the detailed analysis of the bathymetry and seismic data, Goslin et al. (1980) suggested that the Madagascar Ridge could be separated into two distinct domains – northern domain and southern domain, which are separated by a broad saddle deeper than 2000 m. The northern area, between latitude 31°S and the continental shelf of the Madagascar, shows complex seafloor and basement topography and is associated with short wave length magnetic anomalies (Schlich, 1982). The crust underlying this northern domain is considered to be anomalous, as it has neither purely continent nor oceanic affinity (Bhattacharya and Chaubey, 2001). The southern area, between latitudes 32°S and 35°S, is generally flattish at depths of less than 1500 m. This part of the ridge corresponds to a subdued topography, where the Moho is inferred to lie about 14 km below sea level. The velocity-depth distribution of the southern domain is closely related to that of mean oceanic crust (Recq et al., 1979 quoted by Schlich, 1982; Goslin et al., 1981). Gravity studies also suggested a contrast in the isostatic compensation of the two domains. The northern domain appears to have achieved local isostatic equilibrium by crustal thickening, where as the southern domain is isostatically unbalanced with respect to the northern domain and the adjacent ocean basins (Goslin et al., 1981; Bhattacharya and Chaubey, 2001).

The western flank of the Madagascar Ridge, facing the Mozambique Basin, is steep and approximately rectilinear in its southern and central part. The eastern flank shows obvious differences in topography north and south of 31°S/32°S. To the north, the eastern flank has many small-scale topographic features on its slope. To the south, a smooth slope separates the ridge from the adjacent Madagascar Basin (Goslin et al., 1980). It is suggested (Schlich, 1982) that the southern domain of the Madagascar Ridge had probably been created simultaneously during an episode of anomalous volcanism on the flanks of the Southwest Indian Ridge approximately at the time of anomaly 31n or 29n (~68-64 Ma).
2.3.5 Mascarene Basin

The Mascarene Basin (Fig. 2.3) is bordered on the west by the steep, linear and presumably faulted eastern margin of the Madagascar Precambrian massif and to the east by the Mascarene Plateau (Schlich, 1974). This basin corresponds to the northwestern extension of the Madagascar Basin where the limit is marked by the Mauritius Fracture Zone, which offsets the magnetic pattern right laterally by about 700 km. To the south, the Mascarene Basin abuts the complex northeastern flank of the Madagascar Ridge and to the north; the basin extends towards the Farquhar Group, the Amirante arc, Seychelles Bank and the northern tip of the Madagascar Island (Schlich, 1982; Bhattacharya and Chaubey, 2001).

Deep sea drilling at site 239 in the Mascarene Basin suggested a Late Cretaceous age for this basin (Schlich et al., 1974). Based on the analysis of bathymetric, magnetic and seismic data, the structural scheme of the Mascarene Basin was proposed first by Schlich and Fondev (1974) and Schlich (1974) and later updated by Schlich (1982). These studies mapped three fracture zones between the Madagascar margin and the western scarp of the Mascarene Plateau running almost parallel to the Madagascar Basin fracture zone system, and identified two complete sequences of early Paleocene and late Cretaceous magnetic anomalies in the southern Mascarene Basin. Since the youngest magnetic anomaly identified in Mascarene Basin is the anomaly 28n, Schlich (1982) suggested that spreading in the Mascarene Basin ceased just prior to time of anomaly 27n. The extinct spreading centre corresponds to a clearly identifiable topographic high, which could be observed on several bathymetric profiles. To the south, the Late Cretaceous magnetic anomalies 32n to 34n have been easily recognized which lie at the foot of the northern Madagascar Ridge. But, these studies could not identify the magnetic lineations in the northern part of the Mascarene Basin and its age was doubtful (Masson, 1984). However, geophysical studies by Masson (1984) inferred three transform faults in the northern part of the Mascarene Basin.

Subsequent studies by Dyment (1991, 1996) and Sahabi (1993) provided revised magnetic anomaly identifications in the southern Mascarene Basin, but the kinematic evolution of the northern part of the Mascarene Basin remained
unknown. Recently, based on the bathymetric, seismic and magnetic data, Bernard and Munschy (2000) proposed a new structural scheme for the whole Mascarene Basin, including the northern part of the Mascarene Basin. They identified five new compartments and numerous fracture zones in the northern part of the Mascarene Basin (Fig. 2.3). Their interpretation suggests a southward progressive extinction of the Mascarene spreading centre.

According to the presence of anomaly 34n (83 Ma) off the Madagascar margin in the Mascarene Basin, it is believed that the seafloor spreading in the Mascarene Basin commenced during mid-Cretaceous time (Norton and Sclater, 1979; Besse and Courtillot, 1988). Since this time falls under the Cretaceous long normal superchron, the precise age of the rift-drift transition cannot be determined from the magnetic lineations.

2.3.6 Madagascar Basin

The Madagascar Basin is bound on the southeast by the rough topography of the Southwest Indian Ridge, and to the southwest by the gentle upward slope of the Madagascar Ridge. To the northwest, the Madagascar Basin is considered to have been separated from the Mascarene Basin by a clearly defined approximately NE-SW trending Mauritius Fracture Zone. To the northeast, the Madagascar Basin is limited by the rugged topography of the southern section of the Central Indian Ridge. Two major fracture zones, nearly parallel to the trend of Southwest Indian Ridge, have been traced throughout the Madagascar Basin (Schlich, 1974; Schlich, 1975 quoted by Bhattacharya and Chaubey, 2001). The basin contains the entire sequence of anomalies 20n to 30n (~43 to 66 Ma). The oldest anomaly, 30n (~66 Ma), was found in the southern Madagascar Basin close to the Madagascar Ridge, while the youngest anomaly, 20n (~43 Ma) was observed in the area south of the Mauritius Islands. The DSDP site 245 in the southern Madagascar Basin was located south of and immediately next to anomaly 29n (~64 Ma). The oldest sediments recovered from this site are of Early Paleocene (~63.6 – 66.4 Ma). The Madagascar Basin is inferred to have been created by spreading from the Central Indian Ridge and its conjugate northern basin is considered to lie just east of the Laccadive-Chagos Ridge and forms part of the Central Indian Basin.
2.4 Dated volcanism in the areas related to the study

The ages of various volcanic formations in and around Indian mainland, Seychelles Bank and Madagascar mainland have been estimated by various researchers (Fisher et al., 1968; Shipboard Scientific Party, 1972; Duncan and Hargrave, 1990; Valsangkar et al., 1981; Dickin et al., 1987; Storey et al., 1995; Rathore et al., 1997; Radhakrishna et al., 1990, 1994, 1999; Hofmann et al., 2000; Torsvik et al., 2000; Widdowson et al., 2000; Anil Kumar et al., 2001; Pande et al., 2001; Kothari et al., 2001). In Table 2.1, available estimated ages of volcanic formations in and around Indian mainland (Fig. 2.4) have been compiled and presented. This age estimates are of volcanic samples recovered from drill wells over the continental shelf area, volcanic rocks exposed in the coastal belt, and volcanic rock recovered from a drill hole over the Laccadive Plateau. The compiled data shows occurrence of volcanic rocks of broadly three age groups. The rocks of around 65 Ma age, were related to Deccan Trap basalts (Rathore et al., 1997; Hofmann et al., 2000). The rocks older than 85 Ma perhaps represent India-Madagascar separation related volcanism or even older events. The volcanic rocks younger than 65 Ma, perhaps represents those formed by passage of India over Reunion Hotspot. The evidence of two episodes of volcanism over Laccadive Plateau (Padua Bank) and the estimated ages of 102 Ma and 60 Ma for these events (Kothari et al., 2001) appears interesting. The 60 Ma age corresponds well with the ages along the predicted track of Reunion Hotspot, so it could be related to the hotspot volcanism. However, the 102 Ma age volcanism appears to be surprising due to the fact that the Laccadive Plateau is considered by many authors as a trail of Reunion Hotspot implying that all volcanism over the Laccadive Plateau should be same or younger than 65 Ma. May be the presence of 102 Ma age volcanic rocks over the Laccadive Plateau indicates that this plateau region is a continental fragment which was in existence even before the India-Madagascar separation.

The available estimated ages of volcanic formations in and around Madagascar is compiled in Table 2.2. This age estimates are of volcanic samples recovered from the Cretaceous volcanic and dyke rocks from Madagascar mainland (Fig. 2.5). From this compilation, it is observed that the Cretaceous volcanic and intrusive rocks of Madagascar crop out semi-continuously along the
Table 2.1 Estimated ages of volcanic rocks from selected locations over the western part of the Indian mainland and offshore region as compiled from various publications. Where the position locations are not given in the publication, they were digitized from respective figures.

<table>
<thead>
<tr>
<th>ID</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Rock type and sample locale</th>
<th>Age (Ma)</th>
<th>Dating Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>19° 00'</td>
<td>71° 33'</td>
<td>Basalts, continental shelf off Bombay</td>
<td>65</td>
<td>K-Ar</td>
<td>Rathore et al. (1997)</td>
</tr>
<tr>
<td>B</td>
<td>17° 55'</td>
<td>73° 38'</td>
<td>Deccan Trap basalts, Mahabaleswar</td>
<td>65</td>
<td>40Ar/39Ar</td>
<td>Hofmann et al. (2000)</td>
</tr>
<tr>
<td>C</td>
<td>15° 30'</td>
<td>73° 44'</td>
<td>Basaltic dykes, Goa coast</td>
<td>63</td>
<td>40Ar/39Ar</td>
<td>Widdowson et al. (2000)</td>
</tr>
<tr>
<td>D</td>
<td>13° 27'</td>
<td>72° 32'</td>
<td>Basalts, Padua Bank, Laccadive Plateau</td>
<td>102</td>
<td>K-Ar</td>
<td>Kothari et al. (2001)</td>
</tr>
<tr>
<td>E</td>
<td>13° 21'</td>
<td>74° 39'</td>
<td>Felsic volcanic rocks, St. Mary Islands, northwest off Mangalore</td>
<td>93</td>
<td>K-Ar</td>
<td>Valsangkar et al. (1981)</td>
</tr>
<tr>
<td>F</td>
<td>12° 48'</td>
<td>77° 10'</td>
<td>Mafic dykes, Hulyardurga, Karnataka</td>
<td>89</td>
<td>40Ar/39Ar</td>
<td>Anil Kumar et al. (1988, 2001)</td>
</tr>
<tr>
<td>G</td>
<td>12° 22'</td>
<td>75° 16'</td>
<td>Dolerite dykes, North Kerala</td>
<td>129</td>
<td>K-Ar</td>
<td>Radhakrishna et al. (1999)</td>
</tr>
<tr>
<td>H</td>
<td>11° 56'</td>
<td>75° 28'</td>
<td>Dolerite dykes, North Kerala</td>
<td>99</td>
<td>K-Ar</td>
<td>Radhakrishna et al. (1999)</td>
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<tr>
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<td>75° 30'</td>
<td>Dolerite dykes, North Kerala</td>
<td>54</td>
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<td>J</td>
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<td>75° 45'</td>
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<td>112</td>
<td>K-Ar</td>
<td>Radhakrishna et al. (1999)</td>
</tr>
<tr>
<td>K</td>
<td>11° 27'</td>
<td>75° 47'</td>
<td>Dolerite dykes, North Kerala</td>
<td>61</td>
<td>40Ar/39Ar</td>
<td>Radhakrishna et al. (1999)</td>
</tr>
<tr>
<td>L</td>
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<td>75° 53'</td>
<td>Felsic dykes, North Kerala</td>
<td>82</td>
<td>K-Ar</td>
<td>Radhakrishna et al. (1999)</td>
</tr>
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<td>Dolerite Dykes, Agali-Anaikatti, Karnataka</td>
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<td>K-Ar</td>
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<tr>
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<td>76° 40'</td>
<td>Leucogabbro dykes, Central Kerala</td>
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<td>40Ar/39Ar</td>
<td>Radhakrishna et al. (1994)</td>
</tr>
<tr>
<td>O</td>
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<td>Mafic dykes, South Kerala</td>
<td>144</td>
<td>K-Ar</td>
<td>Radhakrishna et al. (1990)</td>
</tr>
</tbody>
</table>

K-Ar : Potassium – Argon
40Ar/39Ar : 40 Argon – 39 Argon
ID : Identification of the locations as given in Fig. 2.4.
+ : In this area, age of basalts in different flows range between 60 and 102 Ma
##: In this area Torsvik et al. (2000) dated some rocks at ~91 Ma, using U-Pb method and Pande et al. (2001) at ~85 Ma using 40Ar/39Ar Method.
Fig. 2.4. Locations of selected volcanic formations in the west coast of India and the adjacent offshore regions with their estimated ages of emplacement. Red solid triangles are locations where volcanic formations were dated and ages (in Ma) of formations at locations are given within parenthesis. Details of age compilation are presented in Table 2.1. Other details are as in Fig. 2.2.
Table 2.2 Estimated ages of volcanic rocks from selected locations over the eastern part of the Madagascar mainland, Seychelles and the adjoining areas as compiled from various publications. Where the position locations are not given in the publication, they were digitized from respective figures.

<table>
<thead>
<tr>
<th>ID</th>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>Rock type and sample locale</th>
<th>Age (Ma)</th>
<th>Dating Method</th>
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<td>24° 34'</td>
<td>46° 11'</td>
<td>Rhyolite, Volcan de L'Androy</td>
<td>87</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>B</td>
<td>24° 24'</td>
<td>46° 05'</td>
<td>Rhyolite, Volcan de L'Androy</td>
<td>85</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>C</td>
<td>23° 09'</td>
<td>47° 35'</td>
<td>Basalt, Eastern Madagascar</td>
<td>89</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>D</td>
<td>22° 42'</td>
<td>47° 40'</td>
<td>Basalt, Eastern Madagascar</td>
<td>89</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>F</td>
<td>21° 17'</td>
<td>48° 05'</td>
<td>Basalt, Manajary, Eastern Madagascar</td>
<td>87</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>G</td>
<td>20° 55'</td>
<td>48° 14'</td>
<td>Basalt, Manajary, Eastern Madagascar</td>
<td>85</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>H</td>
<td>20° 42'</td>
<td>48° 17'</td>
<td>Basalt, Manajary, Eastern Madagascar</td>
<td>84</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>I</td>
<td>19° 24'</td>
<td>48° 38'</td>
<td>Rhyolite, between Manajary and Tamatave, Eastern Madagascar</td>
<td>88</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
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<td>J</td>
<td>17° 14'</td>
<td>49° 16'</td>
<td>Basalt, north of Tamatave, Eastern Madagascar</td>
<td>86</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>K</td>
<td>14° 50'</td>
<td>50° 14'</td>
<td>Basalt, south of Sambava, Eastern Madagascar</td>
<td>91</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>L</td>
<td>13° 29'</td>
<td>50° 02'</td>
<td>Basalt, south of Vohemar, Eastern Madagascar</td>
<td>90</td>
<td>40Ar/39Ar</td>
<td>Storey et al. (1995)</td>
</tr>
<tr>
<td>M</td>
<td>13° 16'</td>
<td>50° 01'</td>
<td>Basalt, Analalava Pluton, northeastern Madagascar</td>
<td>92</td>
<td>U-Pb</td>
<td>Torsvik et al. (2000)</td>
</tr>
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<td>6° 40'</td>
<td>52° 35'</td>
<td>Tholeiitic basalts, Western flank of Amirante arc</td>
<td>82</td>
<td>K-Ar</td>
<td>Fisher et al. (1968)</td>
</tr>
<tr>
<td>O</td>
<td>4° 19'</td>
<td>55° 43'</td>
<td>Dolerite dykes, Praslin Island</td>
<td>73</td>
<td>K-Ar</td>
<td>Dickin et al. (1987)</td>
</tr>
<tr>
<td>P</td>
<td>4° 28'</td>
<td>55° 13'</td>
<td>Syenite, Silhouette Island</td>
<td>66</td>
<td>K-Ar</td>
<td>Dickin et al. (1987)</td>
</tr>
</tbody>
</table>

40Ar/39Ar : 40Argon – 39Argon
ID : Identification of the locations as given in Fig. 2.5.
Fig. 2.5. Locations of selected volcanic formations in the east coast of Madagascar, Seychelles and the adjacent regions with their estimated ages of emplacement. Red solid triangles are locations where volcanic formations were dated and ages (in Ma) of formations at locations are given within parenthesis. Details of age compilation are presented in Table 2.2. Other details are as in Fig. 2.3.
1500 km length of the east coast of Madagascar, which marks the rifted margin. The rocks include basalt flows, dykes and some rhyolitic flows. Along the rifted margin, the flows lie mainly on the Precambrian basement, whereas most of the dykes are parallel to the coast. The Volcan de L’Androy complex in southern Madagascar contains the thickest sequence of Cretaceous volcanic rocks exposed on the Island. The compiled age information in and around the Madagascar mainland and its offshore region indicate that these widespread Cretaceous flood basalts of Madagascar can be related to the track of the Marion hotspot.

The geochronological studies on igneous rocks from Seychelles indicated that the region has been subjected to, or influenced by, several phases of igneous activity. These major igneous rocks are initial granite emplacement (~650 Ma), Karoo Dolerite equivalents (~180 Ma), Lebombo-Movene volcanic equivalents (~140 Ma), Amirante volcanism (~75 Ma) and Deccan Trap equivalents (~63 Ma). However, the volcanics are yet to be recognized in Seychelles related to the rift between Seychelles and Madagascar, which is predicted to have occur at around 96-84 Ma (Plummer, 1995), before the initiation of the seafloor spreading in the Mascarene Basin at around 83 Ma. In Table 2.2 available estimated ages of volcanic formations in and around Seychelles is compiled. These age estimates are of volcanic samples recovered from Amirante Arc, Praslin Island, Silhouette Island and the basement sampled at ODP Site 707 in the saddle between Seychelles and Saya de Malha (Fig. 2.5).

2.5 Passive continental margins – types, formation and evolution

Geologically, continental margins represent a transition zone, where thick granitic continental crust changes to thin basaltic oceanic crust (Davis and Fitzgerald, 2004). Depending on their characteristics, the continental margins have been broadly categorized into two basic types; 1) Passive (or Atlantic type) continental margins and, 2) Active (or Pacific type) continental margins. The passive continental margins are usually characterized by a relatively wide continental shelf, an extensive continental rise and are devoid of significant seismic activity. The passive margins develop during the processes of continental break-up and subsequent ocean basin formation. They are located within the plate interior where the continent and adjacent ocean floor are part of the same
plate. On the other hand the active continental margins are seismically very active and are, instead of a continental rise, characterized by a trench at the foot of the continental slope. The active continental margins develop where an oceanic plate is consumed beneath a continental plate at a subduction zone. Therefore in case of active margins, the continent and adjacent ocean floor belongs to different plates (Seibold and Berger, 1993; Open University Course Team, 1995).

The western continental margin of India was considered as a passive continental margin (Biswas, 1982, 1987) and later a major portion of this margin was further categorized as volcanic passive (rifted) margin (Todal and Eldholm, 1998). Therefore, to appreciate the western continental margin of India in terms of a volcanic passive continental margin framework, the generalized concepts of formation of various types of passive continental margins in general and volcanic passive (rifted) margins in particular have been briefly described in the following paragraphs.

(a) Types of passive margins

Passive margins are broadly divisible into rifted and sheared (or transform) types. Rifted continental margins form where the initial plate separation is approximately perpendicular to the rupture. They show a gradational transition, and their morphology can be subdivided into continental shelf, continental slope and continental rise. Rifted margins comprise a group with two end members: non-volcanic rifted margins - that have evolved without extensive igneous activity, and volcanic rifted margins - those in which igneous processes played a major role in their formation. Since the lithosphere has wide range of responses to rifting at different mantle temperatures, so there is probably a continuum of margin structures between non-volcanic and volcanic (Jones, 1999).

Sheared (or transform) continental margins form when the initial split is along a transform fault and differ from rifted continental margins both in their structure and development. They are usually characterized by steep continental slopes that connect the ridge-trough complexes of the oceanic transform and the transition between the thick continental crust and thin oceanic crust is sharp, usually occurring over distances of less than ~30 km. The outer parts of a
transform margin usually are underlain by a basement ridge, which bounds on its landward side a deep sedimentary basin (Jones, 1999).

(b) Formation and evolution of passive margins

Generally agreed sequence of phases for the evolution (Fig. 2.6) of passive margins is:

i) rifting;

ii) onset of drifting, i.e., separation of continental crust as oceanic crust accretes in the gap between continental blocks; and

iii) post-rift evolution, dominated by massive subsidence of the rifted margins and shaping of those margins by sedimentary and secondary (mostly gravity) tectonics (National Research Council, 1979).

Continental rifting, a thermo-mechanical process, is the first stage of the process by which continents break-up. Crustal thinning, thermal anomalies (high heat flow) and uplift are clearly associated with rifting but regarding the mechanism of initiation of rifting, there exist two models, active rifting and passive rifting. In the passive model, lithospheric tension causes failure of the continental lithosphere and results extension. The two types of the tectonic extensional stresses, which have been suggested for the lithospheric tension, are convective drag on the base of the lithosphere caused by major cellular convection currents, and plate interior stress caused by plate boundary forces. In the active model, the anomalous upper mantle develops first by some form of convective upwelling and doming and consequently, volcanism and extension. The uplifted dome and its deep isostatic compensation give rise to local tension in the lithosphere. As the lithosphere thins by heating up, the stress becomes concentrated into the relatively thin strong layer near the surface, with consequent stretching and faulting. The passive model with its modifications explains the formation of passive margins segments away from hot spot regions. Whereas the active model explains volcanism, plateau uplift and rifting primarily as a consequence of the formation of anomalously hot and low density region in the underlying mantle – a hot spot (Bott, 1995). The end products of active and passive rifting are likely to be very similar and the difference between the two are not easy to recognize in the geological record (Fitton, 1983; Golombek et al., 1996). Geologic record
suggests that rifting process often localize along ancient orogenic belts and suture zones. By weakening the crust, these pre-existing crustal discontinuities play a major role in the localization and distribution of crustal strain (Corti et al., 2003).

(c) The sediment cover and deep crustal structure at rifted continental margins

(i) Sediment cover

Normally passive margins are characterized by thick piles of sediments, however, there are a few starved margins where sediments are thin or absent due to lack of supply of sediments during their development. In general, the sedimentary rocks found at passive margins had been formed during three successive stages. First, the continental basement may be overlain by the pre-rift sediments, i.e. the sediments deposited prior to rifting. Second, syn-rift sediments of the rifting stage of margin development may be deposited at the time of initial rifting and continental stretching. Third, post-rift sediments of the drifting stage are deposited subsequent to continental break-up and onset of seafloor spreading. They are separated from the underlain syn-rift sediments by the post-rift or break-up unconformity. The post-rift sediments are generally un-faulted (except for growth faults at the slope) indicating a quiet tectonic environment apart from slow but persistent subsidence at an exponentially decreasing rate. Syn-rift sediments are typically deposited in half-grabens formed by intense stretching and thinning of the continental crust and lithosphere (Bott, 1995).

(ii) Deep crustal structure

The crustal thinning and stretching eventually lead to the break-up of continents forming a new ocean basin with a pair of passive continental margins on both the sides (Fig. 2.6). The detailed information on the structure of passive continental margins has come from seismic experiments, which suggest that the continental basement thins towards the ocean and that the transition from thick continental to thin oceanic crust occurs over distances of a few tens of kilometers to over a hundred kilometers. Various criteria have been considered to define the position of the ocean-continent boundary. These criteria have been based on crustal thickness, vertical and lateral variations in seismic velocity, seismic mode
Fig. 2.6. Schematic representation of the concept of continental break-up and formation of a pair of passive margins at successive stages. Stage a: initiation of breaking of a continental mass along a weak zone; Stage b: crustal thinning, rifting and initiation of seafloor spreading; Stage c: continents drift away as seafloor spreading continues and Stage d: subsiding continental land masses as they drift away under the load of sedimentation giving rise to shelf, slope and rise configuration. Modified after Trujillo and Turman (2005).
conversion, steep isostatic gravity gradients, the terminations of linear magnetic anomalies and the basement topography and composition. However, none of these criteria has been generally applicable because of the difficulty of distinguishing oceanic crust from continental crust modified by deformation and intrusion of basaltic material during the process of crustal attenuation (Jones, 1999).

The 'seaward-dipping-reflectors (SDRs)' first discovered by K. Hinz (1981) quoted by Talwani and Abreu (2000) and later studied extensively by many (eg. Mutter et al. 1982; White and McKenzie, 1989; Eldholm et al. 1995; Talwani and Abreu, 2000) appear to be a characteristic feature in most of the volcanic passive margins. These reflectors generally exhibit convex upward curvature with dips that increase in a seaward direction and a distinct geometry quite unlike the internal stratification of sedimentary accumulations. The SDR sequences lie adjacent to the oldest magnetic lineations in deeper water, indicating a close association with continental rifting and the onset of seafloor spreading (Jones, 1999). These SDRs are considered to represent enormous volumes of basaltic lavas, and this has been confirmed by deep drilling at places (Eldholm et al. 1995). The mechanism for producing such large amount of lava and the nature of the underlying crust is still a subject of much discussion, but it is generally agreed that eruptions took place in a sub-areal or shallow water setting from a source seaward of the present location of the reflectors and the dips arise from loading and subsidence (Jones, 1999). This sub-areal source is considered to be a sub-aerial spreading ridge along which the early opening of the ocean takes place. It is considered that during early stages of seafloor spreading, melt production rates are unusually high resulting in the development of a sub-areal spreading ridge (Jones, 1999). Some (Smythe, 1983 and Skogseid and Eldholm, 1987 quoted by Jones, 1999; White and McKenzie, 1989) believe that this SDR's were emplaced over an extended continental crust. Others (Talwani et al., 1995; Talwani and Abreu, 2000) consider that the SDRs provinces represent the initial stage of formation of oceanic crust when the spreading rate was high and called it the "Initial Oceanic Crust". They further inferred that the SDRs were emplaced symmetrically at both conjugate margins by a sub-areal spreading ridge and the SDRs constitute the extrusive part of the initial oceanic crust. This sub-areal
spreading produced lavas could flow large distances and gave rise to smooth oceanic basement. One of the view (Hinz, 1981 quoted by Talwani and Abreu, 2000) is that the subaerial extrusion of volcanics are onto a continental crust, while the other view (Mutter et al., 1982; Talwani and Abreu, 2000) is that the entire crustal material lying above Moho and underlying the SDRs wedge was emplaced after separation and the continental crust is present landward of this province. Formation of SDRs province stops when at later stage the spreading rate slows down, the ridge axis subsides below sea level and the lengths of lava flows became reduced due to rapid quenching giving rise to typical magnetic lineations and the rough basement surface.

(d) Ocean basins adjacent to passive margins

When the continental blocks move apart, the sudden upwelling of magma takes place, which solidifies, cools and get magnetized through Curie temperature in the direction of the Earth's magnetic field at that time. From the paleomagnetic results, it has been found that the Earth's dipole moment alternates between two anti-parallel polarity states, a normal state, in which the field at the Earth's surface is directed northward, and a reversed stage in which the field has the opposite direction. Therefore, the new crust formed as a result of seafloor spreading becomes magnetized in the prevalent direction of the magnetic field. When the field reverses, the oceanic crust generated at that time also becomes reversely magnetized. These two states result in formation of symmetrical stripes of normal and reverse magnetized blocks on either side of the spreading centre (Fig. 2.7). Hence, magnetic profiles across these magnetized blocks provide alternate positive and negative magnetic lineations of high amplitude with wavelengths of a few kilometers to several tens of kilometers. These oceanic magnetic lineations can be correlated with the geomagnetic polarity reversal timescales and thus the age of the ocean floor can be inferred. Usually the oldest oceanic magnetic lineations landward can constrain the time at which drifting (by seafloor spreading) began, but some continental slopes are bordered by a magnetically quiet zone, leaving the early history of seafloor spreading unclear (Jones, 1999).
A. Period of normal magnetism

B. Period of reverse magnetism

C. Period of normal magnetism

positive magnetic anomaly

negative magnetic anomaly

Earth's magnetic field

crest of midocean ridge

Normal Reverse

Magnetic stripes

Fig. 2.7 Schematic representation of alternate normally and reversely magnetized oceanic crust and associated magnetic anomaly pattern. (a) Generation of alternate bands of magnetized crust by seafloor spreading; (b) The positive and negative magnetic anomalies along a profile across the normally and reversely magnetized blocks of the oceanic crust.
(e) Continental fragments off passive margins

Slivers of continents or continental fragments have been found to exist in the deep-sea areas adjacent to many of the continental margins. These continental fragments pose problems in fitting conjugate margins. Issues such as, why do continental fragments occur in some places, what is the nature of the crust between the fragments and the continent proper, are not clear in all cases. Some of these fragments seem to be located near areas where seafloor spreading changed direction and where initial rifting was propagating with time along the line between continental masses (National Research Council, 1979). In some cases an existing spreading ridge propagated or jumped onto a continental margin, severing a small segment of stretched crust from a large continent, while the existing mid-ocean ridge became extinct (Müller et al. 2001).