CHAPTER VII

DISCUSSION
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The results obtained from the present investigations are discussed in the context of the evolution of the Indian ocean in general and the evolution of the Central Indian Basin in particular. Before going into the detailed discussions pertaining to the evolution of the Central Indian Basin, a brief overview of evolution of the Indian ocean from the breakup of Gondwanaland to Present is outlined.

7.1 EVOLUTION OF THE INDIAN OCEAN

The evolution of the Indian ocean according to the observed magnetic lineation patterns can be summarized as a consequence of the processes involved in three main phases. These phases include the breakup of Gondwanaland (Late Jurassic to Late Cretaceous), Late Cretaceous to Middle Eocene period and the Middle Eocene to the Present.

7.1.1 BREAKUP OF GONDWANALAND (LATE JURASSIC TO LATE CRETACEOUS)

The initial breakup of Gondwanaland during the Late Jurassic period at about 152 Ma (anomaly M22), resulted in the separation of Africa from
Madagascar and Antarctica. During this process, the western Somali Basin 
(Segoufin and Patriat, 1980; Rabinowitz et al., 1983; Coffin and Rabinowitz, 
1987; Cochran, 1988), the symmetric Mozambique Basin and the basin off 
Droning Maud land, Antarctica (Segoufin, 1978; Simpson et al., 1979; Bergh, 
1977, 1987) have formed.

Subsequent to the initial break up, during the Early Cretaceous period 
at about 127 Ma (anomaly M11), the South Atlantic started to form (Larson 
and Ladd, 1973; Rabinowitz, 1976) and probably during the same period 
India and Madagascar separated from Antarctica creating Mesozoic basins 
along the western margin of Australia (Markl, 1974, 1978; Larson et al., 
1979; Veevers et al., 1985). The movement between Africa and Madagascar 
Figures 7.1a and 7.1b illustrate the initial position of the southern continents 
before breakup and the reconstruction during the Early Cretaceous period 
respectively.

7.1.2 LATE CRETAEOUS TO MIDDLE EOCENE

This period recorded the rapid northward drift of Indian plate towards 
Asia. As a consequence of this process major portion of the Indian ocean 
has evolved and resulted in the formation of ocean basins. India left 
Madagascar in the Late Cretaceous period (80-90 Ma). This event was 
recorded in a suite of lavas along the east coast of Madagascar (Vallier,
Figure 7.1 (a) Initial position of the southern continents before breakup (b) Reconstructed positions of the continents during Early Cretaceous (after Norton and Sclater, 1979).
1974). Norton and Sclater (1979), opined that similar volcanism, if existed on the west coast of India, probably got masked by the later Deccan trap events. The seafloor spreading record of the first few reversals after the Cretaceous normal suggested very high spreading rates (up to 15 cm/yr, half rate) in the Latest Cretaceous (Norton and Sclater, 1979). Larson et al., (1978) suggested a major reorganisation of spreading direction between India and Antarctica during the Late Cretaceous following the separation of India from Madagascar (Figure 7.2).

In the Early Paleocene, the ridge axis between Madagascar and India jumped towards India and separated Seychelles platform from India and formed the Chagos-Laccadive transform (Norton and Sclater, 1979; Patriat and Segoufin, 1988). The Early Paleocene reconstruction indicating the separation of Seychelles from India is shown in Figure 7.3a.

In summary, during this period (Late Cretaceous to Middle Eocene) the basins between Africa-Madagascar and Antarctica (Bergh and Norton, 1976; Patriat, 1979; LaBrecque and Hayes, 1979; Sclater et al., 1981; Fisher and Sclater, 1983); the Madagascar and Mascarene Basins, the eastern Somali Basin and the Arabian sea between Africa-Madagascar and India have evolved (McKenzie and Sclater, 1971; Whitmarsh, 1974; Schlich, 1975, 1982). The mirrored Central Indian Basin and Crozet Basins between India and Antarctica (McKenzie and Sclater, 1971; Sclater and Fisher, 1974; Schlich, 1975, 1982) and Wharton Basin between India and Australia
Figure 7.2 Reconstruction during Late Cretaceous (80 Ma) (after Norton and Sclater, 1979).
(Sclater and Fisher, 1974; Liu et al., 1983) were also formed during this period.

In addition to the formation of these basins, Australia and Antarctica commenced to separate during the Eocene (Weissel and Hayes, 1972; Cande and Mutter, 1982).

7.1.3 MIDDLE EOCENE

The latest period in the evolution of the Indian ocean started with a major reorganization of the spreading centers consequent to the collision of India with Asia in the Middle Eocene.

Indian plate collided with Asia in the Middle Eocene about 43 Ma (anomalies A19, A20) (Patriat and Achache, 1984). The collision initiated a major plate reorganisation and the spreading rates decreased dramatically along the Central Indian Ridge and the Southeast Indian Ridge. The spreading direction on the Southeast Indian Ridge changed from near north-south to northeast direction.

During this phase the seafloor spreading ceased in Wharton Basin. Major changes in the direction and rate of spreading occurred in the Central Indian Basin and Crozet Basin, the Madagascar Basin, the eastern Somali
Figure 7.3 (a) Early Paleocene (65 Ma) reconstruction indicating the separation of Seychelles from India. (b) Reconstructed positions during Early Eocene (53 Ma) (after Norton and Sclater, 1979).
Basin and the Arabian Sea (Patriat and Achache, 1984; Patriat and Segoufin, 1989).

The Australian-Antarctic Basin, the southern Central Indian Basin and the northern Crozet Basins were created during this period (Weissel and Hayes, 1972; Sclater et al., 1976a; Schlich, 1975).

The Mascarene plateau and the Chagos-Laccadive Ridge were separated (Fisher et al., 1971; McKenzie and Sclater, 1971) and the Gulf of Aden opened (Laughton et al., 1970). Because of the differential motion along the Central Indian Ridge and the Southeast Indian Ridge, the South West Indian Ridge propagated rapidly towards east (Tapscott et al., 1980; Sclater et al., 1981; Patriat, 1987a). Figure 7.3b illustrates the reconstructions during the Early Eocene.

7.1.4 MIDDLE EOCENE TO PRESENT

During this tectonic epoch, the plate boundaries in the Indian ocean assumed the present status. The strike slip motion along the Ninetyeast Ridge has ceased, Australia drifted rapidly away from the Antarctica.

In this period the Reunion hotspot crossed beneath the Central Indian Ridge and transferred from Indian plate to the African plate (Patriat and

Figures 7.4a and 7.4b illustrate the reconstructions during Early Oligocene (anomaly 16, 38 Ma) and the Present positions of the continents, respectively.

7.2 EVOLUTION OF THE CENTRAL INDIAN BASIN

7.2.1 TECTONIC IMPLICATIONS OF MAGNETIC LINEATIONS, SPREADING RATES AND FRACTURE ZONES

The identified magnetic anomalies A21 to A26 in the study area define near E-W magnetic lineations (Figure 4.22). Fracture zones at 73°E, 76°30'E and 79°E are delineated based on the offsets of the magnetic lineations (see section 4.4).

The near E-W trend of the magnetic lineations indicates that these magnetic lineations originated from an E-W trending ridge which was spreading in N-S direction. McKenzie and Sclater (1971) suggested that the Central Indian Basin was formed further south and since has drifted to the present position. The paleo-position of the Southeast Indian Ridge as proposed by the reconstructions of Fisher et al. (1971) indicated that the ridge existed further south and was trending in E-W direction during the
Figure 7.4 Reconstructed positions during (a) Early Oligocene (38 Ma) and (b) Present day positions (after Norton and Sclater, 1979).
geological past. Figure 7.5 illustrates the past position and orientation of the Southeast Indian Ridge at 51 Ma, 35 Ma and at Present. The near N-S orientation of the identified fracture zones and the E-W magnetic lineations observed in the Central Indian Basin suggest the dominant N-S spreading during the period A21 to A26 from the E-W trending Southeast Indian Ridge. The model studies have indicated a spreading rate of 8.0 cm/yr (half spreading rate) between anomalies A26 to A23 and 3.6 cm/yr between anomalies A23 to A21.

The Cenozoic rapid northward drift of the Indian plate followed three successive phases (Figure 7.6). In the first phase, down to anomaly A23 Indian plate drifted with a mean velocity of 15 to 20 cm/yr (full spreading rate). Then between anomalies A23 to A13 (36 Ma) the motion of the Indian plate was rather erratic, showing several changes in the spreading direction and the velocity decreased to <10 cm/yr. In the third and the latest phase, from anomaly A13 to present, the Indian plate motion resumed to a stable northward direction with respect to the Eurasian plate, with full spreading rate less than 5 cm/yr (Patriat and Achache, 1984). The spreading rates inferred from the model studies undertaken in the present investigations have documented the Indian plate movement during a part of the first and second phases. The inferred spreading rate of 8.0 cm/yr (half spreading rate) between anomalies A26 to A23, fits in to the first phase. The dramatic decrease in the spreading rate to 3.6 cm/yr half rate (ie., 7.2 cm/yr full rate),
Figure 7.5 Paleoposition of the Southeast Indian Ridge during a) Eocene b) Oligocene and c) Present (Fisher et al., 1971).
Figure 7.6 The Indian plate movement during the geological past. The magnetic anomalies corresponding to various stages are marked (Patriat and Achache, 1984).
corresponds to the initial stages of the second phase of the Indian plate movement.

The initial decrease in the spreading rates at anomaly A23 (53 Ma) may indicate a 'soft' collision that has occurred between the Indian plate and the Eurasian plate which was followed by the 'hard' continent-continent collision in the later period. The initial result of the ODP Leg 116 suggested this process (Shipboard scientific party, Leg 116, 1987). Based on the detailed studies of the ODP Leg 116 results, Cochran (1990) proposed that the Tethyan oceanic crust at the northern portion of the Indian plate was consumed in a subduction zone along the southern margin of Eurasia. He further suggested that the northern continental margin of India encountered the subduction zone in the Eocene. The present location of the volcanic arc associated with the subduction zone is marked by the Trans-Himalayan granitic batholith, which extends over 2500 km to the north of Indus-Tsangpo suture zone (Tapponnier et al., 1981; Brookfield and Raynolds, 1981; Honegger et al., 1982).

The studies carried out on the Southeast Indian Ridge by Royer and Schlich (1988), between the Amsterdam and St.Paul islands indicated several fracture zones on this segment. These detailed kinematic studies have shown that the fracture zones tracked in the Crozet Basin can be traced in the north in the Central Indian Basin. The inferred fracture zones from the
studies of magnetics and bathymetry in the study region are suggested to be the traces of the transform faults on the Southeast Indian Ridge.

The fracture zone at 76°30'E is reported for the first time in the Central Indian Basin based on the offsets observed in the magnetic lineations. This fracture zone was earlier predicted from the studies of reconstruction models by Patriat and Segoufin (1988). These predictions were based on the observed offsets in the magnetic anomalies A20-A22 in the Madagascar Basin and the reconstruction of the same isochrons on the Indian plate (Patriat and Segoufin, 1988). However, earlier predictions indicated that the fracture zone at 76°30'E could have originated either from Southeast Indian Ridge or from Central Indian Ridge. In the present studies the observed anomalies in the region are modeled assuming that these anomalies have originated from the Southeast Indian Ridge and the fit is reasonable. Thus the model studies and the trends of the 79°E and 83°E fracture zones suggest that the 76°30'E fracture zone also originated from the Southeast Indian Ridge.

Studies by Sclater et al., (1976), have shown that the trend of the fracture zone at 86°E changes from N12°E to N45°E during the anomaly times A18 to A20, as a consequence of the reorientation of the Southeast Indian Ridge from east-west to northwest-northeast during Eocene. The trends of the fracture zones at 73°E, 76°30'E and 79°E which are identified in the present study might have undergone similar changes.
Several changes in the spreading rate and direction are observed in the vicinity of the Southeast Indian Ridge (Schlich and Patriat, 1971; Schlich, 1974). The changes in the spreading rate and direction are noticed over the crust generated by all the three ridges (Figure 7.7) (Patriat and Segoufin, 1988). One common feature of these observations is that the spreading rate and direction changes are more prominent on the Central Indian Ridge and the Southeast Indian Ridge than on the Southwest Indian Ridge. The Southwest Indian Ridge along which very slow spreading rates (< 1 cm/yr) were observed was believed to have evolved as a consequence of the eastward migration of the Indian Ocean Triple Junction (Tapscott et al., 1980; Sclater et al., 1981; Fisher and Sclater, 1983). A detailed discussion of the influence of the triple junction on the evolution of the Central Indian Basin, is given in the following section. One more important observation from the changes of spreading rate and direction (Figure 7.7) is that these changes have occurred between A18 to A20 during the Eocene period. These major changes have been attributed to the global plate reorganisation that took place in the Indian Ocean as a consequence of the collision of the Indian plate with the Eurasian plate (Schlich et al., 1976; Patriat and Achache, 1984; Royer and Schlich, 1988). The plate reorganisation is believed to have taken place immediately before anomaly A20 near 45 Ma and was completed before anomaly A19 near 44 Ma (Patriat and Achache, 1984). Consequent to the major plate reorganisation, the spreading rates have dramatically
Figure 7.7 The computed spreading direction and spreading rates with respect to Central Indian, Southeast Indian and Southwest Indian Ridges (Patriat and Segoufin, 1988). The inferred spreading rates in the present study are shown with small dashed line. Note the sudden decrease in spreading rate around anomaly 23.
decreased. The model studies in the present investigations have indicated these variable spreading rates in the Central Indian Basin.

The above discussion highlights the erratic spreading direction changes of the Indian plate movement subsequent to anomaly A20, the suggestion of a 'soft' collision as indicated by the ODP Leg 116 results and the sudden decrease in the spreading rate as observed in the present study, at anomaly A23. These observations suggest that even though the major global plate reorganization took place during the A20 and A19 anomaly times (Middle and Late Eocene period), the effects of 'soft' collision were felt during anomaly time A23 (Early Eocene period) as documented by the decrease in spreading rate.

7.2.2 INFLUENCE OF INDIAN OCEAN TRIPLE JUNCTION

Contemporaneous to the evolution of the oceanic crust in the study area, another major tectonic process that has been active in the region is the evolution of the Indian ocean triple junction. Shortly before 80 Ma, the ridge axis that had existed north of Madagascar jumped south (Figure 7.2), changed direction and commenced to separate Madagascar from India (Tapscott et al., 1980; Fisher and Sclater, 1983). An RRR (Ridge-Ridge-Ridge) triple junction was created at the southern extremity of the Madagascar Ridge, joining the newly formed Central Indian Ridge spreading center to the already active Southwest and Southeast Indian Ridges (Figure
7.2) that had been separating Africa and Antarctica and India and Antarctica respectively (Fisher and Sclater, 1983). At this stage the Southwest Indian Ridge which was spreading at a very slow rate (<1.0 cm/yr, Bergh, 1971), was coupled with faster spreading Central Indian and Southeast Indian Ridges. The Southwest Indian Ridge had been spreading consistently slower than the other two ridges forming the triple junction. The varied spreading rates lead to the eastward migration of the triple junction while the Central Indian Ridge and Southwest Indian Ridges migrated away from Africa and Antarctica. Between 80 and 64 Ma India and Madagascar continued to separate as did India and Antarctica and the triple junction moved both east and south. By 53 Ma (anomaly 22), the triple junction moved further east and to the north (Fisher and Sclater, 1983). At about 44 Ma (anomaly 19), as noted in the previous section, major spreading rate and direction changes have occurred on Central Indian Ridge and Southeast Indian Ridge. No contemporaneous major change was noticed on the Southwest Indian Ridge. Between 44 Ma to present rapid eastward migration of the triple junction was reported by Tapscott et al. (1980) and Sclater et al. (1981).

During the anomaly times A26 to A21 i.e., from about 60 to 50 Ma, which corresponds to the period documented in the present study, the eastward migration of the triple junction had been taking place. The trace of the triple junction separates the crust generated at different ridges and these traces can be mapped on the Indian, African and Antarctic plates. From the
study of kinematics of the three ridges the triple junction traces were predicted on the Indian, African and Antarctic plates (Sclater et al., 1981).

The triple junction trace as delineated by Sclater et al. (1981) extends into the present study area (Figure 7.8). The triple junction trace is typically marked by the change in the bathymetric character demarcating the crust generated by the different spreading centers. The recent SEABEAM surveys in the vicinity of the triple junction (Patriat and Larson, 1989; Munschy and Schlich, 1989) have mapped the triple junction trace. The reconstructions of the triple junction (Sclater et al., 1981; Tapscott et al., 1980; Fisher and Sclater, 1983), suggest rapid eastward migration of the triple junction. In order to maintain the RRR configuration the triple junction undergoes jumps and induces offsets along the ridges forming the junction (Munschy and Schlich, 1989).

The offset near anomaly A24 is very much less when compared to the observed offsets across anomalies A21 and A22 (Figure 7.8) across the 76°30'E fracture zone. The observed decrease in the offsets of the magnetic lineations across the 76°30'E fracture zone, when examined in the light of the triple junction trace on the Indian plate, indicates that the study area was affected by the evolution of the Indian ocean triple junction. Thus, it is implied that the seafloor in the study area which was generated by the fast spreading Southeast Indian Ridge, was influenced by the slower Central Indian Ridge
Figure 7.8 The identified magnetic lineations and fracture zones during the present study. The triple junction trace as delineated by Sclater et al. (1982) and the fracture zone trends on Southeast Indian Ridge (Royer and Schlich, 1988) are shown.
and was subjected to modifications as a consequence of the evolution of the triple junction.

7.3 IMPLICATIONS OF SEAMOUNT DISTRIBUTION

A total number of 80 seamounts are identified in the region based on echosounding data and multibeam sonar data. The heights of the seamounts range from 500m to 1850m. Locations of these seamounts are shown in Figure 5.9. The morphological features of some of these seamounts are studied in detail by Kodagali (1989) and Mukhopadhyay and Khadge (1990). These seamounts are situated on a lithosphere, with crustal ages varying from 50 to 60 Ma. As can be seen from the Figures 5.9 and 7.9, within the study area seamounts follow the fracture zone-trends. The 79°E fracture zone has distinct bathymetric expression as revealed by the Hydrosweep data (Figures 5.1 and 5.2), whereas the 76°30′E fracture has very little bathymetric expression and is only manifested by a small offset in the magnetic lineations. It was observed that in areas where detailed magnetic surveys have been made, the seamounts are located either on or very close to the fracture zones, even though the fractures have little bathymetric expression (Vogt, 1974; McNutt and Batiza, 1980). Based on seamount distribution in the Pacific Ocean, Batiza (1982), observed that the non-hotspot volcanoes have preferentially occurred either on or very near to the fracture zones and suggested that fracture zones may provide easy conduits for the seamounts. Although fracture zones are not necessary for the
Figure 7.9 Summary tectonic map showing the results from the present study with magnetic lineations, fracture zones and seamount locations along with results from other studies. The magnetic lineations along SEIR are from Royer and Schlich (1988). The inset shows the traces of the Reunion and Kerguelen hot spot traces in the Indian Ocean (after Duncan and Richards, 1991).
formation of seamounts, there may be local perturbations to the plumbing system or other zones of weakness that control the location of seamounts (Epp and Smoot, 1989).

Seamount distribution in the Central Indian Basin can be viewed with respect to the known hotspot traces in the Indian Ocean (inset, Figure 7.9). The reconstructions of the Indian plate considering the relative plate motions (Molnar et al., 1987; Royer and Sandwell, 1989) assuming fixed hotspot reference frame, resulted in the Indian plate motion which can be compared to the Reunion and Kerguelen hotspot traces (Duncan and Richards, 1991). Recent studies of the legs 115 and 121 of the Ocean Drilling Program have shown evidences to suggest that the Chagos-Laccadive ridge system and Ninetyeast Ridge are the traces of the Reunion and Kerguelen hotspots, respectively (Duncan and Richards, 1991).

At present it is not clear whether the seamounts identified in the study area are related to the hotspot activity or they are of non-hotspot type. The magnetic signature along the line CB-14 is disturbed and the anomalies are uncorrelatable. This may be due to the presence of the 73°E fracture zone. Furthermore, the region along the 73°E fracture zone, which was further south and west during the geological past, could have come under the influence of the Reunion hotspot. The disturbed and uncorrelatable magnetic signature along the 73°E fracture zone over a wide zone, the reconstructions of the Indian plate, and the proximity of the Reunion hotspot trace suggests
that the seamounts in the vicinity of 73°E fracture zone probably originated from the Reunion hotspot. Whereas, the seamount clusters along 76°30'E and 79°E fracture zones, which occur preferentially along the fracture zones appear to be controlled by the fracture zones.

It was suggested that the hotspot generated seamount chains will tend to form a mosaic of smaller chains along the fracture zones (Vogt, 1974) and the oceanic volcanoes belonging either to the hotspot or non-hotspot type may share a common genetic source (Batiza, 1982).

On the basis of the high resolution seismic tomographic models Anderson et al. (1992), have indicated that the hotspots may not be a narrow pipe like features but have wider sublithospheric dimensions as represented by broad low velocity anomalies (LVA's) within the upper mantle and suggested that the mantle upwelling is controlled by appropriate lithospheric stress conditions.

The fracture zones in the study area could have provided suitable lithospheric conditions for the formation of seamounts. Based on these observations it is suggested that the seamount chains observed along the fracture zones at 73°E, 76°30'E and 79°E represent smaller chains of major hotspots.
7.4 MORPHOTECTONICS OF THE 79°E FRACTURE ZONE

The detailed high resolution swath bathymetric data has revealed various morphotectonic elements in the vicinity of the 79°E fracture zone as detailed in Chapter 5.0. Integration of these results with other geophysical data and corroboration of the tectonic processes observed at various fracture zones enabled to decipher the detailed morphotectonics of the 79°E fracture zone.

Studies of microseismicity at the intersections of transform faults, spreading ridges and fracture zones demonstrate that while numerous microearthquakes occur along the transform faults and spreading ridges, the fracture zones are seismically inactive (Francis et al., 1978; Forsyth and Rowlett, 1979; Rowlett, 1981). It was also suggested that fracture zones are not the zones of weakness and the fractured crust is only a remnant surface expression of past tectonic activity along the transform faults. Furthermore, it was also shown that there is no significant slip along the fracture zones and that the lithosphere bends in the vicinity of the fracture zone, in order to contain the topographic expression (Sandwell and Schubert, 1982). The thermal contrast across the lithospheric sections at the fracture zone setup a small-scale convection cell in the underlying asthenosphere. The upwelling limb of the convection cell uplifts the hot younger side, while the downwelling limb depresses the colder older side resulting in a ridge trough topography along the fracture zone (Craig and MacKenzie, 1986).
The curvature of the bathymetric contours, seen on the maps (Figures 5.2 and 5.4), reflect the surface expression of the ridge-transform intersection (RTI) topographic signature. The clay cake deformation experiments of Macdonald et al., (1986) illustrated the development of riedel shears at the transforms. These riedel shears accommodate significant dip-slip component and evolve into normal faults that curve at the RTI resulting in sharply curved bathymetric scarps.

Multibeam bathymetric studies using Seabeam near a slow slipping transform fault (Vema fracture zone), (Macdonald et al., 1986) and fast slipping transform boundary (Clipperton transform fault), (Gallo et al., 1986) have shown the existence of a transform tectonized zone of varying dimensions along the length of the active transform fault. The transform tectonized zone (TTZ) as delineated at various transform faults represents a zone that had experienced strains associated with strike slip tectonism during the passage along the plate boundary. The TTZ contains braided network of small-scale faults and intrusive contacts of different ages. The width and geometry of the TTZ will vary as a function of regional changes that arise due to the pole of relative motion behavior and the local changes that include adjustments in slip geometry due to the structural incompatibilities in the transform fault environment (Fox and Gallo, 1989). A zone of smooth topography along the 79°E fracture zone has been delineated from the swath bathymetry is shown on the tectonic summary map (Figure 7.10). The varying width of the zone of smooth topographic expression, perhaps
Figure 7.10 Summary tectonic map of the area encompassing the 79°E fracture zone. Bathymetric lineations as delineated from the multibeam bathymetric study, the identified magnetic anomalies (numbered solid lines), the fossil transform tectonized zone (stippled area) and the location of seamounts (solid dots) are shown.
resulted from the fracturing in the transform fault environment and subsequent sedimentation, represent a fossil transform tectonized zone.

The prominence of the bending of E-W bathymetric lineations in the northern portion of the 79°E fracture zone area in comparison to the southern portion (Figures 5.1, 5.2 and 7.10), indicates principal difference in the formation of these features. The features observed at the 79°E fracture zone are compared with fast slipping transform faults like Clipperton (slip rate 110 km Ma⁻¹) (Gallo et al., 1986) and Siqueiros (slip rate 126 km Ma⁻¹) (Fornari et al., 1989) and the slow slipping Vema transform fault (slip rate 24 km Ma⁻¹) (Macdonald et al., 1986). In order to do this, a situation is visualized in which the present fracture zone was a part of active transform fault system. Macdonald et al., (1986) proposed that the transform fault evolves from having young newly accreted lithosphere on one side which is thin and easily deformed to having old, thick lithosphere on both sides which does not deform easily and acts as a strain guide to confine the transform to a narrow zone. The detailed multibeam studies of fast slipping transform faults in the Pacific (Fox and Gallo, 1989) and slow slipping transforms in the north Atlantic (Fox and Gallo, 1986) have described the processes occurring at a wide range of slow (< 30 km Ma⁻¹), intermediate (60 km Ma⁻¹) and fast slipping (> 60 km Ma⁻¹) transforms. The slow slipping and large offset (> 10 Ma) transforms exhibit relatively simple geometry and the accretionary processes are constrained by thick lithosphere edges, proximal to these ridge transform intersections (RTI’s) (OTTER, scientific team, 1985;
Langmuir and Bander, 1984; Macdonald et al., 1986; Fox and Gallo, 1986; Pockalny et al., 1988). The tectonic characters of the small offset (< 2 Ma) and fast slipping (>60 km Ma\(^{-1}\)) transforms are governed by thin, hot and weak edges of the lithosphere at the RTI. At these transforms the shear zone geometries exhibit complex patterns and the accretionary processes are not significantly perturbed by the thin edge of the lithosphere (Searle, 1983, 1986; Gallo et al., 1986; Madsen et al., 1986; Kastens et al., 1986; Langmuir et al., 1986; Gallo et al., 1989).

Based on the identified magnetic anomalies and the spreading rates inferred in the present study, the fracture zone segment between A23-A26 evolved during a fast slipping transform fault environment with a slip rate of 160 km Ma\(^{-1}\), whereas the segment between A22-A23 fits into a medium rate transform fault with a slip rate of 72 km Ma\(^{-1}\). In this scenario the oblique extensional features which are prominent in the segment with fast slipping rate transform fault environment can be considered as due to the strong coupling at the RTI. The thin and hot lithospheric edges at the fast slipping transform which are easily deformed, facilitate strong coupling. Where as the subtle presence of the oblique extensions in the south which is the result of a medium rate slipping transform fault can be attributed to the weak coupling at the RTI. In this case the relatively cold and thick lithospheric edges are encountered at the RTI resulting in weak coupling. In the case of strong coupling at the RTI, the curvature of the riedel shear faults is more prominent and result in curved bathymetric scarps that are oblique to the transform,
where as spreading center parallel features develop when the coupling is weak. It was also suggested that the temporal variations of shear coupling at RTI may vary with the waxing and waning of axial magma chamber (Macdonald et al., 1986). In case of Clipperton transform fault the oblique extensional features were delineated at the fracture zone adjacent to the transform fault zone (TFZ) on a seafloor of 0.5 Ma (Gallo et al., 1986). Similar features existing on older lithosphere (53 Ma) in the present study area indicate the clear preservation of these morphotectonic features.

Because of the thin sediment cover, it has been possible to clearly document various morphotectonic features over the fracture zone. In general, sediment thickness of about 100 m was reported in the Central Indian Basin (Geological map of the Indian Ocean, Heezen et al., 1978). Sediment thickness varying from 0.5 to 0.8 km with a decreasing trend towards south has been inferred based on the multichannel seismic data, north of the study area around 5°S (Bull, 1990). However, pulses of ancient sediment injection from the Bengal Fan have been found as far south as 7°S (Emmel and Curray, 1984) and geochemical evidence suggest the terrigenous input down to 10°S (Nath et al., 1989; Martin-Barajas et al., 1991). These studies suggest that in the study area, the northern portion can be expected to have more sediment thickness. The subdued topographic expression of the fracture zone in the north is therefore probably due to the sediment infill.
Fracture zones in the ocean basins represent a unique asymmetrical contact between two sections of the same lithospheric plate. The wide range of spreading rates and age contrasts under which the fracture zones evolve result in markedly different lithospheric sections with substantially different thermomechanical structures.

Studies of seismicity at transform faults, mid-ocean ridges and fracture zones indicate that fracture zones are seismically inactive. It is therefore considered that the fracture zones are the inactive traces of the transform faults. The various morphotectonic features that were formed in the transform fault domain remain along the fracture zones. It is expected that the nature of these morphotectonic expressions undergo minor modifications due to the processes of subsidence and contraction with age. It is therefore possible to relate the evolution of various morphotectonic features observed at fracture zones, as a consequence of the tectonic processes that took place during the ridge-transform-ridge (RTR) environment. An RTR plate boundary represents a complex and dynamic interface. Along the RTR, strike slip strains are accommodated and the accretionary processes such as, asthenosphere upwelling, melt generation and melt emplacement are truncated. The interplay of these tectonic processes determines the overall morphology of the RTR boundary. These parameters affect the lithosphere on either side of
the transform fault. Expression of these processes can be found along the fracture zones and can be mapped with high resolution swath bathymetric tools.

As discussed in the earlier sections the morphotectonic characters at a fracture zone are primarily dependent on the plate tectonic setting under which it evolves. The plate tectonic setting can be inferred from the regional geophysical data such as magnetics and conventional bathymetry. The offset distance between the magnetic lineations, the age of the ocean floor and the spreading rates can be measured.

The age contrast across the fracture zone leads to the formation of intraplate lithospheric sections with different thermal structure and thickness. The differences in the age and thermal structure subject the lithospheric sections to differential subsidence on either side of the fracture zone. The contrast in rate of subsidence of the lithospheric sections coupled with the thermomechanical interactions, result in flexure of the lithosphere at the fracture zone.

The thermal contrast between the lithospheric sections across the fracture zone would also induce small scale convection cells within the asthenosphere with the hot upwelling limb on the younger side and the downwelling limb on cold, older side. Formation of these small scale
convection cell uplifts the younger side and depresses the older side and contributes to the persistence of ridge trough pair along the fracture zone.

Spreading rate and the rate of magma supply at various stages of evolution of the fracture zone control the coupling at the RTI. The coupling at the ridge transform intersection between the cold older lithospheric section and the newly accreted younger side depends on the mantle weld. During the faster spreading regimes the accretionary processes are in general faster and provide strong coupling at RTI. For the segment evolved in a slow spreading episode the coupling is weak. These factors lead to the presence or absence of bending of the bathymetric lineations at the fracture zone.

A model for the evolution of morphotectonic features at a fracture zone is proposed and the factors influencing the evolution are summarized in Figures 7.11a and 7.11b.
Figure 7.11 (a) The proposed model: summary of factors influencing the morphotectonics at a fracture zone.
Figure 7.11 (b) Schematic illustration of the evolution of the fracture zone.