Chapter 2

Background

2.1 Introduction

The Burmese-Andaman-Sunda arc defines the 5500 km long boundary between the Indo-Australian and Eurasian plates, from Myanmar to Sumatra and Java to Australia. The plate boundary separates the north east moving Indian plate from the southeast Asian plates that includes Burma and Sunda microplates (Fig. 2.1). Global plate tectonic reconstructions suggest that the Indian plate converges obliquely toward the Asian plate at a rate of 54 mm/yr (DeMets et al., 1994a) at N22°E. The effect of oblique convergence has resulted in the formation of a sliver plate between the subduction zone and a right lateral fault system, which has evolved as the Sumatra Fault system in the southern part of the subduction zone and the Sagaing Fault in the Myanmar, as well as the opening of the Andaman Sea (Curray, 2005).

Varying degrees of tectonism, seismic and volcanic activity occur along this subducting margin. The Andaman-Sunda section of the subduction zone had produced many large earthquakes in the past, some of which have also generated destructive tsunamis. Significant historical earthquakes occurred in this region are the 1679 (M ~7.5) in the west coast of North Andamans; 1797 (M ~8.4), 1833 (M 9.0), 1861 (M 8.5), 1907 (M ~7.8), 1935 (Mw 7.7) from Sumatra region; 1881 (Mw 7.9) off Car Nicobar and 1941 (Mw 7.7) off Middle Andaman (Fig. 2.3). While
Figure 2.1: Map showing the major tectonic segments of Andaman-Sumatra subduction zone. Area of interest for this thesis lies in a zone between 5-15°N and 92-98°E. Rupture area for 2004 earthquake is marked in yellow, historic rupture areas also marked. (modified from Kayal, 2004.)
these large earthquakes have ruptured only a few hundreds of kilometers (200-300 km) of the plate boundary, the 2004 earthquake has ruptured more than 1300 km length of the arc, which included regions that have ruptured in the past as well as the intervening unbroken patches. Sumatra region has also witnessed three recent earthquakes; two of these occurred on 4th and 18th of June 2000 (MW 7.8), located south of the 1833 rupture (Abercrombie et al., 2003) and the third one on 2002 (MW 7.3), north of March 2005 rupture.

While Indonesian part of the trench has been extensively studied in the recent years using variety of techniques including GPS based ground deformation studies, as well as coral micro atolls studies (Natawidjaja et al., 2004), similar work is only in early phase in the Andaman part of the arc (Rajendran et al., 2007). This chapter review the seismicity and tectonics of the region and evaluates the ongoing seismogenic processes followed by a discussion on the December 26, 2004 megathrust earthquake.

2.2 Previous studies

The first organized oceanographic study of the Andaman Sea was conducted by Alcock (1902), and later on by Sewell (1925). But there are reports on onshore and offshore surveys way back to 1595 by Van Linschoten and then by Mallet (1895) on the Barren Island volcano. Study of the geology and origin of the Andaman-Nicobar ridge started with Rink (1847), who suggested that this ridge had been formed of sediments uplifted from deep ocean floor. Hochstetter (1869) pointed out that the same ridge extended southward as the outer arc ridge off Sumatra and
Java. Wegner (1966) was the first to postulate a rift origin of the Andaman Sea. First systematic geophysical survey of the Andaman Basin was conducted by Weeks et al., (1967) as part of the International Indian Ocean Expedition, onboard U.S Coast Guard ship Pioneer. They carried out marine magnetic, gravity and limited sub-bottom seismic profiles, and using these data they suggested that the Barisan range of northern Sumatra extends into the Andaman Sea to 10°N.

The earlier available reference to the geology of the islands is made by Helfer (1840), who described the rocks of Ritche's archipelago. Later, Rink (1847) divided the rocks of the Nicobar into three groups namely, 1) brown coal formation, 2) igneous rock and 3) older alluvium. Ball (1870) recorded the geology of the vicinity of Port Blair and correlated these rocks with that of the Nicobars. According to him the sedimentary rocks of the South Andaman are cross-cut by serpentinite intrusions. Tipper (1911) mapped parts of North Andaman Island and Nicobar. During 1959 and 1960 a team led by C. Karunakaran of Geological Survey of India, conducted investigations for sulphur on the Barren and Narcondam Islands and mapped parts of South Andaman Island. The central Andaman Sea is 100-200 km wide trough and marked by steep and elongated sea valleys and sea mounts such as the Nicobar Deep, Barren-Narcondam volcanic islands, Invincible bank, Alock and Sewell sea-mounts (Rodolfo, 1969). Geological expeditions of scientific interest were initiated by Survey of India way back to 1957 (Bandyopadhaya et al., 1971). Fitch (1972) brought out that the NE movement of India was resolved or partitioned into two large components: dextral strike-slip on the Sagaing Fault (5 cm/yr) and high rate normal subduction along the Sunda-Andaman trench (4
The western base of the Andaman-Nicobar trench is filled with sediments of Bay of Bengal (Curay et al., 1979). The structure along the arc in the Andaman-Nicobar region is dominated by east dipping nappes having folding, while intense folding is observed off Sumatra (Weeks et al., 1967; Curay et al., 1979). Curay et al., (1979) proposed the existence of an independent sliver plate absorbing the oblique motion of India with respect to southeast Asia. Eguchi et al., (1979) inferred collision of the Ninety-east ridge with the Sunda trench in the middle or late Miocene. They also reported that the ridge-trench collision transmitted compressional stresses into the back-arc area and collision of India with Eurasia exerted a drag on the back-arc region causing opening of the Andaman Sea.

2.2.1 Seismicity and tectonics

The seismicity and tectonics of Andaman-Nicobar region was analyzed by many and led to many conclusions on the processes that control the subduction. The regional seismicity pattern itself reflects different tectonic regimes within this Island arc system, namely the thrust dominated subduction front, the strike-slip faulting along the west Andaman transform and the extensional processes within the Andaman spreading center. Sinvhal et al., (1978) analyzed the time-space seismicity evolution and the associated neotectonics and reported a seismic gap north of the Andamans. This paper also deals greatly with the mechanism associated with the 1941 Middle Andaman earthquake. Uyeda and Kanamori, (1979) related the back-arc spreading activity in the Andaman Sea to leaky transform tectonics. The geometry of the Wadati-Benioff zone has been studied in detail for the Andaman region by many workers (Verma et al., 1976; Uyeda and Kanamori, 1979; Mukhopadhyay, 1984; Mukhopadhyay and Dasgupta, 1988; Mukhopadhyay, 1988; Ni et al., 1989; Gupta et al., 1990; Mukhopadhyay and Krishna, 1991). Gravity and seismicity data along the Burmese-Andaman arc suggests the presence of a subducted slab (Verma et al., 1978; Gupta et al., 1990). Srivastava and Chaudary (1979) analyzed epicentral data from USGS for a period of 1917-1974, and constrained the dip of the subducting interface (Fig. 2.2). A geodynamic perspective of the region was put forth by Surendra Kumar (1981) using the epicentral data as well as the available focal mechanism data. From marine magnetic anomaly studies Liu et al., (1983) identified a fossil spreading ridge beneath the Nicobar fan.
Curray et al. (1982) inferred that the Andaman Sea and the central lowlands of Burma are parts of a single structural province. Chandra (1984) inferred segments that divide the major tectonic regimes in the Burmese-Indonesian arc based on several lines of evidence, which include change in trend and offset in arcs, bathymetry and sedimentation, faulting in the region, change in composition and trend of the line of volcanoes, spatial distribution of earthquakes and change in dip of Benioff zone. Mukhopadhayay (1984) studied extensively on the shallow earthquakes at the Andaman spreading ridge and reported evidence for extensional stress within the subducting lithosphere, and reported the under-thrusting of Indian lithosphere below the Burma plate down to a depth of 200-220 km. Further, he observed that some of the north-south faults developed on the main islands are seismically active. Hamilton (1979) reported that the subduction is more penetrative (~600 km) under the Sunda arc further south.

The decoupled transcurrent movement of the Burmese sliver plate is presented
in detail by Maung (1987) wherein the opening up of Andaman basin in the mid Miocene is discussed. Banghar (1987) using epicenters of 345 earthquakes between 1967 and 1982 brings out the under-thrusting along the arc. McCaffrey (1988) extensively studied the active tectonics of the Eastern Sunda and Banda arcs and the back-arc thrusting using centroid depths and fault plane solutions. Rajendran and Gupta (1989) studied the stress orientations in the Andaman-Nicobar region and found the maximum compression in the region is NE-SW to N-S, compatible with the motion of the Indian plate. They reported the along-arc variations in the stress orientations. Surendra Kumar (1981) used gravity data in addition to seismicity information to better constrain the geometry of the subduction zone.

Using micro-earthquake survey data Harjono et al. (1991) studied the transtensional Sunda strait and stress tensors for the area were computed and confirmed that the Sunda strait is an extensional tectonic regime as a result of the northwestward movement of the Sumatra sliver plate along the Semango Fault Zone. Guzman-Spezilae and Ni (1993) studied the opening up of the Andaman Sea, and suggested that the strain due to the opening of the Andaman Sea spreading system is seismic. The subduction azimuth varies from frontal/normal subduction in the Java to oblique subduction in the Sumatra-Andaman region (McCaffrey, 1988; Malod et al., 1995).

Crustal evolution and sedimentation history of the Bay of Bengal studied by Rao and Kumar (1997) pointed out that Sumatra-Andaman arc is an intermediate stress subduction zone with relatively few large magnitude events. A detailed study on the neotectonics of the Sumatran Fault was conducted by Sieh
and Natawidjaja (2000) which brought out many new insights into this 1900 km long fault. Radhakrishna and Sanu (2002) using focal mechanism data of shallow earthquakes from this region identified major tectonic segments along the arc and inverted these mechanisms for the stress tensors there. Shallow seismicity and available source mechanisms in the Andaman-west Sunda arc and Andaman sea region suggest distinct variation in stress distribution pattern both along and across the arc in the overriding plate. They inferred that the oblique plate convergence, partial subduction of 90°E ridge in north below the Andaman trench and the active back-arc spreading are the main contributing factors for the observed stress field within the overriding plate in this region. In the background of spurt of seismicity in North Andamans, Rajendran et al. (2003) and Kayal et al. (2004) analyzed the spatio-temporal evolution of seismicity and the volcanism associated (Fig. 2.4) with the archipelago. Rajendran et al. (2003) reported that this area has entered into a phase of renewed activity and suggested an association with the down-dip extension of the subduction earthquakes.

Using multi-beam swath bathymetry, magnetic and seismological data Kamesh Raju et al. (2004) and Kamesh Raju (2005) brought out a three phase tectonic evolution of the Andaman basin. They presented that the current full rate spreading at Andaman Sea (Raju et al., 2004) is about 38 mm/yr, 327° relative to the present north (Curray, 2005). Using data sets from decades long oceanographical surveys and other geophysical studies, Curray (2005) gives a detailed information on the Andaman spreading, its evolution, and in addition an overall review of the tectonics.
An estimation of the relative motion between the plates is of significant importance in studies quantifying the deformation within the Indian Ocean and understanding the relative high level of seismicity there (Gordon et al., 1990). New constraints have been obtained decades later from seismology and GPS measurements, after the pioneering efforts of Fitch (1972) and Curray (1979) using early global plate kinematic models. Modern day geodetic techniques have become increasingly prominent in studies of plate boundary deformation. Using GPS data collected in Bangalore (IISC) and at several global GPS sites, Freymueller et al. (1996) found agreement between the present day Indian plate motion with that predicted by NUVEL-1A (DeMets et al., 1994a). Later on analysis by Chen et al. (2000) and Shen et al. (2000) suggest that the motion of Bangalore is 5-7 mm/yr slower than NUVEL-1A. A significant motion of a large Sundaland block (Sumatra, Java, Vietnam, China, Borneo) with respect to Eurasia was discovered by Chamot-Rooke and Le Pichon (1999); Simones et al. (1999); Michel et al. (2001). Paul et al. (2001) estimated the convergence of 14 mm/yr between India and Port Blair from a single campaign mode control point at CARI, Port Blair and forms the first GPS geodetic observations from this Island archipelago. According to this study, CARI samples only 50% of the India/Andaman convergence due to the unknown degree of coupling there, and this makes it difficult for an independent estimate of the full Andaman velocity. A velocity of 20 mm/yr was established across Sagaing Fault (Vigny et al., 2003) using GPS.

Strain rate field from Andaman Sea region was studied by Kreemer et al., 2003 and obtained new constraints on the partitioning of the compression along the
Sumatra-Andaman trench and the extension along the spreading segments. They predicted the direction of extension, as it occurs along the spreading segments, and showed that it is consistent with earthquake slip vectors ~N30°W. They also reported that there is a significant component of right-lateral shear in agreement with the seismotectonics indicated by the focal mechanisms. Bird (2003) compiled from the available deformational rates and fault locations, proposed a pole of rotation for the Burmese plate with respect to the Sunda plate as 103°E, 13.9°N, 2.1°/Ma.

### 2.2.2 Significant pre-seismic earthquakes

The Andaman-Nicobar section of the Sunda-Andaman plate boundary has produced many large and destructive earthquakes in the past (Table 2.1 and Fig. 2.3). Among the earlier earthquakes, those in 1847 (7.5<M<7.9), 1881 (Mw 7.9) and 1941 (Mw 7.7) are significant (Bilham et al., 2005). An earthquake also occurred in the Arakan coast of Burma on April 2, 1762 (Chhibber, 1934) from the description of felt effects in northern part of the Bay of Bengal and Arakan, it appears that the event was close to the Irrawady delta. Another large earthquake is reported to have occurred in the North Andaman on January 28, 1679 (Iyengar et al., 1999). Felt reports from the Burmese coast as well as parts of the east coast of India (Temple R. C., 1911) suggest this to be comparable to the 1941 earthquake in magnitude and rupture extent (Rajendran et al., 2007). According to Hochstetter (1866) the 1847 earthquake happened near to Kondul Island, an island in between the Little Nicobar and Great Nicobar, with an aftershock duration of 5 weeks, which makes it comparable to 1941 and 1881 earthquakes. Location of the earthquake is still
Figure 2.3: Significant earthquakes and their rupture areas along the Sunda-Andaman plate boundary. An, Andaman islands; Nb, Nicobar islands; Sm, Simuleue island; Bt, Batu island; Mt, Mentwai island; Ac, Aceh province; Ni, Nias island; NER, Ninety East ridge; Sfz, Sumatran fault zone); WFR, Warton fossil ridge; IFZ, Investigator fracture zone. Filled arrows represent Indian and Australian plate velocities and direction (DeMets et al., 1994). Modified from Briggs et al., 2006.
speculative, as there is no other source of information. Bilham et al. (2005) rejects the idea of a strike-slip earthquake to the west of Nicobars, as no earthquakes exceeding $M_w 7.2$ is reported on the adjacent transform fault, and it is much more probable that an event of such a magnitude can occur on a reverse fault on the west of or beneath the islands.

Oldham (1884) compiled a detailed account on the 1881 Car Nicobar earthquake. It caused a tsunami surge not exceeding 0.75 cm at Car Nicobar (Rogers, 1883) and a wave height of 0.25 m was measured from the tide gauge stations at Madras (Chennai) on the east coast of India (Ortiz and Bilham, 2003). Using the tsunami travel times, Ortiz and Bilham (2003) analyzed the source mechanism of this earthquake. The earthquake is calculated to have occurred near and west of Car Nicobar with two reverse slip ruptures. The larger measured 150x60 km, and dipped $25^\circ$E with a slip of 2.7 m equivalent to a $M_w 7.9$ earthquake (Ortiz and Bilham, 2003). The smaller was equivalent to $M_w 7.0$ and occurred some 50 km north of the larger patch. According to them the location of the rupture was so close to Car Nicobar, its western edge raised $\sim 50$ cm relative to the eastern shore.

June 26, 1941 earthquake happened a year before the occupation of Japanese in the Andaman Islands and is reported to have caused an uplift of $\sim 1.5$ m along the western margin of the middle Andaman and subsidence of the same magnitude along the eastern margin, an observation not validated by any direct measurements (Jhingran, 1953). There are no reports of any tsunami impact either from the Andaman-Nicobar Islands or from the east coast of India (Rajendran et al., 2007). The event affected the middle and south Andaman regions, including the town of
Port Blair. The central watch tower of the cellular jail in Port Blair collapsed along with a hospital and other masonry structures. There are also eyewitness accounts on the subsidence of Ross Island, subsequent to this earthquake. Using the available aftershock extent information Bilham et al. (2005) infer the rupture extend between $11^\circ$N and $13^\circ$N and computes a slip of 3 m on a 50 km wide 150 km long down-dip rupture.

The M 8.7 earthquake of 1833, is reported to have ruptured about 550 km segment of this arc; it also generated a tsunami (Natawidijaja et al., 2004). Natawidijaja et al., (2006) re-estimated the 1833 magnitude to 8.9-9.0 based on the rupture extent they measured based on the emergence and subsidence of the coral microatolls there. Briggs et al., (2006) fixed the magnitude to M 9.0 (see, Fig. 2.3). Another great earthquake of 1861 (M 8.5) broke a segment north of the equator, also triggering a tsunami. The 1833 and 1861 earthquakes and the attendant tsunamis occurred before the introduction of harbour tide gauges in most parts of the world and no tidal gauge data exist for these events. However, better documentation exists for the 31st December 1881 earthquake which caused run-up in eastern coast of India. This earthquake is also the oldest for which slip geometry has been inferred. See, Table 2.1 for major significant pre-seismic earthquakes from Andaman-Sumatra subduction zone.

2.2.3 Volcanism

Subduction along the Andaman-Sumatra trench system has given rise to a discontinuous belt of submarine volcanic seamounts. The andesitic volcanoes of Bar-
Table 2.1: Significant Pre-seismic earthquakes that occurred in recent and historic times in and around the Andaman-Nicobar region. Rupture areas and earthquake locations, where-ever available are plotted in Fig. 2.3. 1Iyengar et al., 1999, 2Imperial Gazetteer of India, 1909, 3Ortiz and Bilham, 2003, 4Bapat et al., 1983, 5NEIC, USGS, 6Rajendran et al., 2003.

<table>
<thead>
<tr>
<th>Date</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Magnitude</th>
<th>Region</th>
</tr>
</thead>
<tbody>
<tr>
<td>28/Jan./1679</td>
<td>12.50</td>
<td>92.50</td>
<td>7.5(^1)</td>
<td>M/N. Andamans</td>
</tr>
<tr>
<td>31/Oct./1847</td>
<td>07.30</td>
<td>94.75</td>
<td>7.5 &lt; (M) &lt; 7.9</td>
<td>Off Car Nicobar</td>
</tr>
<tr>
<td>31/Dec./1881</td>
<td>09.25</td>
<td>92.70</td>
<td>7.9(^3)</td>
<td>Off Car Nicobar</td>
</tr>
<tr>
<td>16/Nov./1914</td>
<td>12.00</td>
<td>94.00</td>
<td>7.2(^4)</td>
<td>S-W of Barren Island</td>
</tr>
<tr>
<td>28/Jan./1929</td>
<td>10.00</td>
<td>93.00</td>
<td>6.5</td>
<td>Little Andamans(^5)</td>
</tr>
<tr>
<td>01/Aug./1929</td>
<td>10.00</td>
<td>93.00</td>
<td>6.5</td>
<td>Car Nicobar(^2)</td>
</tr>
<tr>
<td>09/Dec./1929</td>
<td>04.90</td>
<td>94.80</td>
<td>7.2(^4)</td>
<td>Sumatra</td>
</tr>
<tr>
<td>19/Mar./1936</td>
<td>10.50</td>
<td>92.50</td>
<td>6.5</td>
<td>Little Andaman(^5)</td>
</tr>
<tr>
<td>14/Sep./1939</td>
<td>11.50</td>
<td>95.00</td>
<td>6.0</td>
<td>East of Car Nicobar(^5)</td>
</tr>
<tr>
<td>26/Jan./1941</td>
<td>12.00</td>
<td>92.50</td>
<td>7.7(^4)</td>
<td>West of M. Andaman</td>
</tr>
<tr>
<td>08/Aug./1945</td>
<td>11.00</td>
<td>92.50</td>
<td>6.8</td>
<td>North of Little Andaman(^5)</td>
</tr>
<tr>
<td>23/Jan./1949</td>
<td>09.50</td>
<td>94.50</td>
<td>7.2(^4)</td>
<td>East of Car Nicobar</td>
</tr>
<tr>
<td>17/May./1955</td>
<td>06.70</td>
<td>93.70</td>
<td>7.3(^4)</td>
<td>East of Great Nicobar</td>
</tr>
<tr>
<td>14/Feb./1967</td>
<td>13.70</td>
<td>96.50</td>
<td>6.8</td>
<td>Andaman Sea(^5)</td>
</tr>
<tr>
<td>20/Jan./1982</td>
<td>06.95</td>
<td>94.00</td>
<td>6.2</td>
<td>Great Nicobar(^5)</td>
</tr>
<tr>
<td>20/Jan./1982</td>
<td>07.12</td>
<td>93.94</td>
<td>6.1</td>
<td>Great Nicobar(^5)</td>
</tr>
<tr>
<td>13/Sep./2002</td>
<td>13.08</td>
<td>93.11</td>
<td>6.4</td>
<td>South East of Diglipur(^6)</td>
</tr>
</tbody>
</table>
Figure 2.4: Cone of Barren Island volcano as on May, 2002 (view from west). See Fig. 2.1 for location. Inset shows the composite eruption rate, smoothened using a moving average filter, shows an accelerated eruption ~50 years after the 1941 earthquake. (Rajendran et al., 2003).

Barren and Narcondam Islands are prominent among them; the Narcondam being now extinct, but barren is still marked by an active volcano and lie on the neogene inner volcanic arc. It erupted in March, 1991 after lying dormant for about two centuries (see Fig. 2.4-inset for eruption history). The first known historical eruption was on 1787 when a cinder cone grew in the center of pre-historical caldera. Intermittent eruptions were there on 1832, 1991 and 1995. Presently the island is just 3 km across, with a reported maximum elevation of ~400 m (Haldar et al., 1992).

Further south, this volcanic chain is represented by the Barisan range in Sumatra, and in North, the trend is correlated with the chain of volcanoes in Burma (Curray et al., 1982). Global observations on earthquake-volcano interactions, suggest large-scale eruptions following large earthquakes over periods of 7-50 years at dis-
tances of 30-150 km, and 30-50 years at distances up to 1000 km (Hill et al., 2002). Rajendran et al., (2003) analysed the eruption history of Barren (Haldar et al., 1992) in wake of the spurt of earthquake activity in the North Andamans and suggested a 50-year correlation may be applicable to the Barren Island volcano, located within the 250 km of the 1941 epicentre (Fig. 2.4, inset). Recently Kumar et al., (2006) reported a minor eruption of Barren volcano on May, 2005 using remotely sensed satellite imageries.

2.3 26 December, 2004 $M_W$ 9.3 Great Sumatra-Andaman earthquake

The 26 December main-shock rupture began at 3.36°N, 96.0°E at a depth of 30 km at 00:58:53 GMT (National earthquake information centre (NEIC), United States geological survey (USGS)). Harvard moment tensor solution suggests thrusting on a shallowly dipping plane ($8^\circ$), striking $328^\circ$. It ruptured a 1300 km long plate boundary north-westward along the Sunda trench and the Andaman trench and caused static offsets as far as 4000 km away from the epicenter (Banerjee et al., 2005). The aftershock zone extends to nearly $15^\circ$N. Distribution until the beginning of March 2005 suggests little change in the extent of the aftershock zone (Fig. 2.5). One notable feature is the absence of aftershock activity north of about $15^\circ$N latitude, a zone that has not generated much earthquakes in the past.

Significant vertical displacements of the sea floor were responsible for a tsunami that propagated throughout the world's oceans (Bilham, 2005). Tsunami runup heights were measured at $\sim25$ m near Banda Aceh region of Sumatra (Borrero,
Figure 2.5: Location of December 26, 2004 earthquake shown by centroid moment tensor (CMT) solution beach ball, and aftershocks (black dots) till 1st March, 2005. Epicentral data source: NEIC, USGS, CMT: Harvard University CMT database. Extent of rupture zone can be clearly marked by the extent of aftershocks.
2005), 6 m in Thailand (Titov et al., 2005), and 3-12 m along the coast of Sri Lanka (Liu et al., 2005). Models of body-wave amplitudes in the first few minutes of the rupture indicated that the slip on the rupture surface was heterogeneous, varying from several meters in many places to more than 20 m near the epicenter (Ammon et al., 2005; Lay et al., 2005; Park et al., 2005). The characteristics of this earthquake were studied by many workers using various methods, to constrain the moment magnitude of the event, its rupture duration, direction, extent and the associated mechanism.

Focal mechanisms of the aftershocks suggest arc-normal compression (thrust faults) along the subduction front and extension (normal and strike-slip faulting) in the back-arc region (Mishra et al., 2007). Other than the arc normal compression expressed by the thrust faulting all along the subduction front, one notable feature is the cluster of aftershocks in the back-arc region, characterized by normal and occasional strike-slip faulting. Lay et al. (2005) note that although such swarms have occurred in this region in the past, the one associated with the 2004 earthquake is the most energetic swarm ever observed, globally. During the two months that followed, nearly 1000 shallow earthquakes have occurred here among which about 600 events occurred during a short duration from January 27–30, 2005 and nearly 100 of them were of magnitude >5.0 (NEIC), and this activity continued there till September 2005. This region is a transition area between the Sumatra Fault and developing Andaman back-arc spreading center. This high aftershock activity along the back-arc ridge-transform faults indicates accompanying slip partitioning along that boundary (Engdahl et al., 2007). According to them, most of these swarm
earthquakes were characterized by shallow depth right-lateral strike-slip, but also including a number of normal faulting mechanisms oriented perpendicular to the predominant strike-slip faults in the region along a linear trend. There is a marked transition in the distribution of aftershocks at ~5.5°N and it broadly corresponds to changes in the physical properties of the plate interface (Kennet and Cummins, 2005). Also, this transition corresponds to the second region of significant slip in many main-shock rupture models (Ammon et al., 2005; Ishii et al., 2005).

Various modes of rupture mechanisms were put forth by many including single and multi segmented rupture, involvement of slow slip in the northern segment and the varying rupture dimensions along the arc. Some common features of almost all rupture models are, a rupture of ≥1300 km along the Sumatra-Andaman megathrust, with rapid slip concentrated in the southernmost 500 km (Ammon et al., 2005; Lay et al., 2005; Banerjee et al., 2005; Chlieh et al., 2007; Rhie et al., 2007). Ammon et al., (2005) suggests the possibility of slow slip in the first 50-60 sec of rupture in the southern region near the epicenter. Several studies have required a slower rupture velocities for the northern segment of the rupture, with periods greater than 600 sec (Banerjee et al., 2005; Lay et al., 2005; Stein and Okal, 2005; Tsai et al., 2005; Singh et al., 2006). Some studies, again using various seismic and geodetic datasets, have found that the slow slip and low rupture velocities in the northern region are not necessary to explain the data (Ishii et al., 2005; Tsai et al., 2005; Vigny et al., 2005; Banerjee et al., 2007; Chlieh et al., 2007).

The centroid moment tensor of the 26 December, 2004 Sumatran event was of point source mechanism and of moment magnitude of (Mw) 9.0 (Dziewonski et al.,...
1981). The magnitude of the earthquake has been a matter of energetic debate, with most estimates falling in the range of $M_w$ 9.0-9.3. Stein and Okal (2005) estimated a moment of $10^{23}$ Nm ($M_w$ 9.3) based on normal mode excitation, although Banerjee (2005) argue that this is overestimated because they assumed a point source with too large a dip angle. Ammon et al., (2005) have produced three different model slip distributions, each of which has $M_w$ 9.0-9.2. Park et al., (2005) using normal mode spectral data find that a model with $M_w$ 9.1. Braitenberg and Zadro (2007) compare the free oscillation amplitudes of 1960 Chile and 2004 Sumatra-Andaman earthquake and give a direct comparison that Sumatra-Andaman event was smaller than the Chilean event by a factor between 1.5-3.0. Using an extension of the empirical Green's function, Choy and Boatwright (2007) demonstrated that the second half of the rupture radiated less high-frequency energy than the first half of the rupture. The rupture process of the Sumatra-Andaman earthquake lasted for approximately 500 sec. The moment energy release by the rupture processes took place in the first 250 sec, with some minor patches of high frequency release up to 8 min after the onset of $P$ (Lomax, 2005; Kruger and Ohrnberger, 2005 and Ishii et al., 2005). Several finite fault inversions (Ammon et al., 2005) using low frequency surface wave data derived moment rate functions that had similar durations of 500-600 sec. Seismic rupture models which based entirely on analysis of body waves were unable to resolve the rupture in the northern segment because of the mixing of the arrivals from that segment with the secondary phases arriving from the intial phases of the rupture (Stein and Okal, 2005). All the authors argued that the major source of additional moment release from a slow slip in the northern
segment of the aftershock zone (Stein and Okal, 2005; Ammon et al., 2005; Bilham, 2005; Lay et al., 2005a). Tsai et al. (2005) estimated a similar magnitude using a CMT modelling approach, with 5 distributed point sources along the length of the rupture. Later, doubts have been cast on the slow slip hypothesis, as many later analyses showed that almost entire seismic energy release is from normal speed seismic rupture (Ishii et al., 2005; Neetu et al., 2005). The published models show a wide variation in features such as the depth and overall length of significant rupture.

The largest earthquake to follow the great event (Mw 8.6) occurred on 28 March. This earthquake had a similar mechanism as the 26 December earthquake, showing predominantly thrusting on a shallow-dipping (7°) fault plane. The rupture was about 300 km long, as defined by the extent of aftershocks. However, this one did not generate a huge tsunami like the December earthquake. It has been suggested that the March earthquake did not breach the sea floor, resulting in the transfer of lesser energy to the water column. Further, the earthquake occurred under relatively shallow water, displacing lesser volume of water.