Geochemical and Isotopic studies of sediments from the Andaman Islands and the Andaman Sea

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Declaration

I, Neeraj Awasthi, hereby declare that the work presented in this thesis entitled "*Geochemical and Isotopic studies of sediments from* <u>the Andaman Islands and the Andaman Sea</u>" is original and has not formed the basis for the award of any degree or diploma by any university or institution. I have acknowledged the work and material taken from other sources and solely own the responsibility for the originality of the contents.

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Dedicated to my parents, my wife and my family



Our team members on Andaman and Nicobar Islands studies (From left to right): Mr. Neeraj Awasthi, Dr. H.C. Sheth, Mr. Alok Kumar, Dr. J.S. Ray and Dr. Rajneesh Bhutani.

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Abstract

The prime focus of the present study is to understand the evolutionary history of Andaman Subduction Zone by unravelling the evidences preserved in the sediments. For this, we studied ash deposits preserved in the sedimentary records and determined the timing of past major volcanic activities in the Andaman region. The oldest record of volcanism from the Andaman region comes from the tephra interbedded in the Eocene and Mio-Pliocene age sedimentary rocks on the Andaman Islands. From our study on seven discrete ash layers in the core raised from the Andaman Sea, we established that the ash layers are uniquely sourced from the nearby Barren Island Volcano. We reconstructed the eruptive history of this volcano by dating foraminifers (AMS ¹⁴C dating) in sediment layers and found that the seven ash layers represent major eruptions at ~72, 71, 62, 24, 17, 12, and 8 ka. Isotopically correlating the precaldera volcanics exposed on the volcano to the uppermost ash layer (AL-1) in the core, we infer that the caldera of Barren Island volcano is younger than 8 ka. The sedimentary records preserved in the Andaman Islands (Late Cretaceous to present) and the Andaman Sea (Pliocene to present) are also studied in order to understand the tectono-sedimentary processes occurring in the subduction zone environment. The study of sedimentary rocks from the Andaman Islands clearly suggests that the Mithakhari Group sediments, deposited during the early to middle Eocene, were derived predominantly from mafic igneous sources comprising suprasubduction ophiolites and volcanic arc rocks with minor contributions from the Himalayan/Indian Shield sources. In comparison, the Andaman Flysch Group sediments appear to have been derived from mixed sources with dominance of Himalayan sources. The local arc/ophiolite sources possibly contributed >80% of sediments during the deposition of rocks of the Mithakhari Group, whereas the same sources contributed about 60-80 % during the deposition of the Andaman Flysch Group. We believe that the substantial increase in the sediment input from the rising Himalaya during the deposition of the Andaman Flysch Group was result of large scale weathering, erosion and transportation of sediments through the paleodrainage system developed along arc and suture zone. In order to understand the impact of climate on weathering and erosion, and supply of sediments in the past, sediments in the core (SK-234-60) from the Andaman Sea were also studied. The study reveals that the western Andaman Sea show relatively higher contribution of sediments from mafic sources of the Indo-Burman Ranges, while sediments from the Irrawaddy river system dominate in the sediments deposited in the central and eastern Andaman Sea. The elemental and isotopic compositions of the sediments show significant variations in the relative supply of sediments from sources over glacial-interglacial timescale. The changes observed reflect influence of climate on erosion in source areas and relative supply of sediments to sea. Significant increases in the relative contribution of sediments from mafic Indo-Burman sources at ~8 kyr, ~20 kyr (LGM), ~36 kyr, ~44 kyr, ~52 kyr and ~58 kyr are related to the weakening of the Asian summer monsoon, which restricted material contribution from the Himalayan source. Also, at ~6 kyr, ~10 kyr, ~15 kyr, ~46 kyr, ~54 kyr and ~60 kyr and ~72 kyr there were higher sediment contributions from higher Himalayas and continental Myanmar sources through Irrawaddy and Ganga-Brahmaputra rivers. These could have been resulted from intensification of Asian summer monsoon, which in turn could be correlated to the global events of warm climate. The overall contribution of sediments derived from the Indo-Burman sources increased since the LGM. This is inferred to be related to the strengthening of the surface currents in the north-western Andaman Sea due to increase in the sea level after the LGM which resulted in reopening of "Preparis North Channel" through which substantial quantity of sediments from the NE Bay of Bengal entered into the Andaman Sea. From the study of uplifted coastlines of two islands, seismic history of the islands for past 9 kyr was reconstructed. Earlier reports and our results reveal that the Andaman region had experienced a major earthquake and associated tsunami event at \sim 500 (or \sim 600) cal yr BP. Combining our data with the available data on such events in this region we have been able to determine that there have been at least 14 major landscape changing seismic events between ~40 kyr BP to present, with a hiatus between ~19.5 and ~8.5 cal kyr BP. We propose that, in a similar fashion as observed subsequent to the 2004 earthquake, the Andaman Islands have been experiencing tectonic upliftments in the north and subsidences in the south, for the last ~40 kyr, along the so called "pivot line" proposed by Meltzner et al. (2006).

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Chapter-1 Introduction

1.1 Subduction Zone

1.1.1 Plate Tectonics and importance of Subduction Zone

On the basis of physical properties, the silicate portion of the Earth has been classified into two distinct layers, known as: the lithosphere and asthenosphere. The lithosphere is the rigid outermost rocky part of the Earth, which comprises the crust and a portion of the upper mantle, while asthenosphere, underlying it, is the deeper part of the upper mantle and is comparatively weaker, hotter and more viscous. The lithosphere of the Earth is broken into several continental and oceanic tectonic plates, which constantly move with respect to each other. The driving force behind these movements is the mantle convection currents produced in the Earth's interior due its internal energy. The science which deals with the study of cause of, and relative motions of, lithospheric plates is known as plate tectonics. Seven major and several minor plates have been recognized. The motions of these plates are extremely slow but they continuously interact among themselves in both constructive as well as in destructive manner and reshape the surface morphology of our planet. Based on the relative motions of these plates, three types of plate boundaries have been recognized: divergent, transform and convergent.

At divergent plate boundaries, two plates move away from each other resulting in upwelling of mantle, which then melts to fill the gap and form new oceanic crust. Within continents, these plate boundaries produce rift valleys while between oceanic plates; these exist as mid-oceanic ridges (MOR). The transform plate boundaries are another type, where the plates slide past one another predominantly in horizontal direction which can be either sinistral or dextral. These are also known as conservative plate boundaries, since at these boundaries lithosphere is neither created nor destroyed. The transform faults are connected on both ends to other faults, ridges, or subduction zones. Most of the transform faults are located in the deep oceans where they are present as zigzags accommodating seafloor spreading, but a few present on land too, e.g., San Andreas Fault in USA.

At the convergent boundaries the lithospheric plates are destroyed. These are also known as subduction zones. At subduction zones, the colder and denser tectonic plates (subducting plates) sink into the mantle beneath comparatively hotter and lighter tectonic plates (overriding plates). Subduction zones are our planet's largest recycling system and are also the places where new continental crust gets generated on the overridding plate (Hawkesworth, 1993). These are also tectonically most active regions on our planet and often known for their association with volcanism, earthquakes and mountain building activities.

A Subduction Zone provides exposures and allows easy access to study the lower oceanic crust and upper mantle in place and provide a unique opportunity to study the magmatic products associated with the initiation of subduction. Study of rocks in convergent plate margins can provide answers to several questions pertaining to role of fluids in modifying the chemistry of recycled material, in metamorphism at such zones and in generation of partial melts in the mantle wedge. These can also help understand the pathways of elemental recycling into the mantle and processes that lead to chemical modification of the mantle. Sedimentary rocks deposited at such zones can reveal a lot about the tectonic processes and their timings in the development of important geomorphic features that shape the convergent margins.

1.1.2 Earthquakes and Volcanism in Subduction Zones

The earthquake and volcanic activities are indicators and consequences of the movements along the lithospheric plate boundaries and based on their worldwide distribution, tectonically active regions of the world and the margins of the lithospheric plates have been defined. The most destructive, high magnitude (M>6) and deepest earthquakes are known to occur at subduction and collision zones in comparison to divergent and transform boundaries which are characterized by smaller magnitude and

much shallower earthquakes. In subduction zones, the earthquake hypocentres lie in a plane known as the Wadati-Benioff zone, which represents the brittle region in the upper 10-20 km of the subducting lithospheric slab. Earthquakes at convergent margins many a times cause rapid deformation of the sea floor leading to tsunamis.

Subduction Zone volcanism has the greatest impact on humans, because of their violent eruptions and long term effect on global climate. At subduction zones, oceanic lithosphere, sediments and seawater reequilibrate with Earth's mantle and then return to surface through arc volcanoes (Stern, 2002). The source and type of magma genarated depends on tectonic environment, e.g., angle of subduction, depth of subduction, thinness of mantle wedge etc. At subduction zones, magma generated has its origin from the partial melting of heavily metasomatized mantle wedge in presence of fluids derived from subducting oceanic crusts. The compositions of lava generated are complex, but incipient island arcs tend to be more basaltic in composition, whereas mature volcanic arcs tend to be more andesitic in composition. The volcanoes of subduction zones are usually stratovolcanoes, which eject large amounts of pyroclastics that get dispersed to large distances due to high energy eruptions.

1.1.3 Structure of a subduction zone

Despite the fact that subduction zones are the most well studied regions of the globe, the processes that lead to the evolution of various geo-morpho-tectonic features in these zones are not very clearly understood. This is partly because, the efforts in this regard are limited by absence of exposures of active regions and lack of understanding of important tectonic controls that lead to morphological changes (Dickinson and Seely, 1979; Underwood and Bachanan, 1982; Moore, et al., 1982). The current understanding of subduction zones is based primarily on geological and geophysical studies done on active and ancient subduction complexes now exposed on land (Karig, 1974; Seely, et al., 1974; Karig and Sharman, 1975). Although, a complete set of morphological components of subduction zone is rarely exposed at a single margin, studies on numerous subduction zones suggest a generalized morphology consisting of features,

which in order includes: a peripheral bulge (or outer swell), trench, accretionary prism, forearc basin, volcanic arc and backarc basin (van der Pluijm and Marshal, 2004) (Fig. 1.1). In the following paragraphs brief description on some of these features are given.



Fig. 1.1: A schematic diagram showing various morphological components of an ocean- ocean plate subduction zone: viz. trench, accretionary prism, forearc basin, volcanic arc and backarc basin. (Source: http://en.wikipedia.org/wiki/File:SubZone.jpg).

(A) Trench

In a convergent plate boundary, a peripheral bulge is encountered on the subducting plate before the trench, where the surface rises to form a broad arc due to flexural rigidity of the lithosphere (Uyeda and Kanamori, 1979). But the 'Trench' marks the actual boundary between the downgoing plate and the overriding plate where the former bends from its horizontal position to dipping position (Fig. 1.1). The trench is long but narrow topographic depressions of the sea floor, created by the gravitational pull of the relatively dense subducting plate pulling the leading edge of the plate downward. Trenches, near continents, are sometimes buried under great volumes of sediment

supplied by large rivers or glaciers. Such trenches lack bathymetric expressions. Trenches distant from influx of continental sediments lack accretionary prism and the inner slope of such trenches is commonly composed of igneous or metamorphic rocks.

(B) Accretionary Prism

An accretionary prism or wedge is formed from sediments that are accreted on the overriding plate near the trench (Fig. 1.1). Materials incorporated in accretionary wedges include: ophiolitic rocks, pelagic sediments and deformed turbidites and in some cases the erosional products of volcanic island arcs. An accretionary prism develops either by frontal accretion whereby sediments are scraped off from the edge of downgoing oceanic plate, or by underplating of subducted sediments and perhaps oceanic crust (ophiolites) along the shallow parts of the subduction decollement. Ophiolite rocks representing obducted uppermost part of oceanic lithospheres (Dewey and Bird, 1971; Coleman, 1977; Nicolas et al., 1989) are categorized into two types-subduction related and subduction-unrelated types (Dilek and Furnes, 2011). Subduction related ophiolites are of suprasubduction zone varieties, develop during the initial stages of subduction prior to the development of any volcanic arc (Pearce et al., 1984) and are geochemically similar to island arc magmas.

Structurally, an accretionary prism is imbricate stack of fault slices formed by thrustcontrolled emplacement of discontinuous slices of ophiolites and trench fill sediments (Condie, 1989; Platt, 1986; Moores et al. 1984). The continuous accretion of new thrust wedges of ophiolite occur from the bottom, which expose the older thrust wedges to the seafloor (Platt, 1986). With continuous subduction and upliftment with time, a series of thrust slices of ophiolites were thus emplaced with complex folds in them. These thrusts slices show listric faults, steeply dipping towards the arc (Dickinson, 1977; Roy, 1992; Pal et al., 2003). Several isolated slope basins develop in between these thrust-bounded structural ridges of ophiolite, which trap the material derived from the eroding thrust wedges in a deep-water environment. Sediment transport in this environment is controlled by submarine landslides, debris flows, and turbidity currents. Submarine canyons transport siliciclastic sediment from beaches and rivers of nearby landmasses down the upper slope which is further carried via channels and a series of faultcontrolled basins. Frontal accretion results in younger sediments defining the outermost part of the accretionary prism and the oldest sediments defining the innermost portion.

The accretionary prisms are also sites for mud volcanoes and fluid activities, wherein Earth's most dynamic and complex interactions between aqueous fluids and rocks occur. The upper lithosphere and sediments of the subducting plate, and sediments accreted to the forearc contains large volume of water entrapped in their pores and fractures. These sediments having hydrous minerals are progressively squeezed with increasing pressure and temperature as they are subducted which forces fluid out along the decollement and numerous thrust faults up into the overlying forearc. Water released by dehydration accompanying phase transitions, is another source of fluids. These fluids along with solid materials may travel through the accretionary prism diffusely, via interconnected pore spaces in sediments, or may follow discrete channels along faults. Sites of venting may take the form of mud volcanoes or seeps and serve as windows to the subducting slab. These fluids are dominated by water but also contain dissolved ions and organic molecules, especially methane. At greater depths the fluid released alters the bulk composition of the mantle wedge and trigger partial melting reactions which are responsible for magma generation (Peacock, 1990).

(C) Magmatic/Volcanic Island Arc

As discussed above during the process of subduction, when the descending slab saturated with water and volatiles comes in contact with mantle wedge, causes release of fluids which cause partial melting in the mantle wedge. The melts generated rise from the point of melting through the overriding plate and erupt on surface to form volcanoes (Fig. 1.1). These volcanoes generally occur in arc-shaped chain of volcanic islands, often parallel to trench and popularly known as Magmatic/Island Arc. The volcanoes of Japan, Aleutian, Mariana and Lesser Antilles are some of the best examples of volcanic arcs.

(D) Fore-Arc and Back-Arc Basins

In later stages of subduction, a forearc basin develops between the accretionary prism and volcanic arc due to extensional processes (Fig. 1.1). Typically, it is filled with siliciclastic sediments (flysch) from the adjacent landmass and material trapped from oceanic crustal sources. Subsequent shallowing of the basin results in deposition of calcareous sediments along with frequent pyroclastic influx from volcanic arc sources. It is observed that the strata of a forearc usually cover the top of the accretionary wedge and/or trapped oceanic crust or submerged parts of volcanic arc. Sometimes, outer part of forearc gets uplifted to form outer arc ridge running parallel to it (ten Veen, and Kleinspehn, 2003; Yanagi, 2011).

A back arc basin develops on the inner concave side of the volcanic arc (Fig. 1.1). Several hypotheses have been proposed for its development. Earlier it was believed that convergent plate margins were zones of compression, thus zones of extension (back-arc basins) above subduction zones were not expected. It was only in 1970, when Dan Karig, based on several marine geologic expeditions to the Western Pacific, first proposed that some convergent plate margins were actively spreading. Karig (1971) proposed that development of back arc basin, due to extension, resulted from the forcible intrusion of basaltic mantle diapirs. Packham and Falvey (1971) believed that the spreading was because of passive magma upwelling, due to the trench suction force resulted from a roll-back of the trench. The term 'roll-back of the trench' describes the backward motion of the subduction zone relative to the motion of the plate which is being subducted. This causes regional extensional stresses in the back arc region and makes the lithosphere considerably thinner (Chase, 1978; Fein and Jurdy, 1986). Hsui and Toksoz (1981) and Jurdy and Stefanick (1983) suggested that extension occurs due to formation of secondary convection cells in the mantle wedge overlying the benioff zone, induced by descent of the subducting slab.

Back-arc basins are usually very long and relatively narrow, with spreading rates which vary from very slow, a few cm/yr to very fast >10 cm/yr (Taylor and Martinez, 2003). The ridges in back-arc basins erupt basalts derived from depleted mantle similar to those erupted at the mid-ocean ridges (Hart et al., 1972). The main difference lies only in their magmatic water content and correlated trace elements, which is higher in back-arc basin basalt magmas (Stolper and Newman, 1994). Sedimentation in the back arc basin is strongly asymmetric, with most of the sediment supplied from the active magmatic arc and adjacent landmasses. Not all subduction zones have back-arc basins, some of the active back-arc basins are found in the Marianas, Tonga-Kermadec, S. Scotia, Manus, N. Fiji, and Tyrrhenian Sea regions, but most are found in the Western Pacific.

1.1.4 Sedimentary records in Subduction Zones

Deposition of sediments within a subducton zone occurs in the trench, and in the fore arc and back arc basins. Although the trenches receive sediments throughout the evolution of a Subduction Zone, the oldest trench materials are generally found obducted onto the accretionary wedges. Since a part of the trench sediments ultimately undergo subduction into the mantle, study of these rocks can shed some light on the recycled crustal components. Sediment deposited in the forearcs, which are generally found on the accretionary wedges, can reveal a lot about the evolution of the subduction zone. In particular these may provide information on timing and processes that led to the formation of accretionary wedge, and magmatic arc. In addition determining the sources of these sediments can lead us to understand the tectonic evolution of the surrounding landmasses and the drainage patterns therein. Sediments deposited in the backarc basins usually represent the latest phase of sedimentation in subduction zone and likely to yield information about the events that occurred subsequently to the opening of the backarc sea. These can also provide information about the volcanism in the arc during recent past.

1.1.5 Paleoseismic records in accretionary wedges

As already discussed, subduction zones are known for their high seismic activities and associated tsunamis, which are great natural hazards and cause large scale devastations of life and property. Many times paleoseismology studies have changed comforting notions about the seismicity of a region by unearthing evidences of extremely large earthquakes and tsunamis. Therefore, paleoseismological studies of subduction zones become necessary to understand the causes, effects and patterns of large earthquakes. In active subduction zones, local geomorphic features such as coastal terraces of accretionary prisms are often formed by tectonics and not by local fluctuations in the sea-level. In many cases, these coastal terraces have been proven to be very useful in understanding the relationship between earthquakes, active tectonics and individual faults in a region (e.g., Chappell, 1974; Kayanne et al., 2007; Ota, 1991, 1992; Rajendran et al., 2007, 2008).

1.2 Andaman Subduction Zone (ASZ)

The ASZ in the northeastern Indian Ocean is a part of the Burma-Andaman-Sumatra-Java (BASJ) subduction zone (Fig. 1.2). The ASZ and its southward extension to Sumatra are famous for high magnitude earthquakes. It is also well known for being home to some of the large and destructive volcanoes of the world. The young and seismically very active ASZ is one of the few convergent margins that contains the complete set of the morphological features of an ideal subduction zone setting. ASZ is believed to have come into existence after the early Cretaceous break-up of the Gondwanaland, when the Indian plate moved northward and started subducting under the Eurasian plate (Curray, 2005). At ~50 Ma, the northern edge of the Indian plate collided with the Eurasian plate (Rowley, 1996; Hodges, 2000), which subsequently gave rise to the Himalayas in the north. It is also believed that in the eastern and southeastern margin, the subduction of Indian oceanic plate continued under the Eurasian plate (Acharyya, 1992). Studies suggest that the northward motion of the Indian plate was very rapid, 16-20 cm/yr between Late Cretaceous and early Paleogene (Patriat and Achache, 1984; Besse and Courtillot, 1988; Klootwijk et al., 1992; Lee and Lawver, 1995) and it covered a distance of about 2,000 km to 3,000 (Molnar, 1986).

Figure 1.3 shows schematic cross section across the ASZ along the A-B transect on Fig. 1.2. The regional tectonic setting of the ASZ is very complex. Chronological evidences from sedimentary record in the Bengal fan and obducted ophiolites (on the Andaman and Nicobar Islands) suggest the subduction in this region got initiated during the middle-late Cretaceous (Cenomanian, ~95 Ma) (Roy, 1992; Das Gupta and Mukhopadhyay, 1993; Moore et al., 1982; Pederson et al., 2010). The subduction has resulted in the formation of morhpho-tectonic features like trench, accretionary prism, a volcanic arc and a back-arc basin with spreading center. The convergent margin is marked by the sinuous-arcuate Sunda-Andaman trench. Along the Andaman and Nicobar islands it is called the Andaman trench. The Andaman trench marks the active subduction zone where, at present, the northeast-moving Indian plate is obliquely subducting below a small tectonic sliver plate that has been referred to as the Burmese microplate (Dasgupta and Mukhopadhyay, 1993; Ortiz and Bilham, 2003; Kayal et al., 2004). Burmese microplate, believed to have formed during the Neogene (Curray, 2005), is separated from the rest of the Eurasian plate along a transform boundary running N-S through the Andaman Sea. The Andaman accretionary prism is located on this plate and has been discussed in detail later in this chapter. The Andaman Sea is an active extensional basin that encompasses backarc and forearc basins of the Andaman subduction zone, separated by a volcanic arc ridge containing subaerially exposed volcanoes of Barren Island and Narcondam. The complex tectonic setting has resulted in the development of several N-S trending thrust and strike-slip faults in the region.

The Andaman Subdution Zone has experienced several large-magnitude earthquakes in recent history (Rajendran et al., 2003; Bilham et al., 2005). The most devastating was the recent M9.1 earthquake that occurred on December 26, 2004, which generated a massive tsunami that killed approximately 229,800 people in continents surrounding the Indian Ocean. Most of the earthquakes in this region are generated by thrust and strike-



Fig. 1.2: Map showing northeastern Indian Ocean with major geological and tectonic features. The Indian Plate is obliquely subducting under the Burmese microplate along the Andaman Trench at the rate of 4 cm/yr (Gahalaut et al., 2010), arrow showing the direction of plate movement .Further south the Indian Plate is subducting under rest of Sunda Plate along the Sunda-Banda Trench. Red triangles represent volcanoes of volcanic arc whereas dotted line represents connected Sagaing and Semangko dextral strike–slip fault systems.

slip faulting (Eguchi et al., 1979; Banghar, 1987). The seismicity of the region is marked by earthquakes with magnitudes varying from 4 to 8. Several reports and published data suggest that both shallow (10 km or less) and deeper (\geq 250 km) earthquakes are common in this region (e.g., Mukhopadhyay 1984, Eguchi et al., 1979, Banghar, 1987, Ortiz and Bilham, 2003, Kayal et al., 2004). These earthquakes are generally concentrated along the trench, on the east and west of the exposed accretionary prism and along the spreading center of backarc basin.

1.2.1 Andaman accretionary prism

Underthrusting of the Indian plate beneath the Eurasian plate resulted in the upliftment of ophiolites and pelagic sediments that developed a 3000-5000 km long outer-arc ridge. This outer arc ridge and associated forearc constitute the Andaman accretionary prism. Continued subduction and resultant tectonic upliftment exposed the accretionary prism above sea level as islands of Andaman and Nicobar (Fig. 1.2 and 1.3). The exposed part of the accretionary prism, consisting of 572 islands, extends from 6° to 14° N latitudes and from 92° to 94° E longitudes. These islands are aligned in a north-south direction in an arcuate shape. The Andaman group of islands is separated from the Cape Negaris of Myanmar by Preparis Channel and from Nicobar group of islands by the Ten Degree Channel. Except a few, most of the islands possess long hilly ranges, narrow valleys and sandy beaches. The highest point, known as Saddle Peak (732 m), is located in the North Andaman Island. The dense tropical forests that extend across the terrain cover ~86% of the total area. Mangroves are present almost on the entire coastline and coral reefs surround most of the islands.

The interpretation of geology of this accretionary prism is complex due to folding, thrusting and dynamic environments of deposition. Earlier studies have shown that it mainly consist of Late Cretaceous ophiolites and their erosional products, along with siliciclastic, carbonate and volcanoclastic sediments overlying it (Pal et al., 2003). Figure 1.4 shows a model for the tectono-sedimentary evolution of the Andaman accretionary prism, as proposed by Pal et al. (2003). According to this model, the accretionary prism started forming sometime between late Cretaceous and late Eocene in a compressional regime through frontal accretion process with emplacement of ophiolites and deposition of ophiolite-derived clastics and trench sediments in small isolated basin. During the Oligocene with change of compressional regime to extensional, siliciclastic turbidites alternated with ash turbidites started depositing. In Miocene-Pliocene time, carbonate turbidites also started depositing in the forearc region.



Fig. 1.3: A schematic cross-sectional profile of the Andaman Subduction Zone (along A-B in Fig 1.2) showing Andaman Trench, Andaman Accretionary Prism, Barren Island Volcano and Andaman Sea Back Arc Spreading Ridge.



Fig. 1.4: A schematic diagram showing tectono-sedimentary events of Andaman Subduction Zone in different time periods (redrawn from Pal et al. 2003) (a) shows initial movement of subducting plate under overriding plate; (b) & (c) accretion and thrust controlled emplacement of ophiolite sheets and formation of trench-slope basins with deposition of ophiolite-derived clastics and trench sediments; (d) formation of extensional forearc basin on the rear side of wedge and deposition of siliciclastic turbidites alternated with pyroclastic flows (e) transition of siliciclastic to carbonate turbidites deposition along with pyroclastic flows.

The western, outer slope of the accretionary prism is very steep, rising sharply from sea level within a short distance, whereas the eastern margin is bounded by faults. The ophiolite thrust sheets trend N–S and show dip towards the east except some back thrusts which dip towards the west. A few E–W trending out-of-sequence thrusts have also been reported (Pal et al., 2003). The thrusts on the western margin of the prism have been reported to have a ductile character while brittle behaviour has been reported in few thrusts from the eastern margin (Pal et al., 2003).

The sedimentary rocks of the accretionary prism show variable bedding orientation. The lower formations deposited in trench-slope basins show folds with axial planes varying from inclined to vertical while upper formations deposited in forearc basin environment show folds with very regular geometry (Pal et al., 2003). Several SE–ESE-striking normal and reverse faults and a few right-lateral strike–slip faults have been reported within the accretionary prism (Dasgupta et al., 2003; Chakraborty and Khan, 2009).

The Andaman accretionary prism consists of several other interesting features like mud volcanoes, raised coral terraces and limestone cave deposits. There are at least 11 mud volcano fields on the Andaman Islands (Chakrabarti et al., 2006). These mud volcanoes constantly emit hydrocarbon gases, water and mud breccias and are known to occur in Middle and North Andaman islands and are linked to major east-dipping fault systems in the region (Roy, 1992). The raised coral reefs forming coastal terraces are present along numerous coastlines of the islands and appear to be quite young and believed to have been developed as a result of active tectonism in the region (Rajendran et al., 2007, 2008). The limestone cave deposits are found on Baratang Island in the middle Andaman.

1.2.2 Andaman Back Arc Basin

The Andaman back-arc basin is represented by the Andaman Sea, which extends 1200 km in N-S direction and 650 km in E-W direction. The Andaman-Nicobar ridge separates the Andaman Sea from the Bay of Bengal (Fig. 1.5). The central depression of the Andaman Sea, known as 'Central Andaman Basin,' is believed to have been

developed due to rifting and extension (Curray, 2005) (Fig. 1.5). The basin has a basaltic oceanic crust (Rodolfo, 1969). According to Khan and Chakraborty (2005), this rift basin opened up as a result of counter clockwise rotation of south-eastern part of the Eurasian plate. This initiated the extensional activity along the transform fault segments connecting the two major right-lateral strike-slip fault systems, Sagaing and Semangko, during the middle Miocene (Curray et al., 1979; Hall, 2002; Kamesh Raju, 2005). The basin is home to prominent morphological features such as Barren and Narcondam volcanic islands, Invisible Bank, and Alcock and Sewell seamount complexes (Rodolfo, 1969) (Fig. 1.5). The swath bathymetry, magnetic, and seismological studies carried out by Kamesh Raju et al. (2004) suggested that the true seafloor spreading commenced only at about 4 Ma, which resulted in the separation of the Alcock and Sewell seamounts. The central deep valley divided the basin into two parts, east basin and west basin (Fig. 1.5) (Curray, 2005). During these phases of extension and rifting, the western shelf of the Malay Peninsula (the Martaban shelf and further south) was under the influence of rapid terrigenous sediment influx (Kamesh Raju et al., 2004). The arc volcanism resulted in upliftment of the western Andaman Sea and gave rise to several bathymetric highs and N-S trending fault systems which made it more complex compared to relatively smoother east basin (Fig. 1.5). This further prevented the high sediment dispersal into the west basins while the east basin underwent subsidence and received high volumes of sediments (Kamesh Raju, 2005).

The Andaman Sea received and currently receives large amount of sediments from the north through Irrawaddy, Salween and Sittang rivers of Myanmar and occasional minor volcanic tephra from arc volcanism east of the Andaman-Nicobar ridge (Rodolfo, 1969) (Fig. 1.2). This large sediments input was probably got initiated as a result of high weathering and erosion in rising Himalaya, Indo-Burman ranges and Tibet in response to tectonics and climate change. The South Asian summer and winter monsoons are known to affect the physical and chemical weathering of these source regions whereas surface ocean currents driven by monsoonal winds, play an important role in dispersal of sediments in the Andaman Sea (Rao et al., 2005; Ramaswamy et al., 2004). The



Fig. 1.5: Bathymetric map of the Andaman Sea showing the islands, continents, seaways and prominent ocean-bottom ridges/highs and basins. Abbreviations for this figure: N= Narcondam Volcano; BI= Barren Island Volcano; IB= Invisible Bank; ASC= Alcock Seamount Complex; SSC= Sewell Seamount Complex.

Andaman Sea is a semi-enclosed basin and exchanges water and sediments with the Bay of Bengal and marginal seas of the Pacific Ocean through narrow channels such as the Preparis channel, Ten Degree channel, Great channel, Strait of Malacca (Rodolfo, 1969; Dutta et al., 2007; Keller, 1967; Ibrahin and Yanagi, 2006) (Fig. 1.5). Therefore, the sedimentary record of the Andaman Sea is likely to provide information about the changes in the tectonic and climatic conditions in the surrounding continents, volcanic activities in the region and changes in the ocean circulation patterns resulting from climate and sea-level fluctuations.

While the paleoceanographic and paleoclimatic studies from the Arabian Sea and Bay of Bengal are numerous (e.g., Anderson and Prell, 1991, 1993; Clemens, et al., 1991, Siroko et al., 1991, 1993; Naidu and Malmgren, 1995, 1996, 1999, 2005; Naidu, 2004; Rostek et al., 1997; Thamban et al., 2001, 2002, 2007; Singh et al., 2011; Cullen, 1981; Chauhan et al., 2001, 2004; Burton and Vance, 2000, Goodbred and Kuehl, 2000 etc.), only a handful of such studies exist from the Andaman Sea (Rodolfo, 1969; Rao, 1983; Curray, 2005; Naqvi et al., 1994, Ramaswamy et al., 2004; Rao et al., 1996, 2005; Chernova et al., 2001; Roonwal et al., 1997; Ahmad et al., 2000; Colin et al., 1998, 1999, 2006; Kurian et al., 2008; Rashid et al., 2007; Alagarsamy et al., 2010). In addition, studies from the Andaman Sea are not exhaustive compared to their counterparts in other parts of the Indian Ocean. These studies generally focused on clay mineralogy of the sediments (Rao, 1983), nature of organic matter (Ramaswamy et al., 2008), grain sizes and role of monsoon in their dispersal and transport (Rao et al., 2005; Ramaswamy et al., 2004), carbon and oxygen isotopic signatures on foraminifers (Naqvi et al., 1994; Rashid et al., 2007) and hydrothermal activity (Rao et al., 1996 Kurian et al., 2008; Chernova et al., 2001). Curray (2005) described the tectonic history of the Andaman Sea while Colin et al. (1998, 1999, 2006) discussed the weatheringerosional history and magnetic properties of sediments.

1.3 Rationale

1.3.1 Significance of this work

In Indian context, the Cenozoic time period is very important because many important geological events of global significance happened during this Era. These include the India–Asia collision, the disappearance of Neo-Tethys, evolution of the Himalayas, creation of currently active volcanoes of south and southeast Asia, evolution of the South Asian monsoon system and major changes in the oceanic circulation in the northern Indian Ocean. The paleogeographic positions of the Andaman Islands and Andaman Sea during these times makes them most suitable localities for preservation of records of most of these events. The sedimentary records from these terrains are likely to reveal a great deal of information about the events like closure of the Neo-Tethys, tectonics related to Indian Plate subduction, uplift, weathering and erosion in the Himalayas, the evolution of Bengal fan and the initiation of volcanism in this region.

In the absence of extensive studies from the Andaman Islands, the knowledge about provenance for the sedimentary rocks deposited on them has remained inconclusive (Karunakaran et al., 1968; Pal et al., 2003; Curray et al., 1979; Curray, 2005; Allen et al. 2007). The different tectono-sedimentary processes which resulted in development of these islands in the subduction zone environment and the depositional history of the sediments on them have remained largely unknown. A comprehensive geological and geochemical study on the sedimentary rocks is required to settle much of the outstanding issues related to the regional geology, stratigraphy and evolution of the Burmese microplate since its creation. This would not only resolve the issues pertaining to provenances but also help us determine if the Himalayan-Tibetan orogen contributed sediments to the Andaman Islands. If indeed the Himalayan-Tibetan orogen supplied significant detritus to these islands, then the sedimentary record can reveal a lot about the early evolution of the orogen and the development of river systems in India and Myanmar, those originate from the Himalayas.

The Quaternary was most eventful of all geologic times and has observed major climatic changes, fluctuations in eustatic sea level in accordance with changing continental ice cover (Chappell and Shackleton, 1986). Understanding of past climatic variations and oceanographic changes during the Quaternary is important to fully understand present day climate. Climate and tectonics both largely control the physical and chemical weathering in the source regions and contribution of sediments to the ocean. Thus, marine sedimentary record is the most important and easily accessible repository of palaeoclimatic variations. The study of sediments from the Andaman Sea would not only reveal the pattern of sedimentation in the basin but also throw light on the impact of climate on weathering and erosion, and supply of sediments in kilo year timescale. The study might also provide link between the South Asian and East Asian monsoons. Therefore, it is essential to delineate the sources contributing sediments to the Andaman Sea. The variations observed in provenance would likely to reveal information about climatic and tectonic changes those took place in the past.

Numerous volcanic and seismic activities are known to have shaped the morphology of the Andamans. The evidences of volcanic eruption are likely preserved in the sedimentary record of the region in the form of ash deposits. Ash can be linked to various eruptive centres in the region through isotope/geochemical fingerprinting. Once the links are established these records can be utilized to determine the timing of past major volcanic activities in the region. Also, the pattern of earthquakes and their effects on the geomorphic evolution of the region can be understood by studying the tectonically formed coastal terraces in Andaman and Nicobar Islands.

1.3.2 Aim of the thesis

Our interests in the Andaman Subduction Zone lie in understanding its evolution since the Paleocene to the present by gathering valuable information from the sedimentary records of the Andaman Islands and the Andaman Sea. In the present work we have made an attempt to use petrolographic, geochemical and isotopic techniques to determine the sources of sediments, their weathering history and control of climate and
tectonics on them. Since, such studies require a satisfactory time framework, we have used geochronological data from literature (for ages >50 ka) and dated younger records (oceanic) using radiocarbon method.

The specific objectives of this thesis work were to:

- decipher the past volcanic activities in the Andaman region through the sedimentary records preserved on the islands and in the sea.
- determine the timing of deposition of various formations on the islands, and sedimentation in the Andaman Sea.
- determine sediment provenances and drainage patterns that transported the sediments
- understand the role of climate and tectonics on sedimentation.
- understand the origin and evolution of the Andaman region.

To achieve the above objectives we studied the sedimentary records on the Andaman Islands and that in the Andaman Sea. For this purpose rock samples were collected from various formations on the islands and a sediment core was raised from the central (or eastern) Andaman Sea.

1.4 Structure of the Thesis

The present thesis is divided into six chapters.

The **First Chapter (Introduction)**, as already described above, provides an introduction to the Andaman subduction zone and various morphological features associated with it and introduces the scope of the work and its importance in understanding of the geology of the region. It also lists the major objectives of this work.

The Second Chapter (Geology, Samples and Analytical Methods) deals with geology of the Andaman Islands, field observations, sampling procedures and details of the samples collected from the Andaman Islands and core collected from Andaman Sea. Sample preparation methods for various geochemical techniques are discussed in this chapter. The analytical methods utilized in this work include:

- Radiocarbon dating by conventional and Accelerator Mass Spectrometry (AMS) methods.
- Petrographic and mineralogical studies of sedimentary rocks of the Andaman Islands using microscopy (thin section), and EPMA (for ash layers in the core).
- Major and trace elemental geochemistry of the core sediments and sedimentary rocks of the Andaman Islands.
- Sr-Nd isotopic ratio analyses in silicate fraction of the sediments.

The third chapter (Study of volcanic tephra deposits in the Andaman region) presents the analytical results of our study on ash layers, interbedded with marine sediments, in the core collected from the Andaman Sea. Here, I discuss the volcanic history of the region as recorded in the literature and as revealed by our results using isotopic and geochemical proxies.

In the **Chapter four (Provenance of sediments deposited in the Andaman Region)** an attempt has been made to delineate the provenances of the sediments deposited in the Andaman region through time using geochemical and isotopic tracers on sediments from the Andaman Islands and Andaman Sea and their implications in understanding tectonics, paleogeography, paleodrainage patterns, climate, erosion (of source areas) and sedimentary transport have been discussed.

In the Chapter five (Chronology of major terrace forming events in the Andaman Islands during the last 40 kyrs) results on the deformation history of the Andaman Islands have been presented. Focusing on tectonically formed coastal terraces and determining the timing of their formation from the exposed corals, we have been able to reconstruct the history of major earthquakes in these islands for the last 40 kyr.

This last **sixth chapter** (**Summary & Conclusions**) summarizes important findings of the study and describes future research directions in the Andaman region.

Chapter-2 Geology, Samples and Analytical Methods

2.1 Geology of the Andaman Islands

The Andaman group of islands includes about two hundred islands with an exposed area of 6,408 km² (Fig. 2.1). As mentioned earlier, in geological context, these islands represent the accretionary prism of the Andaman Subduction Zone which is believed to have been developed as a result of continuous subduction of Indian plate under Eurasian plate since the Late Cretaceous (Pal et al., 2003). According to Karig and Sharman (1975), an accretionary prism usually develops during a subduction as a result of scrapping off slices of ophiolite and pelagic sediments from the downgoing oceanic plate by the edge of the overriding plate. However, according to Pal (2011), the dismembered ophiolite slices present in the Andaman accretionary prism did not form in the above fashion, instead are generated from the mantle wedge in a suprasubduction zone environment at the time of initiation of subduction. Subsequent to obduction of ophiolites and associated pelagic sediments, various tectno-sedimentary processes resulted in formation of forearc basin and deposition of clastic sediments in it, in both deep and shallow water environments (Allen et al., 2007).

The focus of most of the earlier geological work on the Andaman and Nicobar Islands was on establishing the geology and stratigraphy of the region (Rink, 1847; Hochstetter, 1869; Ball, 1870; Oldham, 1885; Tipper, 1911; Gee, 1927). In the Andaman accretionary prism setting, because of lack of exposures and high degrees of folding and faulting, it has always been difficult to determine the stratigraphy accurately. The earliest proposed stratigraphy of these islands came from the work done by pioneer workers like Rink, (1847); Hochstetter, (1866); Ball, (1870); and Oldham, (1885). However, their studies were restricted only to a few islands. Oldham, (1885) was first to give a comprehensive description of the Andaman geology by dividing it into an older Port Blair Series consisting sandstones, shales, coal seams, conglomerates, limestones

and jasper beds and a younger Archipelago Series, consisting of limestones, calcareous sandstones and white clays separated by volcanic rocks and serpentinites.



Fig. 2.1: Map of the northeastern Indian Ocean showing major geological and tectonic features along with locations of Andaman Islands and volcanoes of the Andaman Sea and Southeast Asia (solid triangles). 4 cm/y vector on the map shows the direction and rate of movement of the Indian Plate (Gahalaut et al., 2010).

The other geologists who made major contributions in improving the geological understanding of the Andaman stratigraphy were Tipper (1911), Gee (1927), Jacob (1954), Chatterjee (1967, 1984), Parthasarathy (1984), Roy et al. (1988), Bandopadhyay and Ghosh (1998), Acharyya et al. (1990), Acharyya (1997, 1998), Chakraborty and Pal (2001) and Chakraborty et al. (1999, 2002). In all these works although the lithological descriptions were similar, the proposed stratigraphic schemes were different with changes in the names of various formations. During 1960s, biostratigraphic principles were used along with lithostratigrapy to constrain depositional ages of various sedimentary units in the islands (Guha and Mohan, 1965; Karunakaran et al., 1968a & b). In mid-1970s, Oil and Natural Gas Commission (ONGC) of India started a detailed seismic reflection study across the island chain and correlated their results with those known from subsurface (through drilling) and outcrop stratigraphy, and placed the geology of the region within the context of the accretionary prism setting. In a series of papers, Srinivasan and his coworkers (e.g., Srinivasan and Azmi, 1979; and Srinivasan, 1977, 1979, 1984, 1986, 1988) published the first comprehensive work on the Archipelago Group or series. In the last few years there have been attempts to understand the origin and emplacement of ophiolites and other magmatic rocks, evolution of siliciclastics (Pal et al., 2003; Ghosh et al., 2009; Pederson et al., 2010), paleontological, petrological and geochemical evolution of chert sequences, and some of the ash (volcanoclastic) deposits (Jafri et al., 1993; Shastry et al., 2001; Srivastava et al., 2004; Bandhopadhay, 2005; Pal et al., 2010a). Pederson et al. (2010) and Sarma et al. (2010) constrained the age of the Andaman ophiolites to ~95 Ma by U-Pb dating of magmatic zircons. Through field, petrography and use of multiple isotopic tracers Allen et al. (2007) attempted to determine age and provenance of the sedimentary rocks of the Andaman Islands.

The currently accepted stratigraphy of Andaman Islands was established on the basis of the data collected and observations made by earlier workers on limited exposures in accessible areas, drill cores, road cuttings, seismic profiles etc. The stratigraphy followed in this work is given in Table 2.1 and comprises five groups, which from bottom to top consists of rocks of the Ophiolite Group and the overlying sedimentary rocks of the Mithakhari Group, the Andaman Flysch Group, the Archipelago Group and the Nicobar Group grouped together with unclassified arc volcanic and recent sedimentary deposits (Curray, 2005) (Fig. 2.2). In the following paragraphs we have briefly described each of these groups.

2.1.1 Ophiolite Group

The Ophiolite Group is the lowermost unit in the Andaman and Nicobar Islands (Fig. 2.2). The north-south trending discontinuous thrust slices of ophiolites are present throughout the Andaman Islands in association with sediments, low-grade metamorphic rocks and tectonic melanges. The intensely schistose greenschist to amphibolite facies metamorphic rocks consist of metasediments (quartzites, quartz-mica schists and phyllites) and metabasics (amphibole-bearing chlorite-epidote carbonate schists) (Pal et al., 2003). These rocks occur either as blocks in a melange zone or as discontinuous patches along the thrust contacts. Melanges include fragments of basalt, ultramafic rocks, metamorphic rocks, brecciated sandstone and chert embedded in sheared basaltic and serpentinized matrices (Pal et al., 2003). In the Andaman Ophiolite suite all the members of a classic ophiolite stratigraphy are present with the sequence of a plutonic complex (harzburgites, dunites), intrusives (gabbros and plagiogranites), extrusive lava series (pillow lavas of dacitic to andesitic and tholelitic basalt composition) and marine pelagic sediments (red clay, jasper, red/white chert, cherty/micritic limestone and reddish-brown and purple shale-mudstone beds) except for the sheeted dykes which have been identified only in some zones that are heavily affected by faulting and folding (Halder, 1985; Ray et al., 1988; Roy, 1992; Bandopadhyay and Ghosh, 1998). Shales of tuffaceous origin, cherty limestone with thin layers of basaltic flows, deformed thin beds of fine-grained sandstones and siltstones have also been reported within the Ophiolite Group and these posses soft-sediment deformation features like the slump folds and cross-cutting by normal and thrust faults (Allen et al, 2007).

2.1.2 Mithakhari Group

Karunakaran et al. (1968a) first introduced the name "Mithakhari" to describe the group of rocks deposited immediately above the Ophiolites. On the basis of the bed geometry and sedimentary structures, Pal et al. (2003) recognized many different lithofacies in this group, which include disorganized and graded matrix supported conglomerate, graded pebbly sandstone, massive and thick bedded sandstone, plane laminated and cross-stratified sandstones, interbedded sandstone and mudstone, massive to faintly laminated shale and interbedded shale and coal. This 1.4 km thick sedimentary unit (Ray, 1982) is mainly divided into three formations with the Lipa Black Shale at the base, the Hope Town Conglomerate unit in the middle and the Namunagarh Grit at the top (Fig. 2.2).

(a) Lipa Black Shale

The Lipa Black Shale is described as a minor unit with very few outcrops (Allen et al, 2007) and mainly includes pyritiferous black shales, thin coals and gypsum layers (Fig. 2.3a).

(b) Hope Town Conglomerate

The Hope Town Conglomerate Formation consists of polymictic conglomerates and pebbly sandstone with alterations of thin beds of coarse and fine-grained sandstones. Rock clasts are mainly derived from basic-ultrabasic sources, with minor contributions from quartz veins, shale, limestone, sandstone, chert and porcellinite (Allen et al, 2007) (Fig. 2.3b & c). The outcrop at Hope Town (type locality) has been reported to have fining- and thinning-upward sequences with evidence of slumping and soft sediment deformation (Allen et al., 2007). Bed contacts are generally sharp and planar.

(c) Namunagarh Grit Formation

This unit is characterized by coarse to fine-grained, greenish to grayish coloured matrix supported sandstones beds, consisting of volcanic minerals and volcanic rock fragments, and minor conglomerate at the base (Bandopadhyay, 2005; Allen et al.,



Fig. 2.2: Stratigraphic litholog showing different rock units exposed on the Andaman Islands. Modified after Guha and Mohan (1965); Karunakaran et al. (1968a); Ray (1982); Pal et al. (2003); Curray (2005) and based on our observations.

Table 2.1: Stratigraphy, lithology, depositional environments, tectonic settings and fossil records of the rocks of the Andaman & Nicobar Islands (based on studies by Guha and Mohan, 1965; Karunakaran et al., 1968; Ray, 1982; Pal et al., 2003; Curray, 2005)

Discene Nicobar Group (Series)/Arc Nicobar Group (series)/Arc Raised beaches, coral reefs, alluvium, swampy mud flats Nanofossils Pelag foraminifers, Lepidocyclina, Miogypsina, Late Pliocene (1.95-3.7) Early Pliocene (3.7-5 Ma) Nacobar Group (series)/Arc Shampenian Taipian Raised beaches, coral reefs, alluvium, swampy mud flats Nanofossils Pelag foraminifers, Lepidocyclina, Miogypsina, Miocene to (1.95-3.7) Early Pliocene (3.7-5 Ma) Archipelago (400 m thick) Neilian (400 m thick) Interbedded sequence of tuff, cross-stratified and graded sdst, silty mdst, lmst. Wave and current agitated shallow water shelf forearc setting, palaeocurrent direction: E. W and majority in SW (Pal et al., 2003). Nanofossils Pelag foraminifers, Lepidocyclina, Miogypsina, 1982) Miocene (16-25 Ma) Andamanian Andamanian Interbedded biostromal) Mdst., silty mdst., lmst. mats, and chalky lmst. (Lmst is biolermal & biostromal) Wave and current agitated shallow water shelf forearc setting, palaeocurrent direction: E. W and majority in SW (Pal et al., 2003). Nanofossils Pelag foraminifers, Lepidocyclina, Miogypsina, al., 2003). Under enercity Jarawaian Mdst., silty mdst., lmst. Inglisian Mdst., silty mdst., lmst. Inglisian Creamish yellow calcareous chalk & marl Grey snady lmst., white siliceous chalk and silt Andaman Flysch Bouma sequences, interbedded skst -shale Submarine fan/forearc settino) Rotalia Nodosani	Approximate depositional		Ltihostratigr	aphic Unit	Lithology		Sedimentary environments/	Fossil Record
Image: Charly plicene (3.7-5 Ma) Sawaian Mdst., silty mdst., lmst. Mdst., silty mdst., lmst. Mast., silty mdst., lmst. Nanofossils Pelag shallow water shelf forearc setting. palaeocurrent direction: Nanofossils Pelag for animifers, Lepidocyclina, Miogypsina, Miogypsina	Pliocene- Recent (Curray, 2005)	Plesitocene/ Holocene (0-1.95 Ma) Late Pliocene	Nicobar Group (series)/Arc Volcanics/recent sedimentary deposits	Shampenian Taipian		Raised beaches, coral reefs, alluvium, swampy mud flats Shell Lmst. Silty Mdst, Imst.		Nanofossils Pelagic foraminifers, Lepidocyclina, Miogypsina,
Miocene to Pliocene Late Miocene (5- (Chatterjee, 10 Ma) Archipelago Group (400 m thick) Neilian Interbedded sequence of tuff, cross-stratified and graded sdst., silty Mdst., silty mdst., lmst. Wave and current agitated shallow water shelf forearc setting. palaeocurrent direction: Nanofossils Pelag for aminifers, Lepidocyclina, 1964; Ray, 1982) Middle Havelockian graded sdst., silty Mdst., silty mdst., lmst. Wave and current agitated shallow water shelf forearc setting. palaeocurrent direction: Nanofossils Pelag for aminifers, 1982) Miocene Ongeian mdst. & lmst. Mdst., silty mdst., lmst. E, W and majority in SW (Pal et al., 2003). Miogypsina, Early Miocene Inglisian marls, and chalky Imst. (Lmst. is creamish yellow biohermal & biohermal & biotstromal) creamish yellow calcareous chalk & marl Grey sandy lmst., white siliceous chalk and silt submarine fan/forearc setting) Rotalia Nodosaria		(1.95-3.7) Early Pliocene (3.7-5 Ma)	-	Sawaian		Mdst., silty mdst., lmst.	-	
—Unconformity— Oligocene— (25-45 Ma) Andaman Flysch Bouma sequences interbedded sdst -shale Submarine fan/forearc setting) Rotalia Nodosaria	Miocene to Pliocene (Chatterjee, 1964: Ray, 1982)	Late Miocene (5- 10 Ma) Middle Miocene (10-16 Ma) Early Miocene (16-25 Ma)	Archipelago Group (400 m thick)	Neilian Havelockian Ongeian Inglisian Jarawaian Andamanian	Interbedded sequence of tuff, cross-stratified and graded sdst., silty mdst. & lmst. marls, and chalky lmst. (Lmst. is biohermal & biostromal)	Mdst., silty mdst., lmst. Mdst., silty mdst., lmst. Mdst., lmst. creamish yellow calcareous chalk & marl creamish yellow calcareous chalk & marl Grey sandy lmst., white siliceous chalk and silt	Wave and current agitated shallow water shelf forearc setting. palaeocurrent direction: E, W and majority in SW (Pal et al., 2003).	Nanofossils Pelagic foraminifers, Lepidocyclina, Miogypsina,
Oligocene- (25-45 Ma) Andaman Flysch Bouma sequences interbedded sdst -shale Submarine fan/forearc setting) Rotalia Nodosaria	Unconformity							
Late Eocene (Pawde & Ray, 1963) Group (300 m thick) Group (300 m t	Oligocene– Late Eocene (Pawde & Ray, 1963)	(25-45 Ma)	Andaman Flysch Group (300 m thick)		Bouma sequences, in rhythmites and mdst.	terbedded sdstshale	Submarine fan/forearc setting). palaeocurrent direction: NE, SW and majority in SE (Pal et al., 2003).	Rotalia, Nodosaria, Amphistegina, Globorotalia, Opimanana, Globigerinoids

Table2.1: Continued

Late Cretaceous to Middle/Upper Eocene (Curray 2005)	(45-70 Ma)	Mithakhari Group (1400 m thick)	Namunagarh Grit	Pebbly and coarse to fine-grained volcaniclastic sdst. & grits, interstratified massive and graded polymict conglomerates, massive cross-stratified and graded sdst., shales, and thin coal seams.	andesitic debris flows and turbidites deposited in forearc- volcanic are setting basin adjacent to volcanic arc (Bandhopadhyay, 2005). palaeocurrent direction: NW, a few show trend between NE-SE (Pal et al., 2003).	Cumulates atacicus, Assilina papillata,Pelatispira, Biplanispira
			Hope Town Conglomerate	Interstratified massive and graded polymictic conglomerates, cross-, massive-and graded - bedded sdst., up-section change over to interbedded shale and coal with minor conglomerates	Submarine slope, shelf and alluvial environments/deposited in faulted-controlled basins within a trench-slope setting (Chakraborty et al., 1999)	
			Lipa Black	Pyritiferous black shale with olistoliths in	Euxinic environment/Trench	1
			Shale	sheared argillite matrix	setting	
		1		—Tectonic/unconformity—		
Late Cretaceous to Paleocene (Roy et al. 1988)	(~95 Ma)	Ophiolite Group (?)		Serpentrzed harzburgite, pyroximite, unclassified plutonic rocks. Cumulates, tectonites and metamorphic. Pillow lavas, intrusive and pelagic sediments	Uphfied segments of ocean floor forming the basement of the accretionary prism	Radiolaria: Nasselaria spumellarion Planktonic foraminifera: Globigerina eugubina, G. trilloculinsides, G. fringa, G. velascoensis. Globorotal ia compressa, Globotruncana arca, Globigerinelloides Pseudotextularia sp., Reugoglobigerina sp. Hagiastrids

Mdst: mudstone; Sdst: sandstone; Lmst: limestone



Fig.2.3: Field photographs of various formations of the Mithkhari Group (a) Lipa Black Shale from a road side section in Sippighat; (b) Hopetown conglomerate unit exposed at Chidiyatapu; (c) Hopetown conglomerate unit showing clasts derived from Ophiolite Group rocks; (d) Namunagarh Grit unit from quarry section in Namunagarh village, South Andaman.

2007) (Fig. 2.3d). Sandstone units are massive and thickly bedded with plane-laminated and cross-stratified facies. The finer grained beds consist of fine- to medium-grained sandstones inter-bedded with laminated shale. Bandopadhyay (2005) identified the sandstones of Namunagarh as tuff beds, and suggested their direct derivation from volcanic arc sources. However, a similar origin for Namungarh units, exposed elsewhere on the islands, has not yet been established.

Considering the presence of coal, gypsum, cross-bedded sandstone lenses and pyritebearing shale in the Mithakhari Group, Pal et al. (2003) suggested that these rocks were possibly deposited in a low gradient alluvial plain that suffered occasional marine transgression. According to these authors the coarse sediments including polymict conglomerates and grits were directly derived from the local ophiolite sources while the fine to very fine grained sediments (turbidites) were derived from distant sources. Highly variable bedding orientation and a complex deformation pattern are shown by the rocks of Mithakhari Group. These also show sedimenatry features like clast imbrication, sole marks, wave ripples and large-scale channel bedform. The palaeocurrent direction derived from sandstone facies indicate that dominant sediment influx was from the NW to NE (Pal et al., 2003).



Fig. 2.4: Field photographs showing formations of the Andaman Flysch Group, Archipelago Group and exposed coral reefs. (a) Shale and sandstone beds at Corbyn's Cove, Port Blair; (b) Shale and sandstone beds at Lamiya Bay, Kalipur; (c) Foraminiferal limestone beds of Archipelago Group exposed at Radhanagar beach, Havelock Island; (d) Dead corals exposed on beach at Lamiya Bay, Kalipur, North Andaman.

2.1.3 Andaman Flysch Group

The Andaman Flysch unit overlying the Mithkhari Group is believed to have been deposited in a forearc environment (Pal et al., 2003). Chakraborty and Pal (2001)

interpreted that the sequence was deposited in a submarine fan and observed three different facies associations: upper fan, mid fan and basin plain facies. The upper fan facies consists of fine-grained sediments derived from ophiolites and deposited in a high-density turbidity current or channelized debris flow environment. Mid-fan facies consists of thick beds of coarse to fine-grained sandstone and shale with well-defined Bouma cycles (Fig. 2.4a & b), also described as poorly sorted matrix-supported quartz wackes with framework grains of detrital quartz, lithic fragments, feldspar and pyroxene. Basin plain facies is characterized by hemipelagic mudstone interbedded with irregular and lenticular silt laminae and thin sand beds. The total thickness of this unit is not known, but estimates vary from 750 m (Roy, 1983) to 3000 m (Pal et al., 2003). N-S and NNE-SSW striking individual beds of this group can be traced for several kilometers. Sedimentary structures like flute cast, groove cast, and current bedding have also been identified in the sandstone beds and on the basis of their orientation southward-directed paleocurrent patterns have been identified (Pal et al., 2003). The siliciclastic turbidites of the Andaman Flysch Group are less deformed and are little affected by tectonics in comparison to rocks of underlying units of the Mithakhari Group (Macdonald, 1993). The beds show regular fold patterns and have a wide lateral extent.

2.1.4 Archipelago Group

The Archipelago Group, like the Andaman Flysch Group, is also believed to have been deposited in a forearc basin environment (Pal et al., 2003). The ongoing tectonics resulted in gradual shallowing of the basin and change in depositional environment subsequent to the deposition of flysch sediments. Due to which the sediment facies changed from siliciclastic to carbonate turbidites with alternations of pyroclastics layers. The pyroclastics were apparently derived from the inner arc volcanoes and deposited as ash falls or subaqueous pyroclastic flows (Srinivasan, 1988). Previous studies have divided the Archipelago Group sediments mainly into two major litho-assemblages. The lower one contains pyroclastics deposited in subaqueous condition and siliciclastic turbidites consisting of coarse-graded, cross-laminated and parallel laminated greywackes, siltstone and mud (Pal et al., 2003). The pyroclastics and turbidites are well

developed in South and Middle Andaman Islands. The upper part consists of ash turbidites overlain by carbonate turbidites (Fig. 2.4c). This unit is well exposed in the Havelock Islands in the east. Although, synsedimentary basinal disturbances are indicated by micrograbens and slump structures, in general beds show very regular fold geometry. Environment of deposition of the sediments of the Archipelago Group is still debated. Roy (1983) suggested their deposition in a shallow marine condition, while Srinivasan (1986) interpreted it to have been deposited in deep water to neritic or outer neritic condition. Curray (2005) inferred the presence of both the facies.



Fig. 2.5: Field photographs showing (a) uplifted coral bed at Radhanagar Beach, Havelock Island; (b) Mangroove swamp deposits at Nilambur Jetty, Baratang Island; (c) Narcondam Island volcanics; (d) Volcanic deposits on northern caldera wall of the Barren Island Volcano.

2.1.5 Nicobar Group/Volcanics/Recent sediments

This unit mainly comprises of rocks and sediments exposed on the Andaman and Nicobar Islands in the form of alluvium, raised beaches, coral reefs (Fig. 2.4d; 2.5a), calcareous tuffs, beach sand, mangrove swamp deposits (Fig. 2.5b), shell limestone along with volcanic rocks and ash produced by the two subaerial volcanic islands (Fig. 2.5c & d). Coral reefs, mangrove swamps, shell limestones and beach sands are present along the coast lines of the islands with large number of bays, lagoons and serpentine creeks, while alluvium is present along the local rainfed streams, in the valleys and on the hill tops. Volcanic products derived from younger volcanoes are restricted to the volcanic islands.

2.1.6 Geochronology of rocks of the Andaman Islands

The presently accepted chronology of the ophiolite and sedimentary sequences of the Andamans is primarily based on biostratigraphy. Based on the presence of key/index fossils of radiolarians (e.g. Hagiastrids, Nassellarian) in cherts, and planktic foraminifera (Globoratalia and Discocyclina) in pelagic sediments of the ophiolite sequence, the age of the Ophiolite Group has been constrained to Late Cretaceous to Paleocene (Jafri, 1986; Roy et al., 1988; Pal et al., 2003). However, from recent geochronological information based on U-Pb dating of zircons from plagiogranite, it has been suggested that the ophiolitic rocks could have formed as early as ~95 Ma (Pederson et al., 2010 Sarma et al., 2010). Compared to the Paleogene sedimentary rocks of the Mithakhari Group, the Andaman Flysch Group rocks are largely barren in nature in terms of their fossil content (Allen et al., 2007). The reason behind this could be that either fossils were not present in these sedimentary rocks from the beginning or got dissolved/removed later by weathering processes (Allen et al., 2007). The possibility of dissolution of planktic shells below the carbonate compensation depth and abrasion by reworking cannot be ruled out. The presence of fossils of shallow benthic life forms (e.g. Nummulites atacicus, Nummulites spp., small miliolids and rare Morozovella spp., fragments of rhodophyte algae, and dasycladaceans) in the Mithakhari Group, suggests Late Palaeocene to Eocene age of deposition (Karunakaran et al., 1967, 1968a; Roy et al. 1988). Biostratigraphic evidences are vague in the Andaman Flysch Group but based on a few fossils recognized, it is believed to have been deposited between Oligocene to the early Miocene (ca. 36–21 Ma) (Pal et al., 2003). Radiolarians, planktic foraminifers, and calcareous nanofossils are present in abundance in the calcareous sedimentary rocks of Archipelago Group (Singh et al., 2000) and are inferred to have been deposited during Miocene to Pliocene (Pal et al., 2005).

Since the chronological constraints based on biostratigraphy were very imprecise, Allen et al. (2007) made an attempt to date these sequences using Fission-Track (FT) and (U/Th)-He methods on apatites and zircons and 40 Ar- 39 Ar method on detrital micas. Although these methods are known to be more suitable for understanding thermal history of the basin, Allen et al. (2007) could successfully constrain the depositional age of the Mithakhari Group to <60 Ma and that of the Andaman Flysch Group to <40 Ma.

2.2 Field studies and Sampling details

2.2.1 Field Studies and Mapping

A comprehensive study on geological mapping and stratigraphy has not yet been done for the Andamans, therefore, it is quite likely that earlier studies have missed or misidentified many lithounits. Due to poorly developed biostratigraphy and lack of isotopic ages, stratigraphic correlations on the islands are very rudimentary. With exposures limited to quarries, coastal areas, and road cuts and presence of thick forest that includes large restricted tribal reserves, sampling has been the most difficult part of this work. Most of the earlier studies have been carried out in and around Port Blair, the capital city of the Andaman and Nicobar Islands. A large part of the middle and north Andaman islands are either restricted or difficult to access, resulting in very little, if any, geological information, whereas owing to easier access, the south Andaman island is well mapped. For the current study, we have made our sampling strategy on the basis of the existing maps and sample locations given in earlier works. The samples were mainly collected in three field campaigns conducted during March-April and December



Fig. 2.6: The geological maps of the (a) entire Andaman archipelago, (b) North Andamans, (c) Middle Andamans and (d) South Andamans. Sampling locations are marked as stars.

Sample	Lat/Long	Location	Group/Formation	Description
South Andaman				
AND-09-01*	N 11°30.338', E 92°42.430'	Chidiyatapu	Mithakhari Group	Coarse sandstone
				with clasts of reef
				carbonate, ophiolite rocks
AND-09-02	N 11°30.338', E 92°42.430'	Chidiyatapu	Mithakhari Group	Gritty coarse sandstone
AND-09-03	N 11°30.168', E 92°42.029'	Chidiyatapu	Andaman Flysch	Greenish sandstone
AND-09-04*	N 11°38.058', E 92°43.180'	Dollyganj	Andaman Flysch	Dark grey
				carbonaceous shale
AND-09-05	N 11°38.058', E 92°43.180'	Dollyganj	Andaman Flysch	Dark grey carbonaceous
				siltstone
AND-09-36*	N 11°37.985', E 92°42.180'	Attampahar	Andaman Flysch	Weathered fine
				grained mudrock
AND-09-37	N 11°37.585', E 92°42.439'	Gyaracharma	Mithakhari Group	Sandstone
AND-09-39	N 11°37.585', E 92°42.439'	Gyaracharma	Mithakhari Group	Gritty coarse sandstone
AND-09-41	N 11°37.437', E 92°42.466'	Gyaracharma	Mithakhari Group	Rusty-brown shale
				with calcite veins
AND-09-42*	N 11°36.674', E 92°42.544'	Gyaracharma	Andaman Flysch	Sandstone
AND-09-43	N 11°36.674', E 92°42.544'	Gyaracharma	Andaman Flysch	Grey shale
AND-09-44	N 11°36.112', E 92°40.956'	Sippighat	Mithakhari Group	Shale
AND-09-45	N 11°36.112', E 92°40.956'	Sippighat	Mithakhari Group	Sandstone
AND-09-48	N 11°35.523', E 92°40.330'	Nayasar	Mithakhari Group	Dark grey shale
AND-09-49	N 11°35.523', E 92°40.330'	Nayasar	Mithakhari Group	Purple coloured shale
AND-09-51*	N 11°40.347', E 92°41.046'	Namunagarh	Mithakhari Group	Clasts
AND-09-52	N 11°40.347', E 92°41.046'	Namunagarh	Mithakhari Group	Shale and coal
AND-09-53	N 11°40.347', E 92°41.046'	Namunagarh	Mithakhari Group	Clasts
AND-09-54*	N 11°40.248', E 92°42.492'	Namunagarh	Mithakhari Group	Gritty coarse sandstone
AND-09-55*	N 11°40.280', E 92°41.339'	Hopetown	Mithakhari Group	Grit/conglomerate
AND-09-56	N 11°30.262', E 92°42.065'	Chidiyatapu	Mithakhari Group	Clasts
AND-09-57	N 11°29.557', E 92°42.419'	Chidiyatapu	Mithakhari Group	Clasts
AND-09-61	N 11°37.246', E 92°43.567'	Protherapur	Andaman Flysch	Sandstone & shale
PB-08-07	N 11°39.700', E 92°45.336'	South Point	Andaman Flysch	Shale
PB-08-08*	N 11°39.700', E 92°45.336'	South Point	Andaman Flysch	Sandstone
PB-08-09	N 11°35.423', E 92°37.139'	Wandoor	Andaman Flysch	Sandstone
PB-08-13*	N 11°35.849', E 92°40.916'	Sippighat	Andaman Flysch	Sandstone
AND-11-05	N 11°29.190', E 92°40.144'	Rutaland	Mithakhari Group	Gritty coarse sandstone
AND-11-20	N 11°36.473', E 92°39.924'	Chouldhari	Mithakhari Group	Weathered hard sandstone
AND-11-21	N 11°39.971', E 92°38.191'	Hobdaypur	Mithakhari Group	Sandstone
AND-11-22	N 11°41.327', E 92°36.468'	Collinpur	Mithakhari Group	Black sandstone
AND-11-23	N 11°41.394', E 92°36.309'	Collinpur	Mithakhari Group	Yellow weathered
		•	-	micaceous sandstone
AND-11-54	N 11°47.881', E 92°39.192'	Mile Tilek	Mithakhari Group	Greenish-white tuff
Middle			-	
AND-09-06	N 12°07.822', E 92°46.989'	Katang	Mithakhari Group	Greenish sandstone
AND-09-07*	N 12°05.731', E 92°44.702'	Baratang cave	Archipelago Group	Sandy limestone
AND-09-08	N 12°05.654', E 92°44.653'	Baratang cave	Archipelago Group	Sandy limestone
AND-09-09*	N 12°11.277', E 92°47.543'	Oralkatcha	Mithakhari Group	Greenish sandstone
AND-09-10	N 12°11.277', E 92°47.543'	Oralkatcha	Mithakhari Group	Greenish sandstone
AND-09-11*	N 12°11.277', E 92°47.543'	Oralkatcha	Mithakhari Group	Gritty coarse sandstone

Table 2.2: Description of samples collected from the Andaman Islands

Table 2.2: Continued

AND-09-12 N 12'31.692', E 92'49.890' Kaushalyangar Andman Flysch Sandstone Sandstone Mithakhari Group Gritty/coarse sandstone With clasts AND-09-14 N 12'31.870', E 92'58.424' Panchvati Mithakhari Group Coarse sandstone with clasts AND-09-16 N 12'34.146', E 92'57.901' Panchvati Mithakhari Group Coarse sandstone with clasts AND-09-17 N 12'34.146', E 92'57.901' Panchvati Mithakhari Group Coarse sandstone with clasts AND-09-17 N 12'34.146', E 92'57.901' Panchvati Mithakhari Group Coarse sandstone with clasts AND-09-17 N 12'34.06', E 92'52.473' Bakutala Mithakhari Group Limestone AND-11-32 N 12'30.067', E 92'52.473' Bakutala Mithakhari Group Limestone AND-11-33 N 12''30.02', E 92''52.687' Bakutala Mithakhari Group Limestone AND-11-41 N 12''32.303', E 92''57.405' Shivapuram Mithakhari Group Limestone AND-11-51 N 12''35.733', E 92''57.405' Shivapuram Mithakhari Group Coarse sandstone With calcite veins AND-09-18 N 13''14.284', E 92''57.51' Subhasgram Mithakhari Group Coarse sandstone AND-09-20 N 13''15.514', E 92''58.612' Madhupur Mithakhari Group Coarse sandstone AND-09-21 N 13''15.514', E 92''58.612' Madhupur Mithakhari Group Coarse sandstone AND-09-21 N 13''15.514', E 92''58.612' Madhupur Mithakhari Group Coarse sandstone AND-09-21 N 13''15.514', E 92''58.612' Madhupur Mithakhari Group Gritty coarse sandstone AND-09-21 N 13''15.514', E 92''58.612' Madhupur Mithakhari Group Gritty coarse sandstone AND-09-21 N 13''15.514', E 92''58.612' Madhupur Mithakhari Group Gritty coarse sandstone AND-09-23 N 13''15.14', E 92''58.612' Madhupur Mithakhari Group Gritty coarse sandstone AND-09-24 N 13'12.186', E 93''02.430' Lamiya Bay Mithakhari Group Gritty coarse sandstone AND-09-28 N 13'12.005', E 93''02.403' Lamiya Bay Mithakhari Group Gritty coarse sandstone AND-09-28 N 13'12.005', E 93''02.403' Lamiya Bay Mithakhari Group Gritty coarse sandstone AND-09-28 N 13'12.005', E 93''02.403' Lamiya Bay Mithakhari Group Gritty coarse sandstone Gritty coarse sandstone AND-09-30 N 13'16.718', E 93''02.777' Lamiya Bay Mithakhari Group Gri	Sample	Lat/Long	Location	Groun/Formation	Description
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AND-09-69 N 11°58.806', E 92°57.646' Radhanagar Beach Recent deposits Coral fragments	AND-09-69	N 11°58.806', E 92°57.646'	Radhanagar Beach	Recent deposits	Coral fragments

Samples marked with '*' were used for petrographic studies.

of 2009 and during December of 2011. Since the focus of this work was sedimentary rocks, considerable efforts were put to cover all the accessible outcrops and most of the formations/lithounits. We have divided our sampling areas into four sectors: North, Middle, South Andamans and Havelock Island within Ritchie's Archipelago (Fig. 2.6). After a thorough field survey we have prepared new geological maps for North, Middle and South Andamans and attempted to establish the interrelationship of different stratigraphic units in various sectors (Fig. 2.6).

On the Andaman Islands, the Ophiolite Group rocks are exposed mainly on the east coast of North, Middle and South Andaman Islands, where individual magmatic/sedimentary units of the ophiolite sequence occur (Fig. 2.6). There are also some small isolated patches of ophiolite exposed in the western part of North Andaman and South Andaman Islands. The continuous and best-preserved outcrops of the Ophiolite Group rocks occur in South Andaman, extending from Corbyn's Cove to Chidiyatapu and continues into Rutland Island in south. The type areas for the Mithakhari Group formations are also located in South Andaman, with a few good easily accessible outcrops of the Hope Town Conglomerate at Hopetown and Chidiyatapu, while the Namunagarh Grit is well exposed in a quarry at of Namunagarh Village. Extensive outcrops of these rocks occur in North and Middle Andaman Islands as well. In South Andaman, the lithology is mainly dominated by the Andaman Flysch Group rocks with the best exposures found at Corbyn's Cove where outcrops of steeply dipping beds are seen (Fig. 2.6). The Andaman Flysch rocks are generally found along the eastern and the western coastlines (Fig. 2.6). The Archipelago Group occurs as a thin band in the centre of the main Andaman Island in NNE-SSW direction, which is the general strike of the island chain. However, the best exposures are confined to the islands of Richie's Archipelago and Interview Island. It is believed that the Archipelago Group of rocks had formed a blanket over the Andamans earlier; however, later events of upliftment resulted in erosion of these relatively softer rocks, leaving behind small patches of outcrops on the main islands (Allen et al., 2007).

2.2.2 Sampling of sedimentary sequences

One of the main objectives of this study was to decipher the provenances for the siliciclastic formations belonging to the Mithakhari and Andaman Flysch Groups. Apart from the sediments from these groups, we also collected rock samples from the Ophiolite Group, considering that the igneous and metamorphic rocks of this group could be one of the most likely sources of sediments. The lowermost unit of the Mithakhari Group, the Lipa Black shale is not well exposed and hence could not be sampled. Samples of conglomeratic and gritty sandstones were mainly collected from Chidiyatapu, Sippighat, Namunagarh, Hopetown and Hobadaypur in South Andaman, Baratang Island in Middle Andaman and from area around Diglipur in North Andaman, where these units are well exposed. In the main island of Middle Andaman Flysch Group samples were mostly collected from various locations on South Andaman Island. List of samples with their locations and other relevant information are given in Table 2.2 and are marked in Fig. 2.6.



Fig. 2.7: Map of the northeastern Indian Ocean showing the locations of Andaman Islands, Barren Island and the studied core SK-234-60 (N12°05′46", E94°05′18") in the Andaman Sea.

2.2.3 Sampling of Coral terraces

To study the timing of upliftment of Andaman coastline and its causes we collected samples from exposed dead coral from coastal terraces of some islands of Andamans. Our study areas include Lamiya Bay (Kalipur) on the eastern coastline of North Andaman and Radhanagar Beach on the west coast of Havelock Island (Fig. 2.6). Three coral samples were collected from terraces of Lamiya Bay and two from Radhanagar. These terraces represent recent sedimentary deposits belonging to topmost unit of our stratigraphic divison. At Lamiya Bay three step-like terraces were recognized, while at Radhanagar Beach section, two terraces were recognised. At Radhanagar, like that in Lamiya Bay dead corals occur at sea level and on a raised terrace at a height of ~3.5m. On this terrace, the dead coral reef is present above the foraminiferal limestones of the Archipelago Group.

2.2.4 Sampling of modern sediments from the Andaman Sea

To decipher the provenance of the modern sediments deposited in the Andaman Sea, a sediment core (SK-234-60) was raised from a location 32 km (great circle distance) southeast of the Barren Island Volcano (Fig. 2.7). This 4-m-long and 12-cm-diameter core was collected during the expedition no. 234 of R/V 'Sagar Kanya' on June 02, 2007 at station no. 60 (N12°05′46", E94°05′18"; Fig. 2.7) at a water depth of 2000 m using a cylindrical gravity corer. The core was cut along with the casing into four ~1m sections on the board before being shipped and back to our laboratory for further study. In the laboratory, each ~1m section of the core was split into two halves; one half of such a section is shown in Fig 2.8(a). One of the halves was sub-sampled at 5 cm intervals and the other half at 2 cm intervals for geochemistry and geochronology, respectively. We did not observe any deformation on the edges of the sediment layers in the core except for the top 20 cm, which was disturbed during coring (Fig. 2.8b).

Bathymetry of the region suggests that the site of the core was far from the submarine base of the Barren Island Volcano (Fig. 2.7), and thus well shielded from direct slides or slumps on the flanks of the volcano. Based on colour and grain size of sediments, seven discrete ash layers were identified within the olive-grey to very dark grey pelagic sediments (Fig. 2.7b, c & d), which were later confirmed by their



Fig. 2.8: (a) Photograph of one half of a 1m section of the core; (b) Litholog of the same core, 4m in length, showing volcanic ash layers (AL-1 to 7) in normal ocean sediments. The top 20cm is disturbed; (c) photograph of pure volcanic ash separated from one of the layers; (d) photograph of normal sediment.

contents of glass and primary igneous minerals through microscopy and X-ray diffractometry. These ash layers occur at intervals of 53-57, 73-75, 90-94, 109-112, 302-321, 372-377 and 380-387 cm and are labelled AL1 to AL7 from the top to the bottom of the core (Fig. 2.8b). A detailed description of these layers is given in Chapter-3.

2.3 Analytical Methods

To achieve the objectives set for the work, various methodologies were used to generate analytical data for chronology, mineralogy and geochemistry. The chemical

procedures for separation of Sr and Nd from samples and associated mass spectrometric analyses on Thermal Ionization Mass Spectrometer (TIMS) are mentioned only briefly in this section as they were based on already established routine procedures of our laboratory. Appropriate care was taken during the analyses to avoid contamination. For geochemical and isotopic analyses well homogenized powdered samples were used.

The sediment core from the Andaman Sea contained silt-clay sized siliciclastics, foraminiferal carbonates and coarser ash material consisting of lithic fragments, minerals and glass shards. We separated coarse ash from the rest by gravity separation. ~250g of sediments were taken in a clean glass beaker and ultrasonicated repeatedly in order to disaggregate ash from the mud. The less dense slurry was then decanted off and collected in a separate beaker and dried down for further analysis. The ash and mud fractions were individually treated with 20 % acetic acid and washed several times to remove carbonates. These were then dried at ~70°C and powdered using an agate mortar. Rock samples were wrapped in polyethylene bags and crushed into smaller pieces. Selected chips were cleaned in water and alcohol using an ultrasonicator to remove finer impurities. Dried chips were then powdered by an agate mortar.

2.3.1 Petrography

Thin section studies of the Andaman sedimentary rocks were done to decipher mineralogy, identify fossils present and determine degree of secondary alteration. Grains from the coarse grained layers in the core were studied under microscope to identify magmatic minerals and lithic fragments. X-ray diffraction of some of these layers was also done to confirm the mineralogy and hence the volcanic nature of the coarse grained layers.

2.3.2 Major element analysis

(A) X-ray Fluorescence (XRF) Spectrometry

Analyses of major element concentrations in rocks and sediments were done by XRF spectrometry. Measurements were done in a Philips AXIOS X-Ray Spectrometer fitted with an Rh X-Ray tube, operated at 50 kV and 55 mA, of 4 kW

power. The instrument belongs to the National facility for Planetary and Exploration Program (PLANEX) of Indian Space Research Organization (ISRO) and located at PRL. The instrument was set-up, installed and calibrated for routine measurement of rock samples during this research activity (Ray et al., 2008). Several international standards such as AGV-1, G-2, GSP-1, JLS-1, JMS-2, JSo-1, MAG-1, NOVA-13, NUSSI, SCo-1, SDO-1 and SOIL-5 supplied by United State Geological Survey (USGS) and Japanese Geological Survey were used for calibration purpose during our measurements. Oxides of major elements: SiO₂, TiO₂, Al₂O₃, Fe₂O₃, MgO, CaO, K₂O, Na₂O, MnO, and P₂O₅ were analysed on pressed pellets of finely powdered samples, which were already demoisturized at 110 °C. Sample pellets were prepared using 2.0 gm of sample powder mixed with 0.5 gm wax binder. The mixtures were homogenized in an agate mortar and taken in 37 mm aluminum cups with more wax at the bottom. The cups were pressed in a hydraulic press at 150 kN pressure for a minute to make the pellets. Figure 2.9 gives the typical XRF calibration curves for various major element oxides obtained during our analyses. With each set of samples an international rock standard was also analyzed as unknown for accuracy check. The measured values are in good agreement with the reported values within 2σ error (Table 2.3). The precision of measurements based on repeated analyses of standards is better than 3% at 2σ level except for Fe, Na and Mn, for which it is better than 10% at 2σ .

(B) Major element analysis by Electron Probe Micro-Analyzer (EPMA)

For major element analysis of ash in core, bulk analysis was not possible because of limited sample amount. Grains of minerals and lithic fragments were picked up from the ash under a stereomicroscope and EPMA mounds were prepared. The mounds were polished on an automated polish wheel and coated with carbon. Major element concentrations were measured on 42 samples using a Cameca SX100 EPMA, at PLANEX facility of PRL, in wavelength-dispersive mode. Analyses were performed with an electron beam current of 10 nA and accelerating voltage of 15 kV. A glass standard (USNM 113716, Jarosewich et al., 1980) was periodically analysed to check both the accuracy and precision. Table 2.4 presents the recommended and measured data for this standard. Analytical data of individual glass analyses were normalized to 100 weight percent.

	Measured		Reported*	
	(n=15)	±2 σ		±2 σ
(%)				
Si O ₂	49.29	0.75	49.28	1.26
TiO ₂	0.71	0.02	0.71	0.06
AI_2O_3	12.46	0.28	12.27	0.46
Fe ₂ O ₃	9.65	0.69	9.34	0.42
CaO	1.04	0.04	1.05	0.05
MgO	1.52	0.08	1.54	0.08
MnO	0.05	0.01	0.04	0.01
K₂O	3.24	0.10	3.35	0.12
Na ₂ O	0.43	0.06	0.38	0.05
P_2O_5	0.11	0.01	0.11	0.01

Table 2.3: A comparison of measured and reported concentrations of various element oxides in the international standard SDO-1 (Devonian Ohio shale)

*Reported data are from USGS

	Measured		Reported
	(n=7)	±2 σ	•
(%)			
SiO ₂	50.67	0.54	51.52*
TiO ₂	1.34	0.06	1.30*
AI_2O_3	14.83	0.23	15.39*
FeO	9.52	1.07	9.24*
CaO	11.24	0.22	11.31*
MgO	7.91	0.14	8.21
MnO	0.16	0.10	0.17*
K ₂ O	0.08	0.02	0.09*
Na ₂ O	2.70	0.19	2.49*
P_2O_5	0.26	0.05	0.13*
Cr_2O_3	0.02	0.04	0.06#
CoO	0.03	0.07	-
NiO	0.03	0.07	0.02 [#]

Table 2.4: A comparison of measured and reported concentrations of various element oxides in the glass standard USNM 113716 (Basalt glass)

Reported data marked with '*' are taken from Jarosewich et al. (1980) and marked with ^{(#,} are taken from Thompson, (1980).

2.3.3 Analysis of Trace elements

(A) Inductively Coupled Plasma Mass Spectrometry (ICP-MS)

Concentrations of trace elements including fourteen Rare Earth Elements (REEs) in our samples were measured using a Thermoelectron X-Series^{II} ICPMS at PLANEX facility of PRL. The instrument was set-up, installed and calibrated for routine measurement of rock samples during this research activity (Ray et al., 2008). A



Fig: 2.9: Typical calibration curves for various major element oxides generated on XRF using multiple international rock standards.



Fig. 2.10: Typical calibration curves for different trace elements generated on ICP-MS using various dilutions of BHVO-2.

known weight of finely powdered and carbonate free sediment was taken in quartz crucibles and combusted at ~600 °C to oxidize and remove organic matter. About 50 mg of organic carbon free sample was taken in a Savillex Teflon vial and digested using a combination of ultra pure HF and HNO₃ (2:1) acid mixture (Acids were from Seastar Chemicals[®]). A stock sample solution (~50 ml) was prepared in 2% HNO₃ with ~1000 dilution factor.

To check the accuracy and precision of analyses, several aliquots of BHVO-2, an international rock standard from USGS, were digested and analysed as unknowns. Calibration curves were generated using blank solutions and various dilutions of BHVO-2 standard. Calibration curves for some elements are given in Fig. 2.10. Reproducibility based on the repeated analyses of standards was $\leq 5\%$ at 2σ level, for REEs and $\leq 10\%$ for all other trace elements. In Table 2.5 we report our data for BHVO-2 along with the recommended values.

2.3.4 Analysis of Radiogenic Isotopic Ratios

Sr and Nd isotopic ratios (⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd) were measured on selected samples from siliciclastic rocks of the Andaman Islands, decarbonated sediments and ash layers of the Andaman Sea core. These analyses were carried out on powdered samples, using the standard HF-HNO₃-HCl dissolution procedure for silicate rocks. Sr separation was done by conventional cation exchange column chemistry (Resin AG 50WX8, 200-400 mesh size), and Nd was separated from other rare earth elements using Ln-specific resin from Eichrom(R) (50-100 µm grain size) with dilute HCl (0.18N) as elutant. Measurements were carried out in static multicollection mode on an ISOPROBE-T thermal ionization mass spectrometer (TIMS) at the Physical Research Laboratory (PRL), Ahmedabad. Sr samples were loaded with 0.1 M phosphoric acid on pre-degassed, oxidized single Ta filaments while Nd samples were loaded on the outer Ta filaments of triple (Ta-Re-Ta) filament arrangements. Some Nd isotope analyses were also performed on Multicollector-Inductively Coupled Plasma Mass Spectrometer at PRL.

Sr and Nd isotope ratios were corrected for fractionation using 86 Sr/ 88 Sr = 0.1194 and 146 Nd/ 144 Nd = 0.7219, respectively. The average values for NBS987 and JNdi analyzed over a period of 4 years on TIMS are 87 Sr/ 86 Sr = 0.71023 ± 0.00001 and 143 Nd/ 144 Nd = 0.512104 ± 0.000004 (±0.1 in ε_{Nd} units) at the 2 σ level of uncertainty. The value of 143 Nd/ 144 Nd = 0.512104 for JNdi corresponds to a value of 0.511847 for the widely used La Jolla Nd standard (Tanaka et al., 2000). For comparison with literature data, the 87 Sr/ 86 Sr data were normalized to a value of 0.71025 for NBS987 and the 143 Nd/ 144 Nd data were normalized to a value of 0.511858 for La Jolla. All plots and discussion in this work are based on the normalized ratios. Also, USGS standard BHVO-2, processed regularly with each set of samples, was analysed for 87 Sr/ 86 Sr is 0.70345 ± 0.00004 (2 σ) (n=7) and for 143 Nd/ 144 Nd is 0.512949 ± 0.000080 (2 σ) (n=13), which are well in agreement with the reported values of 0.70344 and 0.512957, respectively (Raczek et al., 2003).

Elements	Measured concentration		Reported concentration*	
	(ppm)	±2 σ	(ppm)	±2 σ
Sc	29.00	2.98	31.00	2.00
V	304.71	30.75	329.00	18.00
Cr	263.38	27.36	285.00	28.00
Со	44.28	4.44	47.00	4.00
Ni	103.32	10.21	112.00	18.00
Rb	10.17	0.48	10.10	1.20
Cs	0.106	0.009	0.110	0.040
Sr	359.55	28.05	382.00	20.00
Y	21.69	1.74	23.00	2.00
Zr	149.81	12.42	160.00	16.00
Nb	16.33	0.91	16.40	1.40
Ва	113.46	25.31	128.00	8.00
La	15.37	0.83	15.25	0.04
Ce	38.03	2.12	37.84	0.38
Pr	5.45	0.29	5.35	0.03
Nd	24.95	1.35	24.39	0.04
Sm	6.21	0.34	6.03	0.02
Eu	2.08	0.12	2.04	0.01
Gd	6.39	0.36	6.23	0.01
Tb	0.89	0.05	0.86	0.12
Dy	5.45	0.29	5.30	0.02
Но	0.94	0.05	0.91	0.12
Er	2.62	0.14	2.55	0.01
Tm	0.31	0.02	0.30	0.10
Yb	2.01	0.11	1.96	0.01
Lu	0.28	0.01	0.27	0.00
Hf	4.22	0.24	4.10	0.80
Та	0.93	0.04	0.94	0.14
Pb	1.62	0.57	1.40	0.40
Th	1.21	0.06	1.18	0.18
U	0.45	0.02	0.44	0.06

Table 2.5: Measured Trace element concentrations in BHVO-2 compared with reported values

*Gao et al., (2002); Raczek et al., (2003); Kent et al., (2004).

2.3.5 Dating of the sediment core and dead corals

(A) Conventional C-14 Dating

Radiocarbon dating was done on coral samples from the coastal terraces of the Andamans and bulk sediments of the core (SK-234-60), to determine their ages of deposition, by the conventional β -counting method. For radiocarbon dating in the core, samples were selected in such a way that their depositional ages could be utilized to estimate the ages of the ash layers, and the rates of sedimentation of the

layers in- between. Since this method required ≥ 1 g of carbon for high precision, a large amount of sediment was needed (assuming an average inorganic carbon content of 1 wt%) and therefore, we sampled the core in 5-cm intervals, and 5 such layers were dated.

Radiocarbon dating of the carbonate fraction in the bulk sediments was carried out by liquid scintillation spectrometry method at the PRL following procedures described in Yadava and Ramesh (1999). For coral samples, ~10 g of powdered sample, selected from the least altered portion close to the top surface of a fragment and for marine core ~100 g of bulk sediment, was taken in an evacuated flask and reacted with orthophosphoric acid. The resulting carbon dioxide was converted to benzene and its C-14 activity was measured using a liquid scintillation counter (Quantulus 1220). An aliquot of each sample in CO₂ phase was analyzed for δ^{13} C on a stable isotope ratio mass spectrometer (GEO 20-20) and used for fractionation correction. The average δ^{13} C value used for fractionation correction was 0±2‰, with respect to V-PDB.

(B) C-14 dating by Accelerator Mass Spectrometer (AMS)

For more precise dating of the sediment core we employed AMS C-14 dating using planktic foraminifera separated from various layers (2 cm thick). ~15mg (200 individuals) of planktic foraminifers were handpicked from13 different layers down the core. AMS C-14 dating was carried out at the NSF Arizona AMS Facility, University of Arizona, USA (Linick et al., 1986; Jull et al., 1989; Somayajulu et al., 1999). The following planktic foraminifera were utilized for this study-*Globigerinoides ruber, Globigerinoides sacculifer, Orbulina universa and Neogloboquadrina dutertrei.* For both the dates (conventional and AMS) obtained on bulk carbonate and on foraminifers, ages <26000 years were calibrated using radiocarbon calibration program CALIB 6.0 (Hughen et al., 2004), considering reservoir correction (Δ R) value of 11 ± 35 yrs for the Andaman Sea (Dutta et al., 2001). Ages >26000 years were calibrated using "Marine09" and "IntCal09" radiocarbon age calibration curves (Reimer et al., 2009; Stuiver and Reimer, 1993). Our results are reported with errors at one standard deviation (1 σ).

Chapter-3 Study of volcanic tephra deposits in the Andaman region

3.1 Introduction

Cenozoic volcanism has been described throughout the Andaman region including Myanmar in the north to Sumatra-Java in the south (Chhibber, 1934) (Fig. 3.1a). From Myanmar mafic to intermediate type calcalkaline volcanism has been reported (Chhibber, 1934, Mitchell, 1985) while in Sumatra, Java and Sundaland felsic volcanism is predominant (Wakita, 2000; Wakita et al. 1998; Hall 2002, 2009; Van Bemmelen, 1949). This volcanism is related to subduction of the Indian plate under the Eurasian plate and represents the volcanic/magmatic arc since Cretaceous (Curray and Moore, 1974). Pyroclastics and ash derived from these volcanoes have been deposited along with the terrigenous sediments in trench-slope, forearc and back arc basins and many such deposits are currently exposed in the Andaman-Nicobar Islands. Identification and study of these ash beds are likely to reveal a great deal about the volcanism in the region since Cretaceous.

Ash beds found in sediments from the Indian Ocean surrounding continents point to a widespread volcanic activity in the region in the recent past (e.g. Ninkovich et al. 1978; Rose and Chesner 1987; Dehn et al. 1991; Westgate et al, 1998; Pattan et al, 1999; Schulz et al. 1998). However, chronology and geochemistry of these ash deposits suggest that most of them are linked to the last three major eruptions of Toba in Sumatra those happened during the Quaternary (e.g. Pattan et al. 2001 and references therein; Rose and Chesner 1987). In the Andaman region, there have been several reports of ash (pyroclastic) deposits interbedded with forearc sediments of the south Andaman Island, Rutland Island and Havelock Island and are approximately of Eocene through Pliocene in age (e.g., Bandopadhyay, 2005; Pal et al., 2003, 2005, 2010a; Pawde and Ray 1963; Srinivasan 1988; Pal et al., 2002). The source(s) of these ash deposits are not yet known. Andaman Sea, which came into existence during late Miocene to early Pliocene (Khan and Chakraborty, 2005), is also an important site for investigation of records of past volcanism in the region because of its proximity to the volcanic arc of Banda-Sunda and Andaman subduction zones (Fig. 3.1a). In addition, since the Andaman Sea is a semi-enclosed basin (Kurian et al., 2008), with restricted submarine transport and dispersal, the preservation of volcanic record in the sediments deposited in the basin is likely to be complete, therefore, its marine sedimentary record would be the best proxy of regional volcanism since the opening of the basin in Tertiary.



Fig. 3.1: Map of Southeast Asia showing major tectonic features and eruptive centres including the Barren Island Volcano (N12°16'40", E93°51'30") and other volcanoes of the Indonesian Arc. The numbered volcanoes are (1) Narcondam (2) Toba, (3) Sorimerapi, (4) Krakatoa, (5) Galunggung, (6) Merapi, (7) Rinjani, (8) Tambora, (9)Sangeang Api and (10) Iya Flores. The enlarged portion shows the locations of the deep-sea cores in red and yellow stars, respectively. Core SK-234-60 is from this work while cores MD-77-169 (Colin et al., 1999) and SK 168/GC-1(Sijinkumar et al, 2010) are other cores studied from nearby locations. The figure also shows Alcock, Sewell and other volcanic seamounts (in triangles) of the Andaman Sea.

3.2 Records of volcanism from the Andaman Islands

The oldest record of volcanism in the Andaman region comes from the Eocene age Namunagarh Grit Formation (Fig. 3.2a), which contains pyroclastics: abundant glass shards, lithic fragments, plagioclase crystals and altered glassy matrix (Bandopadhyay, 2005). On the basis of sedimentary, petrographic and geochemical attributes, these materials are believed to have been derived from explosive arc volcanoes (Bandopadhyay, 2005). Bandopadhyay (2005) inferred that this deposit was produced as a result of shallow subaqueous or subaerial phreatomagmatic eruption and rapid deposition in deep waters in a forearc setting, as debris flows and turbidites. Several thick sequences of felsic ash (tuff) beds have also been described from the Mio-Pliocene calcareous and non-calcareous sediments of the Archipelago Group of Rutland Island (Pal et al., 2010a), from Hubdaypur and Mile Tilek area of the south Andaman and from Havelock Island of the Ritchie's Archipelago (Pal et al., 2005) (Fig. 3.1b). From these localities, greenish to white and pinkish white facies of the ash beds (tuff) have been reported. These deposits were believed to have been generated by subaerial eruptions, however, evidences suggest that these were transported in subaqueous condition and deposited by high concentration turbidity currents (Pal et al., 2005, 2010a).



Fig. 3.2: Field photographs showing (a) pyroclastic sandstones of Namunagarh Grit Formation at Namunagarh (b) Felsic Tuff in the Archipelago Group at Mile Tilek.

Several hypotheses have been proposed by earlier workers to explain the source(s) of these ash beds. As the occurrences of these ash beds were predominantly restricted to the eastern part of the main islands, Bandopadhyay, (2005) speculated that the inner arc volcanoes located on the western margin of the Burma-Thai-Malaya Peninsula, which were predominantly felsic in nature, were the sources of these deposits. Ocean-continent type subduction must be occurring at that time to produce felsic magmas, however, at present there exists no evidence for such an arc on the continental plate east of the Andaman Sea. The inner volcanic arc located east of the Andaman Islands seems to be quite young and probably came into existence only after the opening of the Andaman Sea in the Late Miocene. Pal et al. (2010a) suggested that the early to middle Miocene felsic volcanic activity of Sumatra was a major source of dacitic and trachytic type ash beds found in the Andamans and that these ashes were transported to long distances in submarine conditions. Hall (2002, 2009) have also reported felsic volcanism of middle to late Miocene age from the Sundaland shelf. It is not clear as to why Myanmar volcanic sources have been neglected by earlier studies, however, inspite of the facts that the sedimentary formations of Andaman Islands received much of its sediments from the north (Pal et al., 2003). Locating the exact sources of volcanogenic sediments would require more clues from geochemical and isotopic studies and a clear understanding of the paleogeographic configuration of the various tectonic blocks in this region.

3.3 Records of volcanism from the Andaman Sea

Despite the fact that the Andaman Sea was created in late Miocene and that it lies east and north of the Andaman-Indonesia volcanic arc, not many workers have reported ash layers in sediment cores. In one such study Colin et al. (1999) reported an ash layer in a sediment core (MD77-169) from the Sewell Seamount (Fig. 3.1b), and suspected it to represent the youngest eruption of the famous Toba volcano of Indonesia. The N-S trending inner arc of the Andaman subduction zone is located on the western flank of the Andaman Sea and it contains two subaerial volcanoes, the active Barren Island Volcano and the dormant Narcondam (Fig. 3.1b). There exist several other prominent underwater mountains possibly of volcanic origin (Fig. 3.1b) (Curray, 2005, Kamesh Raju et al., 2012). Barren Island is the northernmost active center of the volcanic arc and acts as a link between the active volcanic arc of
Java-Sumatra and the extinct or dormant volcanoes of Myanmar (e.g., Mt. Popa, fissure vents of Singu Plateau) (Stephenson and Marshall, 1984) (Fig. 3.1a). Narcondam, located 135 km north-northeast of the Barren Island Volcano, is another volcano, believed to have come into existence during Late Pliocene to Pleistocene (Streck et al., 2011). The composition of lava and ash from the Barren Island Volcano is basaltic to basaltic andesite, whereas of Narcondam varies from basaltic to rhyolitic (Sheth et al., 2009a; Pal et al., 2007b).

Most part of these volcanoes have evolved below sea level, which suggests that their eruptive history is older than the timing of their emergence above the sea surface. The subaerially exposed parts of these volcanoes consist of volcaniclastic deposits and lava flows of unknown age. Inaccessible terrain, thick forest, lack of old exposures and unavailability of suitable samples for dating and geochemical studies have restricted extensive studies to be carried out on these volcanoes. Initial attempts to date the lavas from Barren Island and Narcondam have had limited success (Banerjee, 2010; Streck et al., 2011). Dates from Narcondam, although have very large errors, suggest that the activity of the volcano goes back in time to at least ~700 ka and the youngest eruption occurred sometime during the Holocene (Streck et al., 2011). Dating of these volcanic features and their eruptive products is essential in understanding the evolution of the inner arc in the Andaman region.

Deep marine sediment record have often been used in establishing history of large scale volcanism in a region by studying ash layers preserved in them (e.g., Carel et al, 2011; Fretzdorff and Smellie, 2002). Identifying ash beds/layers preserved in ocean basins and linking them to nearby or distant volcanoes through isotope/geochemical fingerprinting can reveal a great deal of information about the volcanic history of the surrounding region and in a few cases about mega volcanic events of global importance. Detailed mineralogical and geochemical studies can also help in understanding the nature, source, eruptive and depositional history of these volcanogenic sequences both in time and space. In an effort to build a history of volcanism in the Andaman region since the opening of the Andaman Sea, we have studied ash layers from a marine sediment core (SK-234, Fig. 3.1b). The results of this study are discussed in following paragraphs.

3.4 Study of ash layers in a sediment core from the Andaman Sea

3.4.1 Description of ash layers and geochronology

The four meter gravity sediment core (SK-234) collected from location N12°16'40", E93°51'30" (Fig. 3.1b) predominantly contained fine-grained terrigenous sediments and pelagic carbonates. Seven distinct ash layers were identified on the basis of darker color, coarser grain size, and presence of glass shards and magmatic minerals such as pyroxene and plagioclase (confirmed by microscopy and x-ray diffractometry). These ash layers were found to occur below the sea floor at intervals of 53–57 (AL-1), 73–75 (AL-2), 90–94 (AL-3), 109–112 (AL-4), 302–321 (AL-5), 372–377 (AL-6) and 380–387 (AL-7) cm (Fig. 3.3). The grain size of these layers was more than 62 μ m and the 302–321 cm layer was the coarsest of all. Apart from the distinct ash layers we also observed dispersed (< 10 % by volume) glass shards and grains of volcanic fragments in the marine sediments (clay/sand/carbonates) at many places in the core. These grains may represent minor ash eruptions of the volcano or reworked ash from any subaerial exposure.

In the core all distinct ash layers were separated from terrigenous sediments and pelagic carbonates by decarbonation and gravity separation, detailed procedure of which is discussed in Chapter-2. For source identification, Sr-Nd isotopic ratio measurements were carried out on aggregate samples of unaltered glass shards, lithic fragments and crystal grains, of each ash layer of the core. These measurements were also carried on the samples of ten ash beds from subaerial exposures of nearby Barren Island Volcano (Fig. 3.1b), and a couple of lava flows from the Narcondam Volcano in order to compare them with those of the ash layers in the core. To determine the chronology of these events planktic foraminifers from sediment layers were picked for AMS radiocarbon dating (Fig. 3.3). Major element contents of glass and lithic fragments were determined by an EPMA to characterize the eruptions and understand the magmatic evolution of the source volcano(es).

The samples for dating were selected in such a way that their depositional ages could be utilized to estimate the ages of the ash layers. The AMS ¹⁴C ages obtained for the dated sediment layers in the core are given in Fig. 3.3 with 1σ errors.

Progressively increasing ages with depth, of sediments except for the top 20 cm, confirm the core's undisturbed nature. Using these ages we determined the rates of sedimentation between the nine dated bands (Fig. 3.4). All calculations were based on the assumption that the ages represented the mean depths of the sampled intervals (or bands). In the calculations, each ash layer, irrespective of its thickness was assumed to represent an instantaneous event.



Fig. 3.3: Litholog of core SK-234-60 showing ash layers (AL-1 to 7) with calibrated AMS ^{14}C ages (in yr BP) for selected sediment layers and estimated ages of the ash layers on the left.

Assuming that the rates of sedimentation had also remained constant within the dated intervals, the ages of the ash layers were estimated (Fig. 3.4). Due to the limitation of the radiocarbon technique the age of ash layers AL-5 through 7 were not determined directly but extrapolated using the sedimentation rate of 5.5 cm/kyr assuming a constant rate of sedimentation below the bottommost dated band (Fig. 3.4). The estimated ages of the ash layers are given in Figs. 3.3 and 3.4. According to the age model, the 4m long-core covers the last ~74 ka. Based on individual rates of sedimentation the average rate for our core is estimated to about 5.3 cm/kyr.

Core depth (cm)	Tephra code	Conventional radiocarbon age ⁽¹⁾ (cal. kyr BP)	Foraminifer Species	AMS age ⁽²⁾ (cal. kyr BP)	Characteristics of fragments ⁽³⁾
53	Ash Layer-1 (AL-1)	10.1 ± 0.1	Globigerinoides ruber, Globigerinoides sacculifer, Globigerinoides quadilobatus, Globigerinoides bulloides, Globigerinoides triloba, Orbulina universa	7.9 ± 0.1	basaltic andesite to andesite- brown, black vesiculated fragments with transluscent feldspar, pyroxene, green olivine, and spinel
73	Ash Layer-2 (AL-2)	14.7 ± 0.2	do	12.3 ± 0.1	basaltic to andesite- brown, black highly vesiculated fragments with transluscent feldspar and green olivine
90	Ash Layer-3 (AL-3)	18.9 ± 0.3	do	17.3 ± 0.2	basaltic andesite to andesite- brown, black highly vesiculated fragments with transluscent feldspar, pyroxene, and green olivine
109	Ash Layer-4 (AL-4)	23.5 ± 0.3	do	23.5 ± 0.2	basaltic andesite to andesite- grey to black highly vesiculated fragments with transluscent feldspar, green olivine and titaniferrous magnetite
302	Ash Layer-5 (AL-5)	61 ± 5 (extrapolated)	-	61.9 ± 2.9 (extrapolated)	andesitic- grey to black poorly vesiculated fragments with transluscent feldspar, pyroxene, green olivine, amphibole and titaniferrous magnetite
372	Ash Layer-6 (AL-6)	69 ± 7 (extrapolated)	-	71.1 ± 4.1 (extrapolated)	basaltic andesite- brown, black highly vesiculated fragments with transluscent feldspar, pyroxene, and green olivine
380	Ash Layer-7 (AL-7)	70 ± 7 (extrapolated)	-	71.7 ± 4.1 (extrapolated)	basaltic andesite- brown, black highly vesiculated fragments with transluscent feldspar pyroxene, and green olivine

Table 3.1: AMS ¹⁴C ages of planktic foraminifera from core SK-234-60

Superscript (1): Conventional radiocarbon ages were determined on bulk carbonates using liquid scintillation counter and calibrated with the CALIB 6.0 software (Reimer et al., 2009), considering

reservoir correction (ΔR) value of 11 ± 35 yrs for the Andaman Sea (Dutta et al. 2001). The last three ash layers were dated by linear extrapolation.

Superscript (2): AMS ¹⁴C dating done at monospecific planktic foraminifer species and calibrated with the CALIB 6.0 software (Reimer et al., 2009). The last three ash layers were dated by linear extrapolation. Superscript (3): Petrological characteristics of lithic fragments were determined by EPMA.



Fig. 3.4: Depth versus age plot for the studied core. Red diamonds represent dated sediment intervals and yellow triangles are ash layers (AL-1 to 7). The numbers on the lines represent sedimentation rates (cm/kyr). Ash layer thicknesses have been removed in sedimentation rate calculation. The dashed line connects the bottommost dated sediment interval to the ash layers 5–7, assuming a constant sedimentation rate of 5.5 cm/kyr- derived from the overlying dated intervals. The straight solid line represents time-averaged sedimentation rate from a nearby core SK-168/GC-1 (Sijinkumar et al., 2010).

Petrographic studies of grains from these ash layers revealed that they were mostly vesiculated lithic fragments that contained microcrystal of translucent plagioclase, black pyroxene, and green olivine (Fig. 3.5, Table 3.1). Chemistry of lithic fragments also confirmed the presence of plagioclase, pyroxene, olivine, titaniferrous magnetite and amphibole embedded in matrix of glass (see Table 3.3). Most of the feldspar grains in lithic fragments show labradorite (An50-70%) and bytownite (An70-85%) compositions (Fig. 3.6a). The grains of AL-6 fall dominantly in the field of bytownite. The pyroxene grains in lithic fragments are mainly augite and (clino) enstatite (Fig. 3.6b). AL-5 contains a few grains of diopside too. Olivines

in the ash layers AL-1, 4, 6 show uniform composition with Fo content between 75 and 90%. Hornblende, spinel and titaniferrous magnetite are found in AL-5, AL-1 and AL-4 & 5, respectively. Spinels of AL-1 have 22-25 wt. % of Cr_2O_3 and while Fe–Ti oxides in AL-4 & 5 have TiO₂ contents up to 9–10 wt. %.



Fig. 3.5: Backscattered X-ray images of lithic fragments and mineral grains from the ash layers of the studied core. Most of the lithic fragments show large and scattered vesicles with thin walls.

3.4.2 Isotope fingerprinting and Origin

To determine the source(s) of these ash layers, we compared Sr and Nd isotopic compositions of core ash layers (AL-1 to 7) with published data for major volcanoes of the Indonesian Arc, Barren Island and Narcondam (Fig. 3.7). The ⁸⁷Sr/⁸⁶Sr and ε_{Nd} values of the ash layers in the core vary from 0.70395 to 0.70414, and 6.7 to 4.9, respectively (Table 3.2). These data plot well within the field of lava flows and ash of the Barren Island Volcano (Fig. 3.7), whereas data for ash and lava derived from all other volcanoes of the Indonesian Arc and Narcondam, plot well outside this field. The apparent differences between the isotopic ratios of ash from the core (and

lavas of Barren Island) and those of other volcanoes in the Indonesian Arc, despite their common tectonic settings, are related to the differences in the specific mantle sources and crustal architecture and lithology under each volcano.



Fig. 3.6: (a) Classification of feldspars found in the ash layers of the core; (b) classification of pyroxene crystals from the core.

These observations clearly establish that the Barren Island Volcano, located 32 km northwest of our core site, as the unique source for the ash layers. Although some values from Rinjani and Galunggung volcanoes overlap with the range observed in our data, we believe that these volcanoes are too far away to have contributed significantly to the core site. None of the ash layers in our core can be correlated with the prominent and widespread ash from the ~74 ka Toba (Sumatra) supereruption (e.g., Pattan et al., 2001 and references therein) and petrographically more evolved (and more highly explosive?) eruptions from other volcanoes of the Indonesian arc.

Sample	Age of	⁸⁷ Sr/ ⁸⁶ Sr ^a	¹⁴³ Nd/ ¹⁴⁴ Nd ^a	ε _{Nd}
	Eruption			
Ash layers in the core	(Estimated - ka) ^c			
AL1 (53-57 cm)	8.0±0.1	0.704	0.512982	6.7
AL2 (73-75 cm)	12.3±0.1	0.70409	0.512891	4.9
AL3 (90-94 cm)	17.3±0.2	0.70409	0.512943	5.9
AL4 (109-112 cm)	23.5±0.2	0.7041	0.512919	5.5
AL5 (302-321 cm)	61.9±2.9	0.70414	0.512938	5.9
AL6 (372-377cm)	71.1±4.1	0.70395	0.512937	5.8
AL7 (380-387cm)	71.7±4.1	0.70401	0.512925	5.6
Ash deposits on Barren	(Relative Age)			
BI-07-TL-02	Precaldera	0.704	0.512954	6.2
BI-07-TL-03	Precaldera	0.70393	0.51297	6.5
BI07-TL-06	Precaldera	0.70396	0.512989	6.8
BI-08-TL-01	Precaldera	0.70397	0.512921	5.5
BI-08-TL-02	Precaldera	0.70401	0.512971	6.5
BI-07-TL-05	Unknown	0.70409	0.512928	5.7
BI-07-TL-04	Unknown	0.70404	0.512903	5.2
BI-07-TL-01 (2007)	Modern	0.70413	0.512876	4.6
BI-07-06	Modern	0.70404	0.5129	5.1
BI-09-02 (2009)	Modern	0.70411	0.512872	4.6
Narcondam Island Lava				
NCI-09-09	Unknown	0.70444	0.512754	2.3
NCI-09-19	Unknown	0.70503	0.51271	1.4

Table 3.2: Isotopic ratio data for ash layers in the core, ash deposits on Barren Island and lava flows of Narcondam Island.

Superscript 'a': ⁸⁷Sr/⁸⁶Sr is normalized to 0.71025 for NBS987 and ¹⁴³Nd/¹⁴⁴Nd is normalized to 0.511858 for LaJolla. Superscript 'b': $\epsilon_{Nd} = \{(^{143}Nd/^{144}Nd)_{sample}/(^{143}Nd/^{144}Nd)_{chondrite} -1\} \times 10^4$ and is calculated using an average (¹⁴³Nd/¹⁴⁴Nd)_{chondrite} of 0.512638. Superscript 'c' 1 sigma errors, estimated using standard error propagation method, on ages in yrs BP (ka).

The major element oxide data of the lithic fragments from these ash layers also support the inference that ash layers in the core originated from the Barren Island Volcano. In K_2O vs. SiO_2 and TiO_2 vs. Al_2O_3 plots (Fig. 3.8) the data for glass matrix of grains from the ash layers largely overlap with the field of Barren Island. In the core, therefore, it appears that the Barren Island Volcano was the only major eruptive centre in the Andaman Sea during the Late Pleistocene and Holocene and that its debris covered an extensive area around the volcano. Comparing these data with the volcanoclastics preserved in Miocene-Pliocene sedimentary deposits of Andaman Islands and Havelock Island, it was observed that the older ash deposits were completely different in composition and none of these are correlateable to present day volcanoes in the Andaman Sea. The compositions of plagioclase feldspars and pyroxenes in lithic fragments are also similar that reported in samples from Barren Island (Luhr and Halder, 2006; Pal et al., 2010b).



Fig. 3.7: Plot of ε Nd versus ⁸⁷Sr/⁸⁶Sr for the ash layers in the core, ash deposits on Barren Island, and volcanics from some major volcanoes of Indonesia (see Fig. 3.1a). Also shown is a field for isotopic compositions of lava flows on Barren Island. Data sources: Ash layer in the core, ash deposits on Barren Island, and lavas on Narcondam: this work; Barren Island lavas: Luhr and Haldar (2006); Chandrashekhram et al. (2009); Kumar (2011); lavas and ash from volcanoes of Indonesia: Turner et al. (2001). All data were normalized to NBS-987 (0.710250) and La Jolla (0.511858) for ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd isotopic ratios respectively.



Fig. 3.8: (a) $SiO_2 vs. K_2O$ and (b) $Al_2O_3 vs. TiO_2 cross plots showing diagram showing the compositions of the marine ash-layers in the core SK-234-60 compared with the field for the recent volcanics from Barren Island and tuff deposits from Rutland & Havelock. Data source: Barren Island (Kumar, 2011; Luhr and Haldar, 2006; Chandrasekharam et al. 2009; Pal et al., 2010b), Rutland & Havelock (Pal et al., 2005, 2010a).$

3.5 History of volcanism on Barren Island

3.5.1 The Barren Island Volcano

The Barren Island Volcano has been described as a stratovolcano, which rises from the depth of more than 2 km of the Andaman seafloor (Sheth et al., 2009a). The exposed part of the volcano represents only the top part of a vastly larger submarine volcano, mostly evolved below sea level. The Barren Island Volcano must have erupted several times in the past but it was only since 1789, the records of its eruptions have been documented. The first recorded eruption occurred during 1789-1832 and after long period of quiescence, the volcano again became active in 1991 (recent) and eruptions are still continuing (Shanker et al., 2001; Luhr and Haldar, 2006; Pal et al., 2007a; Sheth et al., 2009a, 2010) (Fig. 3.9b). In literature, the undated volcaniclastic deposits and lava flows exposed on the island are mainly divided into pre- and post-caldera volcanic episodes.



Fig. 3.9: (a) Ash cloud from 2010 eruption of Barren Island Volcano, finer ash is drifted away by the wind and deposited later in the sea nearby (b) Map of Barren Island showing various lava flows and ash distributed from the recent and past eruptions of the volcano.

The volcano has generally shown Strombolian and sometimes Plinian type of eruption. Figure 3.9 (a) shows a recent ash eruption of the volcano, the coarser ash particles show explosive ejection with ballistic trajectories unaffected by the wind while finer ones are drifted away by the wind and get deposited in the nearby sea. Probably ash layers in our core were also deposited in similar fashion. In the Andaman Sea, the predominant wind directions are southwesterly during summer (May-September) and northeasterly during winter (October-April). These winds

should promote dispersal mainly into the northeast and southwest part of the sea from the volcano. Our core location is in the southeast of the volcano. We, therefore, believe that the ashes deposited in our core were transported by surface currents from near the volcano and not by direct aerial dispersal, and likely represent major eruptions.

Our core recorded at least seven major eruptions of the volcano during the last ~74 kyr. Presence of the two distinct ash layers near the bottom of the core, with irresolvable ages due to high errors on the estimates, suggests a couple of major eruptions of the volcano at ~71 ka. Another major eruption is recorded at ~62 ka. Interestingly, the ~62 ka ash layer (AL-5) is the thickest in the entire record and contains the coarsest of volcanic grains (~2 mm) in the core. This suggests that the intensity of the ~62 ka eruption was quite large, and we speculate that this was the time the volcano had grown to near sea level or above it. Absence of any ash layer between AL-5 (302 cm, ~62 ka) and AL-4 (112 cm, ~24 ka) points to a significant hiatus between these large ash eruptions (Fig. 3.3). Since the eruption at ~24 ka, the volcano has had fairly regular major eruptions every ~5000 years until the ~8 ka eruption.

The variation of ε_{Nd} value of the ash layers of the core (Table 3.2) between the ~71 ka and the ~17 ka eruptions is very small (marginally higher than the analytical 2σ uncertainty of 0.2 units), which indicates that the Nd isotopic composition of magma of the Barren Island remained almost constant during this large time period. This is significant for the large time interval represented. The first marked change in magma composition is observed in the ~12 ka eruption, and since then the composition has been variable as is apparent from the variable ε_{Nd} of historic and recent eruptions on Barren Island (Table 3.2; Luhr and Haldar 2006; Chandrasekharam et al. 2009). The highest ε_{Nd} value measured for ash layers of the core is 6.7, a composition shown only by the lavas and ash deposits of the Barren Island that are exposed in the caldera wall of the volcano (ε_{Nd} >6.0), and which thus predate the caldera itself. In comparison, all historic and modern (caldera-filling) eruptions of the volcano have produced lava and ash with much lower ε_{Nd} values (~5.0 or less) (Table 3.2; Luhr and Haldar 2006;

	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P_2O_5	Cr ₂ O ₃	CoO	NiO	
Feldspar														
<u>AL-1</u>														
LF-1	46.16	0.04	33.61	0.76	0.02	0.12	17.31	1.68	0.03	0.24		0.03		Bytownite
LF-2	48.44	0.07	31.52	1.19	0.01	0.41	15.52	2.56	0.07	0.19		0.01	0.01	Bytownite
LF-3	55.66	0.43	25.25	2.66	0.04	0.40	10.03	5.00	0.31	0.18		0.05		Labrodorite
LF-4	53.30	0.74	24.39	4.77	0.03	0.88	10.93	4.35	0.30	0.22	0.02	0.02	0.04	Labrodorite
LF-5	53.23	0.10	28.07	1.34	0.05	0.31	12.21	4.31	0.16	0.16	0.02	0.01	0.03	Labrodorite
LF-6	52.34	0.09	29.17	0.92	0.05	0.11	12.55	4.46	0.08	0.16	0.03	0.03	0.01	Labrodorite
<u>AL-2</u>														
LF-1	49.68	0.08	30.85	1.03	0.01	0.11	14.72	3.16	0.06	0.19	0.01	0.06	0.03	Bytownite
LF-3	51.26	0.06	29.58	0.91	0.01	0.21	13.51	4.09	0.16	0.15	0.02	0.01	0.03	Labrodorite
LF-4	51.95	0.07	29.42	0.92	0.02	0.13	12.99	4.22	0.07	0.15	0.02	0.01	0.03	Labrodorite
LF-5	46.93	0.02	33.14	0.72	0.00	0.06	16.80	2.03	0.03	0.20	0.01	0.03	0.01	Bytownite
LF-6	54.65	0.08	27.52	0.98	0.04	0.14	10.81	5.52	0.09	0.09		0.07	0.01	Labrodorite
LF-7	48.01	0.03	32.35	0.57	0.02	0.04	16.17	2.50	0.06	0.19	0.01	0.01	0.04	Bytownite
<u>AL-3</u>														
LF-1	52.09	0.13	28.95	1.27	0.02	0.19	12.82	4.21	0.10	0.17	0.01	0.03	0.02	Labrodorite
LF-2	53.00	0.15	28.39	1.27	0.04	0.15	12.13	4.56	0.12	0.16	0.01	0.02		Labrodorite
LF-3	51.36	0.08	29.31	1.26	0.01	0.15	13.34	4.16	0.07	0.17	0.02	0.04	0.02	Labrodorite
LF-4	53.04	0.24	27.58	1.85	0.05	0.34	12.05	4.42	0.17	0.18	0.02	0.03	0.04	Labrodorite
LF-5	50.28	0.05	30.46	0.97	0.03	0.12	14.25	3.54	0.08	0.17	0.02	0.01	0.01	Labrodorite
LF-6	49.80	0.06	30.64	1.00	0.01	0.11	14.81	3.29	0.06	0.18		0.02	0.02	Bytownite
LF-7	53.86	0.16	27.50	1.60	0.02	0.23	11.38	4.91	0.14	0.16	0.03	0.00	0.02	Labrodorite
<u>AL-4</u>														
LF-2	47.45	0.03	32.47	0.87	0.06	0.16	16.72	2.05	0.03	0.20			0.02	Bytownite
LF-3	50.33	0.03	30.92	0.80	0.08	0.06	13.85	3.62	0.05	0.18		0.08	0.01	Labrodorite
LF-4	50.73	0.09	30.05	1.25	0.03	0.16	13.95	3.46	0.09	0.18	0.02	0.02	0.01	Bytownite
LF-5	52.13	0.25	29.21	1.87	0.03	0.34	12.04	3.63	0.27	0.21	0.02	0.03	0.01	Labrodorite
LF-6	55.72	0.24	28.42	1.60	0.02	0.21	8.55	4.85	0.14	0.19	0.02	0.01	0.02	Andesine
LF-7	52.23	0.26	27.86	2.18	0.01	0.43	12.74	3.92	0.18	0.16	0.03	0.01	0.01	Labrodorite
<u>AL-5</u>														
LF-1	54.70	0.04	28.00	0.56	0.04	0.06	11.44	4.70	0.22	0.19	0.01	0.01	0.02	Labrodorite
LF-2	49.95	0.03	30.93	0.60	0.05	0.10	14.81	3.24	0.07	0.14	0.02	0.03	0.02	Bytownite
LF-5	46.85	0.03	33.17	0.62	0.04	0.06	17.10	1.84	0.03	0.20	0.01	0.01	0.03	Bytownite
LF-7	51.92	0.30	27.76	2.26	0.07	0.47	13.02	3.75	0.26	0.15	0.02	0.01	0.01	Labrodorite

Table 3.3: Major element oxide contents (in wt %) of Plagioclase feldspar; pyroxene; olivine; titaniferrous magnetite, amphibole and spinel in the ash layers of the core SK-234-60

Table	5.5: Com	inued												
	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P_2O_5	Cr ₂ O ₃	CoO	NiO	
<u>AL-6</u>														
LF-1	49.95	0.16	30.01	1.60	0.02	0.59	14.25	3.04	0.10	0.20	0.02	0.03	0.02	Bytownite
LF-2	49.73	0.19	29.82	1.87	0.03	0.47	14.51	3.00	0.11	0.23	0.01	0.02	0.01	Bytownite
LF-3	48.68	0.07	31.43	1.03	0.04	0.19	15.59	2.66	0.06	0.20	0.03	0.01	0.01	Bytownite
<u>AL-7</u>														
LF-4	50.66	0.05	30.89	0.70	0.01	0.15	13.69	3.50	0.05	0.18	0.01	0.00	0.09	Labrodorite
LF-5	51.08	0.04	30.73	0.76		0.13	13.25	3.70	0.06	0.19	0.02	0.03	0.01	Labrodorite
LF-6	49.81	0.15	30.63	1.55	0.03	0.44	14.10	2.91	0.07	0.20	0.04	0.04	0.02	Bytownite
LF -7	48.91	0.07	32.05	1.10	0.04	0.19	14.79	2.53	0.06	0.22	0.01	0.01	0.03	Bytownite
Pvroxen	e													
AL-1														
LF-2	59.18	0.07	3.31	18.70	0.04	16.84	0.49	0.25	0.98	0.04		0.01	0.11	Clinoenstatite
LF-5	50.83	0.78	2.86	11.10	0.35	15.38	18.03	0.28	0.01	0.21	0.12	0.01	0.03	Augite
LF-7	53.60	0.33	1.43	16.90	0.49	25.49	1.65	0.03	0.00	0.02	0.01	0.02	0.02	Clinoenstatite
AL-3														
LF-5	51.40	0.55	2.27	8.62	0.26	16.66	19.18	0.33	0.01	0.26	0.35		0.11	Augite
AL-4														
LF-1	54.22	0.33	1.31	17.41	0.47	25.09	1.55	0.03	0.02	0.01			0.02	Clinoenstatite
LF-2	52.05	0.58	3.00	10.70	0.30	15.65	17.30	0.41	0.01	0.18			0.11	Augite
LF-3	53.34	0.23	1.24	17.79	0.68	25.18	1.42	0.04	0.01	0.02	0.02	0.03	0.01	Clinoenstatite
LF-6	45.69	0.35	8.88	20.78	0.36	19.16	1.89	2.39	0.19	0.26		0.05		Augite
AL-5														C
LF-1	51.78	0.41	1.52	10.59	0.37	13.95	20.78	0.35		0.21	0.04			Augite
LF-3	49.58	0.74	4.57	7.38	0.12	13.98	23.01	0.26	0.01	0.25	0.05	0.03	0.03	Diopside
LF-4	51.73	0.36	1.40	9.92	0.53	14.05	21.40	0.31	0.01	0.23	0.03	0.01	0.02	Augite
LF-7	50.84	0.51	3.50	5.93	0.14	14.88	23.22	0.23		0.25	0.47		0.04	Diopside
<u>AL-6</u>														-
LF-1	52.39	0.49	2.51	7.89	0.23	16.60	18.85	0.32	0.02	0.25	0.38	0.05		Augite
LF-2	51.89	0.52	3.43	6.60	0.12	16.34	20.06	0.22	0.01	0.25	0.47	0.02	0.09	Augite
Olivine														
AL-1														
LF-4	38.06	0.01	0.10	20.90	0.31	40.22	0.20	0.01	0.01	0.03	0.03		0.10	Chrysolite
AL-4														• • •
LF-2	39.29	0.01	0.07	18.70	0.33	41.53	0.28					0.09	0.01	Chrysolite
LF-6	39.00	0.03	0.51	22.31	0.39	37.14	0.25	0.08	0.01	0.13	0.08	0.08		Chrysolite

Table 3.3: Continued

Table 3.3: Continued

Chrysolite
0.32 Chrysolite
0.16 Chrysolite
0.16 Chrysolite
0.13 Chrysolite
0.18 Chrysolite
0.02
0.01
0.02 Ferroan Pargasite
0.13 Cr-Spinel

*Values are normalised to a total of 100 wt. %. AL: Ash Layer; LF: Lithic Fragment. Minerals are classified on the basis of their chemical compositions using software Minpet 2.02 (Richard, 1995).

	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P_2O_5	Cr ₂ O ₃	СоО	NiO	Mg#	FeOt/MgO
AL-1															
LF-2	54.49	1.78	13.63	14.35	0.17	8.34	4.64	1.80	0.59	0.16	0.01	0.01	0.02	50.86	1.72
LF-3	60.17	1.63	15.82	8.66	0.21	2.26	6.11	3.83	0.97	0.31	0.01	0.02	0.01	31.75	3.83
LF-4	53.80	1.44	17.11	9.06	0.23	3.51	10.18	3.79	0.57	0.24	0.03	0.01	0.03	40.85	2.58
LF-5	58.77	0.87	10.22	9.37	0.17	6.68	10.42	1.99	1.12	0.32	0.06	0.01		55.96	1.40
LF-6	59.49	1.53	17.09	7.83	0.18	1.64	6.32	4.74	0.82	0.30	0.02	0.03	0.04	27.18	4.77
AL-2															
LF-1	56.32	1.72	15.97	10.00	0.21	2.67	8.15	3.86	0.75	0.30	0.01	0.02	0.04	32.24	3.74
LF-2	56.85	1.23	16.79	9.19	0.23	3.19	7.77	3.84	0.64	0.21	0.02	0.01	0.03	38.20	2.88
LF-4	50.72	1.54	7.29	11.70	0.41	10.42	16.10	1.37	0.17	0.27	0.02			61.35	1.12
LF-6	59.23	1.63	15.66	9.06	0.22	2.62	6.66	3.73	0.81	0.31	0.02	0.04	0.03	34.01	3.46
<u>AL-3</u>															
LF-1	56.19	1.69	15.72	10.00	0.22	3.24	7.89	3.88	0.78	0.30	0.02	0.04	0.04	36.59	3.09
LF-2	57.41	1.66	15.81	9.24	0.24	2.50	7.68	4.21	0.86	0.33	0.02	0.01	0.01	32.53	3.70
LF-3	55.84	1.61	16.96	9.60	0.21	2.64	8.19	3.81	0.79	0.28	0.02	0.02	0.03	32.88	3.64
LF-4	57.23	1.26	17.77	7.86	0.20	2.52	7.54	4.46	0.81	0.28	0.02	0.04	0.02	36.36	3.12
LF-5	59.56	1.12	17.85	6.53	0.05	1.78	7.19	4.73	0.89	0.31		0.01		32.68	3.67
LF-6	55.98	1.77	14.79	11.16	0.25	3.51	7.61	3.75	0.82	0.32	0.01	0.02	0.02	35.92	3.18
LF-7	55.97	1.31	19.20	7.62	0.16	1.93	8.42	4.46	0.64	0.26	0.01	0.01	0.01	31.06	3.95
<u>AL-4</u>															
LF-1	61.85	1.04	16.88	6.83	0.19	1.95	5.17	5.00	0.94	0.28	0.03	0.02	0.01	33.72	3.50
LF-2	56.33	1.35	17.46	8.22	0.14	2.43	8.87	4.31	0.72	0.22	0.02	0.06	0.01	34.51	3.38
LF-5	62.70	1.16	15.52	7.65	0.10	2.54	4.73	4.23	1.11	0.31	0.02	0.02	0.01	37.14	3.02
LF-6	59.51	1.33	16.95	8.82	0.19	2.23	5.79	4.16	0.60	0.33	0.02	0.06	0.02	31.05	3.96
LF-7	57.15	1.79	16.02	10.10	0.11	2.38	7.69	3.67	0.88	0.28	0.01	0.03	0.01	29.60	4.24
<u>AL-5</u>															
LF-7	58.46	1.33	17.12	8.02	0.09	2.47	7.64	3.88	0.72	0.22	0.01	0.01	0.02	35.40	3.25
<u>AL-6</u>															
LF-1	54.66	1.32	16.37	9.43	0.19	4.03	9.31	3.79	0.58	0.25	0.03	0.01	0.02	43.22	2.34
LF-2	53.85	1.34	15.35	10.36	0.22	5.41	9.21	3.34	0.57	0.25	0.02	0.00	0.05	48.21	1.91
LF-3	54.97	1.42	15.99	9.90	0.22	3.80	9.18	3.54	0.65	0.27	0.02	0.00	0.02	40.65	2.60
<u>AL-7</u>															
LF-4	56.48	1.53	14.46	10.85	0.26	4.44	7.63	3.38	0.65	0.27	0.02	0.02		42.16	2.44
LF-5	55.53	1.38	15.13	9.82	0.22	4.39	8.80	3.79	0.59	0.27	0.05	0.02	0.03	44.34	2.24
LF-6	54.32	1.29	15.66	10.01	0.16	5.51	8.73	3.42	0.50	0.27	0.05	0.03	0.05	49.52	1.82
LF-7	54.84	1.34	15.34	9.88	0.23	5.07	9.44	3.06	0.45	0.28	0.04	0.01	0.02	47.75	1.95

Table 3.4: Major element oxide contents (in wt %) of glass matrix in the ash layers of the core SK-234-60

*Values are normalised to a total of 100 wt. %. AL: Ash Layer; LF: Lithic Fragment.

Chandrasekharam et al., 2009). This relationship may be an indication that the caldera of the Barren Island volcano is younger than ~8 ka, the age of the youngest ash layer in the core with the ε_{Nd} value of 6.7. Absence of an identifiable ash layer younger than ~8 ka is well in accord with the hypothesis that the caldera of Barren Island formed as a result of a single, simple, symmetric collapse along circular (ring) faults (Sheth et al., 2009a), and was not associated with a major eruption as speculated by Shanker et al. (2001). Unlike ε_{Nd} values, 87 Sr/ 86 Sr ratios of ash layers in the core overlap with each other, and with that of the ash and lavas on Barren Island (Table 3.2; Luhr and Haldar 2006; Chandrasekharam et al. 2009), and therefore, are of little use as stratigraphic markers. Also, 87 Sr/ 86 Sr ratio is susceptible to alteration by sea water. Although an attempt has been made here to reconstruct the evolutionary history of the Barren Island Volcano, we acknowledge the fact that data from a single core may not be sufficient to reconstruct the complete eruption history of the volcano, since it is quite possible that many of the past ash eruptions of the volcano were simply not recorded at our core site because of their dispersal in other directions.

3.5.2 Geochemistry of Ash Layers and evolution of the Barren Island Volcano

Major element compositions of glass matrix of the lithic fragments from the ash layers of the core determined using EPMA, are presented in Table 3.4. Based on Total Alkalis-Silica (TAS) classification of volcanic rocks (Fig. 3.10) we determine these fragments to represent sub-alkalic volcanic magma and basaltic to andesitic composition. The majority of these fragments show basaltic to andesitic composition similar to what is observed for most subaerial lava flows on the volcano. This result not only confirms that the all ash layers are generated by the Barren Island Volcano but also suggests that the parental magmas during various eruptions in the past were more evolved than the modern flows (Fig. 3.10).

From the major oxide data it is apparent that the ash in a given layer represents an evolving magma during that particular volcanic event, as various oxide contents in the glass matrix show fractional crystallization trends in plots against MgO content (Fig.

3.11). Mg# (= [atomic Mg/(Mg + Fe²⁺)] x 100) in a given layer is highly variable (Table 3.4), which suggests that the lithic fragments were erupted at various times during an eruption, when the magma was continuously evolving due to fractional crystallization. Decrease in CaO, FeO^T, CaO/Al₂O₃ and increase in SiO₂, Al₂O₃ with decreasing MgO (Fig. 3.11) in glass matrices are indicative of plagioclase and pyroxene fractionation from the magma before it got quenched due to its sudden separation and violent ejection out of the magma chamber. Similar inferences can also be made about fractionation of other minerals from the magma (e.g. titanomagnetite) prior to their removal of the liquid with which it was in equilibrium, as ash grains.



Fig. 3.10: Total Alkalis-Silica (TAS) classification (Le Bas et al., 1986) for glass matrix from lithic fragments for ash layers in the core SK-234-60. For comparison the field for volcanics from Barren Island is also shown. Data sources: Barren Island (Luhr and Haldar, 2006; Chandrasekharam et al., 2009; Pal et al., 2007a, 2010b; Kumar, 2011). The boundary between the alkalic and subalkalic rocks is after Macdonald and Katsura (1964).



Fig. 3.11: Various major element oxides vs. MgO plots for glass matrix of lithic fragments from the ash layers in the core SK-234-60 compared with field for Barren Island volcanic. Data sources: (Kumar, 2011; Luhr and Haldar, 2006; Chandrasekharam et al., 2009; Pal et al., 2010b).

Chapter-4 Provenance of sediments deposited in the Andaman Region

4.1 Introduction

Sediments are either produced by surface processes like weathering and erosion or by (bio) chemical precipitation. Deposited on dry lands and within bodies of water, sediments, store valuable information about processes that occur on the surface of the Earth. In fact, sedimentary records are the main source of our knowledge about tectonics, palaeoclimate, palaeogeography and evolution of life on our planet. Determining the provenance of sediments deposted in a basin is one of the most important aspects of studying sedimentary rocks. In geology, finding out provenance is not only restricted to deciphering lithologic origin of sediments but also involves reconstruction of their parent rock compositions, depositional history, and tectonoclimatic conditions during their deposition. It has been realized that making inferences about sedimentary provenances is not simple. It requires detailed understanding of nature and extent of compositional and textural modifications suffered by sediments during weathering, erosion, deposition and diagenesis. To avoid any bias resulting from over-dependence on a single method, use of multiple proxies is preferred. Over the years, several techniques such as petrography, mineralogy of heavy and clay minerals and elemental and isotope geochemistry have been successfully developed and utilized in such studies (Thurach, 1884; Mackie, 1899; Krumbein and Pettijohn, 1938; Weltje and Eynatten, 2004 and references therein; Taylor and McLennan, 1985; Bhatia, 1983; Bhatia and Crook, 1986; Roser and Korsch, 1986).

In this work, we have used geochemical and isotopic tracers to determine provenances, understand tectonic settings and climatic conditions during the deposition of sediments in the forearc (accretionary prism) and backarc (Andaman Sea) basins of the Andaman Subduction Zone. The sedimentary sequences of the Andaman forearc basin are believed to have started depositing during the Paleocene and therefore, likely to have preserved records of major geological events on and around Indian Plate since then. Information on the provenances and tectono-geomorphic conditions during the deposition of the sediments that make up the sedimentary rocks of the Andaman Islands is meagre. Therefore, a comprehensive geological, isotopic and geochemical study of these rocks is required to settle much of the outstanding issues related to the regional geology, stratigraphy, major tectonic changes and evolution of these islands.

The sedimentary record of the Andaman Sea, which came into existence during late Miocene or early Pliocene, provides another time window to study the provenances of sediments contributed to this region of South-East Asia and understand their transport dynamics and variations with changing climate and tectonics. Although at present, the basin largely receives sediments from the Irrawaddy, Salween, and Sittang Rivers of Myanmar (Rodolfo, 1969), it is yet to be established if the scenario were similar in the past. To answer this question and to understand the evolution of the basin as a whole, a comprehensive study of sedimentary record was deemed necessary.

Prior to the final collision of the Greater India with southern margin of the Eurasia in the north and ancient Myanmar in the east, the Sunda Arc extended well up to Pakistan as the Jurassic–Paleogene Trans-Himalayan Arc and suture zone (Mitchell, 1993). It is believed that this arc and suture zone formed an ancient Andean type margin along the Neo-Tethys (Chu et al., 2006; Scharer and Allegre, 1984) and marked the boundary between the Eurasian plate and the Indian plate (Clift et al., 2001; Sinclair and Jaffey, 2001). The final collision of these two plates occurred at ~50 Ma (Rowley, 1996; Hodges, 2000) and subsequently the Neo-Tethys started closing. Most of the authors propose that the Himalayan thrust belt started developing on the Indian plate margin as a result of continent-continent collision during the Neogene (e.g., Johnson and Alam, 1991; Rahman and Faupl, 2003; Uddin and Lundberg, 1998), while the eastern sector (Myanmar) of the subduction zone continued to remain active. A trench-forearc basin opened up along the subduction trench, on the east of the Indian plate margin, while on

the Indian plate two remnant ocean basins, Katawaz and Bengal, remained active (Alam et al., 2003; Qayyum et al., 2001).

Sedimentary records of the suture zone show that until ~55 Ma the Trans-Himalayan landmass was the predominant source of sediments, while the Greater Indian landmass contributed only subordinately to the basins in the north (Garzanti and Vanhaver, 1988; Wu et al., 2007; Sinclair and Jaffey, 2001). The foreland and Bengal basins lying south of the Himalaya observed multiple changes in provenances of sediments they received during the late Paleocene to mid Eocene, however, only one such change has been reported at the beginning of the Oligocene (Najman et al., 2008). It has been shown by Najman et al. (2008) that the major changes in sediment provenance in the Bengal basin (from ~38 Ma during the deposition >1 km thick sand beds of the Barail Group) are related to the increase in contribution from the thrusted and uplifted Indian crust along the nascent Himalayan mountain belt. Another study by Lindsay et al. (1991) also reported large scale deposition of materials derived from the Himalaya in the northern Bengal basin at ~40 Ma. Such changes have also been observed in the Andaman Islands during the Oligocene (Pal et al., 2003; Allen et al. 2007). This synchronous change in provenance across the basins south of the Himalaya hint at occurrence of a major tectonic (thrusting) event in the Himalaya during this time. Another such major geological event, evident from regional upliftment and stratigraphic changes in the Irrawaddy delta, Mergui Basin, and in the Indo-Burman Ranges, has been reported at ~20 Ma (Acharyya et al., 1989).

4.2 Potential source for Andaman forearc and backarc deposits

In order to characterize the provenances of sediments deposited within the Andaman forearc and backarc basins it is essential to first locate and understand potential sources. In this section we identify four such source regions and describe their chemical characteristics. Table 4.1 lists these sources and their important features. The potential sources identified for both the forearc and backarc basin deposits are: Suprasubduction and Arc magmatic rocks, rocks from the Himalayan-Trans-Himalayan region, rocks of

the Indian Shield and Myanmar. Figure 4.1 shows most probable paleogeographic configuration of various tectonic blocks in south Asia prior to the closure of Neo-Tethys. These blocks contain the potential sources of sediments deposited in the Andaman forearc basin during the Paleogene.

4.2.1 Suprasubduction and Arc igneous sources

These include the suprasubduction ophiolite and magmatic arc rocks of the region. Such sources were present all along the Sunda Arc in the south to the Himalayan Arc and suture zone in the north (Fig. 4.1). As discussed in earlier chapters the ophiolite sequence from the Andaman consists of mantle rocks, plagiogranites, mafic plutonic and intrusives, and extrusive lava series- most of which have been altered. The magmatic arc sources from the region have mafic, intermediate and felsic volcanic products (Chhibber, 1934, Mitchell, 1985; Wakita, 2000; Hall 2002, 2009; Van Bemmelen, 1949). The only isotopic study on the ophiolites of the Andaman Islands (Kumar, 2011) reveals that their ⁸⁷Sr/⁸⁶Sr ratio varies between 0.70342 and 0.70483, whereas ε_{Nd} varies between +6.4 and +11.7. The ophiolitic and early magmatic arc sources would have become important during the initial trench and forearc basin fillings, whereas the recent volcanism (<4 Ma) would have acted as sources to the back arc fillings.

4.2.2 Indian Shield Sources

The Indian shield lying west of the Andaman Basin was made up of 5 cratons- Aravalli, Bundelkhand, Singbhum, Bastar and Dharwar joined by mobile belts (Fig. 4.1). These mostly consisted of Archean gneisses, granites and schists and Proterozoic sedimentary successions. Materials derived from these cratons would have been predominantly arkosic with minerals reflecting old continental crust of Precambrian–Late Palaeozoic age (Najman et al., 2008; Misra and Johnson, 2005; Auge et al., 2003; Mishra et al., 1999). ε_{Nd} of materials derived from Archean rocks should have been \leq -30 (Najman et al., 2008; Peucat et al., 1989; Saha et al., 2004). Considering the paleogeographic positions of various cratons and major drainage patterns during the Paleogene, we

Source Regions	Lithology	U-Pb zircon ages (Ma)	Ar-Ar mica ages (Ma)	⁸⁷ Sr/ ⁸⁶ Sr	$\epsilon_{\rm Nd}(0)$
Igneous Sources					
Andaman and Indo-Burman	Suprasubduction ophiolites	Late Cretaceous (93.6±1.6 Ma, 95±2 Ma)		0.7034-0.7048	+6.5 to +11.8
	Magmatic arc volcanics				
<u>Himalaya</u>					
Higher Himalaya	Quartzofelspathic. Med-high grade metamorphic rocks	Cambro-Ordovician to Archean (500 to >3000 Ma. Pks: 1100, 1500-1700, 2500)	Tertiary (dominant)	0.7331-0.9062	-20 to -5
Tethyan Himalaya & Haimanta Group	Sedimentary & low grade metamorphic rocks	Cambro-Ordovician to Archean (500 to >3000 Ma Pks: 500, 1100, 2500)	Pre-Tertiary; < 950 Ma, mostly <500 Ma (from Cambro-Ord granites)	0.7158–1.1764	-30 to -16
Arc & Suture Zone	Batholiths (Gangdese & Nyainquentanglha), Ophiolites	Jurassic-Eocene (<200 Ma, Numerous peaks: 50, 80-90, 150, 200 Ma)	Cretaceous and Tertiary		+1 to +8
Indian Shield					
Chotanagpur Proterozoic gneissic belt (river sand data)	Gondwanan seds. Gneiss. Granites. Arkosic sediments	Proterozoic (950-1450, pk: 1350-1400 Ma)	Proterozoic (788-938 Ma)	0.7625-0.9724	-13.8
Shillong Plateau (river sand data)	Precambraian-Late Palaeozoic gneiss & granite	Proterozoic & Cambro-Ordovician (500 1800 Ma)	Cambro-Ordovician (467- 524 Ma)	≥0.735	-14.6
Myanmar					
Shan-Thai block	Proterozoic-Cretaceous sedimentary rocks on schist basement.	Tertiary aged zircons along the Burma margin	Mica ages along Shan Scarp: 26–16 Ma		
	Cretaceous arc rocks, Mogok schists, gneisses and intrusives	I-type granitoids ages: 120–150 Ma	K–Ar mineral dating of batholiths: 79–100 Ma		

Table 4.1: Summary of salient features of potential source regions for Andaman Sediments

Data sources: Allen et al., (2007); Allen et al., (2008); Najman et al., (2008); Singh and Krishna, (2009); Zhao et al., (2009); Pederson et al., (2010); Sarma et al., (2010) and Ullah, (2010);.



Fig. 4.1: Schematic map (modified after Replumaz and Tapponnier, 2003) showing the potential source regions for sediments deposited in the Andaman forearc basin during the Paleogene. Data sources: Sharma et al. (1994); Jayananda et al. (2000); Allen et al. (2007); Allen et al. (2008); Najman et al. (2008); Sheth et al. (2009b); Singh and Krishna (2009); Zhao et al. (2009); Ullah (2010) and Shukla (2011).

believe that the parts of the Indian shield that would have contributed sediments to the Andaman trench-forearc basin were the Proterozoic Chotanagpur and Shillong Plateaus (Fig. 4.1).The sediments carried by rivers draining through these areas deposited detrital zircons and micas having (fission track and Ar–Ar) ages older than 467 Ma, with ε_{Nd} values varying from -15 to -13 (Table 4.1, Najman et al., 2008).

4.2.3 Himalayan and Trans-Himalayan Sources

The rocks that today occupy the trans, higher and lesser Himalayas were part of either the greater Indian plate or the sediments deposited in Neo-Tethys. Going by the present day configuration one would expect that the potential sources of the forearc sediments were mainly consisted of rocks of the Jurassic–earliest Tertiary batholiths of the Trans-Himalaya (Scharer and Allegre, 1984), the south-directed thrust belts of ophiolites and Palaeozoic–earliest Tertiary Tethyan sediments (DeCelles et al., 2001; Maheo et al., 2004). The Higher Himalayan sources would have included Oligocene-Miocene metamorphic rocks (Hodges, 2000; Vance and Harris, 1999) and those in the Lesser Himalayan were Precambrian to Cretaceous age, weakly to non-metamorphosed rocks of the Indian plate (Hodges, 2000; Richards et al., 2005). The ε_{Nd} data from the Trans-Himalayan Arc and suture zone sources today vary from +1 to +8 (Clift et al. 2001; Maheo et al. 2004), while the Higher and Lesser Himalayan sources have values varying from -30 and -5 (Ahmad et al., 2000; Robinson et al., 2001; Richards et al., 2005, Zhao et al., 2009).

4.2.4 Myanmar Sources

The Burmese margin to the east of the Andaman forearc basin was characterised by the older Cretaceous Mogok Metamorphic belt during the Paleogene. This belt continued north-west as the Trans-Himalayan (*Gangdese Batholith, Lhasa and Karakoram Terranes*) ancient active margin of Asia (Barley et al., 2003; Mitchell, 1993) (Fig. 4.1). In Myanmar, the active margin was represented by the Tertiary volcanic belt containing the now-extinct volcano- Mt. Popa (Stephenson and Marshall, 1984). The tectonic belt to the east, continued south up to Thailand and exposed rocks of Carboniferous to Permian age (Garson et al., 1976) that were intruded by tin-tungsten bearing granite plutons of Paleocene age having ⁸⁷Sr/⁸⁶Sr ratio \geq 0.717 (Mitchell, 1985).

4.3 Provenance of sediments of the Andaman Islands

4.3.1 Backgound and Earlier work

Of the five groups of sedimentary rocks of the Andaman and Nicobar Islands, the lithostratigarphy of only two, the late Paleocene-Eocene Mithakhari and the Oligocene Andaman Flysch, are dominated by siliciclastic sediments and therefore, we focused our study of provenances on them. The nature of provenance for the sediments of the Andaman forearc, particularly the Andaman Flysch Group, has remained controversial. Many earlier workers (e.g. Karunakaran et al., 1968a and Pal et al., 2003) believed that Irrawaddy sediments were the main source for the Andaman Flysch Group rocks, wheras others (e.g., Moore et al., 1982 and Curray et al., 1979, 2005) believed that these sediments were originally derived from the emerging Himalayas and were transported by the rivers draining into the Bengal Fan. It is, however, not clear from these studies that whether these sediments were deposited directly in the forearc or got offscraped from the subducting Indian slab at a later time. Other sources, which were thought to have contributed sediments to the Andaman forearc, were the northward-drifting greater Indian shield and eastern Myanmar (Allen et al., 2007).

In a recent work, Allen et al. (2007) made attempts to determine provenances of sediments that form the Andaman accretionary prism. Based on petrographic, isotopic (¹⁴³Nd/¹⁴⁴Nd of apatites) and mineral (apatite and zircon) cooling ages they proposed that the sediments of the Mithakhari Group were predominantly derived from the nearby arc sources, with subordinate contributions from an older continental source containing sedimentary and low-grade metasedimentary rocks- most likely located on the northeastern region of Myanmar. They also proposed that the sediments of the Andaman Flysch Group were derived from "recycled orogenic" sources consisting of very low to high rank metapelite-metapsammite sediments and a subordinate arc source. In the absence of appropriate geochemical and/or isotopic data from the sources located in Myanmar, discrimination between Himalayan-Transhimalayan and Myanmar sources was not possible and hence, the authors could not rule out completely the contributions from the Himalayan-Transhimalayan sources to the Andaman forearc sediments. U-Pb

ages of detrital zircons and Ar–Ar ages of detrital white micas from Andaman sediments also suggest contributions from older continental crust as well as younger arc sources (Allen et al., 2007). The Ar-Ar ages of detrital micas from the Mithakhari Group fall between the Paleozoic to Mesozoic, whereas those from the Andaman Flysch fall mainly between Late Mesozoic and Tertiary. The U-Pb detrital zircon ages from the Andaman Flysch span from 100 to 1500 Ma, with two strong modes in Proterozoic and Cretaceous–Eocene. From mineralogical studies, plagioclase, quartz, K-feldspar, pyroxene, epidote, sphene, green-brown hornblende, chromian spinel, apatite, garnet, rutile, and titanium oxides have been reported from the Mithakhari Group while quartz, K-feldspar, plagioclase along with zircon, tourmaline, rutile, and chromian spinel have been reported from the Andaman Flysch Group. From these studies it appears that Andaman sediments contain very little, if any, material from the Cratons in the Indian shield, however, many contain materials derived from the Proterozoic sedimentary sequences of the Greater India including those presently located within the Himalayan mountain belt.

4.3.2 Results of our study

(A) Petrography

Petrographic study on sedimentary rock samples from various formations of the Andaman sequences reveals that these formations have not undergone metamorphism. Important textural and mineralogical characteristics of the formations are described below.

1) Mithakhari Group

a) Hopetown Conglomerate

The Hopetown conglomerate near Chidiyatapu primarily contains feldspars with subordinate quartz, lithic grains and fossils, mainly foraminifers (Fig. 4.2a & b).

b) Namunagarh Grit

The Namunagarh Grit unit, extensively present as coarse grained sandstones, contains quartz, plagioclase and lithic fragments. Both monocrystalline and polycrystalline varieties of quartz have been observed. Fractures are common within these quartz crystals that probably reflect the effect of tectonics (Fig. 4.2c). We observe pumice fragments, glass shards and grains of plagioclase and pyroxene cemented in a siliceous matix (Fig. 4.2d). The presence of angular mineral grains suggests short distance transportation.



Fig. 4.2: Photomicrographs of thin sections: (a) & (b) Hopetown conglomerate unit of the Mithakhari Group showing shells of foraminifers (Nummulites) in coarse gritty matrix (c) Namunagarh grit unit of the Mithakhari Group (d) Tuff in the Mithakhari Group, Middle Andaman.

2) Andaman Flysch Group

The sandstone and shale formations of the Andaman Flysch Group are made up of quartz, feldspars and clay minerals (Fig. 4.3b). In comparison to the Mithakhari Group

rocks, these are much finer in grain size and contain more quartz. Accessory minerals include zircon, tourmaline, rutile and spinel.

3) Archipelago Group

The limestone formations of the Archipelago Group are fossiliferrous and contain shells of cephalopods, gastropods and foraminifers. They also contain sand grains and lithic fragments of volcanic origin. Figures 4.3c & d show photomicrographs of thin sections of two such formations.



Fig. 4.3: Photomicrographs of thin section showings: (a) Grains of quartz in a fine grained matrix from turbidites of Kalipur (b) Shale of Andaman Flysch Group (c) Baratang Cave limestone showing shell of gastropod along with fragments of quartz and feldspar (d) Foraminiferal limestone of Archipelago Group from Havelock Island.

(B) Elemental Geochemistry

Concentrations of major and trace elements, including rare earth elements (REE), in the samples from the siliciclastic formations of the Mithakhari and Andaman Flysch Groups, are presented in Tables 4.2, 4.3 and 4.4. The Chemical Index of Alteration (CIA) (Nesbitt et al., 1996) for each sample, as presented in the Table 4.2, is calculated using molecular proportions of oxides as follows:

 $CIA = [Al_2O_3/(AI_2O_3 + K_2O + Na_2O + CaO^*] 100$

where CaO* represents CaO contents of silicate minerals only. Figure 4.4 shows variations of the major element oxide versus SiO₂ contents in all our samples. Trace element contents, isotopic compositions (87 Sr/ 86 Sr and ϵ_{Nd}) and Nd-model ages (T_{DM}) for these samples are presented in Tables 4.3 (Mithakhari Group) and 4.4 (Andaman Flysch Group). Figure 4.5 shows primitive mantle-normalized trace element patterns (a & b) and chondrite-normalized REE patterns of these samples (c & d). From the geochemical data we make following observations.

- The rocks of the Mithakhari Group are characterized by lower SiO₂ contents, lower K₂O/Na₂O and intermediate to higher Fe₂O₃^T and MgO contents in comparison to the rocks of the Andaman Flysch Group and Post-Archean Australian Shale (PAAS) (Fig. 4.4).
- 2. The CaO content in these rocks are much higher than that in the Andaman Flysch rocks (Fig. 4.4d). Considering that these rocks do not contain much carbonates, we infer that high CaO is a result of high plagioclase content, which is supported by high Na₂O abundance in these rocks (Fig. 4.4f).
- The Andaman Flysch sediments, apart from having higher SiO₂, also contain higher amounts of Al₂O₃ and K₂O.
- 4. Sample AND-09-41, a siliceous limestone, is characterized by very low SiO₂ and high CaO content.
- 5. Unlike other Mithakhari formations, AND-09-48, a shale unit, has very high K_2O/Na_2O ratio.

(wt. %)	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ ^T	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	LOI	K ₂ O/Na	20 CIA
Mithakhari (Group												
AND-09-02	54.03	1.05	14.57	10.32	0.33	5.30	6.10	8.13		0.17	6.45		50.58
AND-09-03	67.39	0.82	14.99	6.58	0.10	3.62	2.47	2.72	1.19	0.12	3.60	0.44	70.75
AND-09-06	60.94	1.06	16.82	10.16	0.44	2.17	4.70	3.18	0.45	0.08	4.65	0.14	67.17
AND-09-09	70.53	0.61	12.32	9.80	0.10	2.68	2.26	1.15	0.33	0.22		0.29	77.04
AND-09-10	63.10	0.68	14.52	7.33	0.18	4.57	5.24	3.20	1.08	0.11	5.15	0.34	61.00
AND-09-11	69.09	0.49	13.02	3.43	0.24	3.36	7.46	1.52	1.31	0.10		0.86	56.39
AND-09-13	71.65	0.70	14.20	5.99	0.09	2.31	2.05	2.36	0.55	0.10		0.23	74.49
AND-09-15	50.43	0.63	13.74	5.98	0.12	5.77	17.31	5.48	0.43	0.11		0.08	37.22
AND-09-16	55.15	1.04	14.60	12.30	0.28	6.82	3.14	6.54		0.13			60.12
AND-09-18	59.43	0.88	15.34	6.66	0.24	2.06	12.99	1.46	0.84	0.09		0.58	50.37
AND-09-20	61.75	1.30	15.74	10.82	0.14	2.91	1.89	4.97	0.34	0.14	2.75	0.07	68.78
AND-09-23	63.22	1.26	16.59	10.26	0.09	3.31	3.25	1.43	0.37	0.22		0.26	76.94
AND-09-24	55.89	0.62	12.56	3.30	0.34	1.51	23.33	1.82	0.45	0.17	12.80	0.25	32.97
AND-09-27	73.53	0.60	14.94	5.34	0.02	2.48		1.13	1.92	0.05	3.95	1.70	84.08
AND-09-28	69.73	0.63	16.89	1.99	0.07	1.32	4.51	4.63	0.12	0.10		0.03	64.60
AND-09-30	62.07	1.10	13.74	10.77	0.15	4.67	3.25	3.47	0.61	0.17	3.25	0.17	65.58
AND-09-37	67.47	0.61	12.10	4.29	0.28	2.09	10.01	2.73	0.32	0.09	6.18	0.12	48.18
AND-09-41	33.43	0.26	3.06	2.40	0.80	1.93	36.01	0.00	0.16	0.28	31.84		7.81
AND-09-44	62.70	0.83	15.83	9.90	0.11	4.26	3.05	1.42	1.78	0.11	6.30	1.25	72.41
AND-09-48	64.58	1.09	16.07	10.41	0.30	1.96	2.51	0.51	2.47	0.10	5.90	4.85	76.59
AND-09-49	64.39	0.89	14.90	9.40	0.47	2.89		6.20	0.78	0.09	3.35	0.13	68.34
AND-09-51	75.16	0.72	11.95	6.89	0.08	2.50		2.60		0.11			82.15
AND-09-52	67.51	1.02	17.19	8.38	0.01	2.27		1.09	2.49	0.03		2.28	84.16
AND-09-55	64.11	0.75	14.05	8.84	0.13	4.17	1.79	5.59	0.43	0.13	2.45	0.08	64.41
PB-08-07	67.40	1.07	17.36	7.75	0.09	2.29		1.27	2.61	0.16		2.05	83.76
PB-08-08	67.66	0.97	16.19	6.76	0.06	3.35	0.28	2.54	2.06	0.14		0.81	78.13
PB-08-13	65.32	0.85	14.82	8.67	0.11	4.62	2.27	2.05	1.20	0.10	3.25	0.59	74.02
<u>Andaman Fl</u>	vsch Grou	p											
AND-09-04	65.90	0.92	16.88	7.35	0.18	2.55	1.89	2.00	2.18	0.14	4.19	1.09	74.25
AND-09-05	67.38	0.90	15.64	7.61	0.13	2.29	1.84	2.17	1.92	0.11		0.89	73.94
AND-09-12	74.11	0.93	15.82	4.15	0.04	1.50		0.94	2.41	0.11	1.90	2.57	84.42
AND-09-14	77.57	0.58	13.02	3.21	0.02	2.13		1.80	1.61	0.07	3.09	0.89	79.91
AND-09-36	68.30	0.87	17.57	6.22	0.05	2.30		2.01	2.51	0.16	3.05	1.25	81.28
AND-09-43	65.88	1.08	18.78	7.79	0.05	2.20		1.08	3.02	0.13		2.80	83.50
AND-09-61	67.59	1.15	24.50	1.11	0.01	1.40		0.72	3.48	0.05	6.25	4.86	86.64
PB-08-09	68.73	1.19	15.68	7.52	0.20	2.03	0.65	1.36	2.52	0.13		1.84	79.57

Table 4.2: Major oxide compositions (in wt. %) for siliciclastic sedimentary rocks of the Mithakhari Group and Andaman Flysch Group of Andaman Islands

All values are normalized to 100% on a volatile free basis. LOI= Loss on Ignition

	AND-09-02	AND-09-03	AND-09-06	AND-09-10	AND-09-19	AND-09-20	AND-09-22
(ppm)							
Rb	6.80	54.80	41.45	41.47	28.10	19.37	13.32
Sr	194.39	70.24	53.47	134.01	41.37	136.63	35.47
Ba	518.95	96.42	110.77	126.30	81.84	118.60	64.28
Y	79.63	9.01	13.83	12.23	15.23	18.85	19.33
Zr	64.80	16.37	24.20	24.12	23.94	61.40	60.61
Hf	2.20	1.07	1.21	1.22	1.52	2.30	2.16
Nb	2.35	5.98	3.43	3.46	3.85	3.83	3.51
Та	0.21	0.53	0.26	0.26	0.25	0.24	0.22
Th	2.59	8.64	5.55	5.55	3.06	2.12	2.28
U	0.94	1.38	0.87	0.87	1.35	0.67	0.72
La	33.68	26.11	16.23	16.26	11.97	10.85	12.41
Ce	54.27	53.24	35.46	35.54	28.90	22.69	26.89
Pr	12.74	6.27	4.35	4.36	4.06	3.33	3.75
Nd	56.79	23.50	17.13	17.17	18.88	14.73	16.52
Sm	15.42	4.57	3.73	3.74	5.36	3.73	3.92
Eu	4.58	0.99	0.98	0.99	1.60	1.51	1.25
Gd	16.13	4.09	3.60	3.62	5.73	4.05	4.20
Tb	2.49	0.48	0.51	0.51	0.80	0.62	0.61
Dy	15.98	2.62	3.10	3.11	4.59	4.26	3.96
Но	2.83	0.46	0.55	0.55	0.76	0.83	0.75
Er	8.18	1.35	1.56	1.57	2.04	2.55	2.24
Tm	1.01	0.17	0.20	0.20	0.25	0.34	0.29
Yb	6.72	1.14	1.36	1.37	1.65	2.37	2.02
Lu	0.97	0.16	0.19	0.19	0.23	0.35	0.30
Sc	24.87	5.44	8.60	6.07	10.21	16.09	14.79
V	167.16	44.86	49.58	41.61	80.78	55.11	98.61
Cr	17.24	71.72	10.90	103.87	121.76	-5.11	-3.74
Co	20.58	12.45	8.02	11.80	26.02	5.35	7.44
Ni	44.81	79.20	4.15	135.11	190.35	-1.57	10.00
Pb	9.27	13.21	12.30	12.31	8.12	7.36	14.30
Cs	0.20	3.36	2.85	2.85	1.85	0.73	0.89
Th/Sc	0.10	1.59	0.65	0.91	0.30	0.13	0.15
Zr/Sc	2.61	3.01	2.82	3.97	2.34	3.82	4.10
La/Th	13.03	3.02	2.92	2.93	3.91	5.12	5.45
Th/Yb	0.38	7.57	4.07	4.07	1.85	0.90	1.13
⁸⁷ Sr/ ⁸⁶ Sr	0.70657	0.71002	0.70665	0.70693	0.70800	0.70551	0.70683
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512851	0.512344	0.512708	0.512502	0.512500	0.512835	0.512823
$\varepsilon_{\rm Nd}(0)$	4.2	-5.7	1.4	-2.7	-2.7	3.8	3.6
T _{DM} (Ga)	0.70	1.13	0.69	1.04	1.86	0.62	0.57
- DM (Cm)	0.70		0.07			0.02	0.07

Table 4.3: Trace element concentration and isotopic ratio data for siliciclastic sedimentary rocks of the Mithakhari Group

The trace elements concentrations are given in 'ppm'. Sr and Nd isotopes are given as ratios ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd. As variations in ¹⁴³Nd/¹⁴⁴Nd ratios are very less, to magnify the variation it is generally,

represented by parameter ε_{Nd} . This parameter ε_{Nd} at present is defined by: $\varepsilon_{Nd}(0) = \{ [(^{143}Nd/^{144}Nd)_{S} - (^{143}Nd/^{144}Nd)_{CHUR}]/(^{143}Nd/^{144}Nd)_{CHUR} \} \times 10^{4}$ where subscipts 's' stands for sample and 'CHUR' stands for Chondrite Uniform Reservoir. The present day $^{143}Nd/^{144}Nd$ value of CHUR is 0.512638 (Depaolo and Wasserburg, 1976). In the results the parameter ε_{Nd} is taken for the present value of CHUR, so the subscript "(0)" has been omitted. $T_{DM} = 1/\lambda \{1 + [(^{143}Nd)^{144}Nd)_{S} - (^{143}Nd)^{144}Nd)_{DM}]/[(^{147}Sm/^{144}Nd)_{S} - (^{147}Sm/^{144}Nd)_{DM}]\}$ where subscripts 'DM' stands for depleted mantle, ($^{143}Nd/^{144}Nd)_{DM} = 0.513114$ and ($^{147}Sm/^{144}Nd)_{DM} = 0.513114$

0.222 (Michard et al., 1985).

	AND-09-24	AND-09-27	AND-09-30	AND-09-37	AND-09-41	AND-09-44	AND-09-48
(ppm)							
Rb	27.00	89.24	25.03	22.05	13.42	76.62	92.89
Sr	127.25	36.59	153.85	101.54	552.49	56.08	92.03
Ba	55.14	83.36	417.42	70.54	142.26	114.97	147.88
Y	16.65	13.56	31.99	11.55	16.08	16.48	10.12
Zr	26.07	68.35	144.96	17.67	13.65	44.75	66.15
Hf	1.25	2.23	4.10	1.03	0.67	1.78	2.53
Nb	4.33	6.81	5.88	3.88	2.91	6.45	10.37
Та	0.27	0.42	0.36	0.25	0.17	0.41	0.63
Th	3.32	7.12	5.06	3.08	1.16	5.85	6.44
U	0.73	2.21	1.32	0.67	0.16	1.28	1.27
La	19.10	22.98	17.45	14.68	21.06	19.43	18.08
Ce	36.99	43.70	39.38	30.17	22.71	43.82	46.79
Pr	4.60	4.86	5.37	3.82	4.57	5.55	5.40
Nd	18.54	17.34	23.15	15.42	18.31	22.93	22.02
Sm	4.20	3.24	5.74	3.44	3.65	5.30	4.91
Eu	1.21	0.64	1.53	0.99	0.93	1.32	1.17
Gd	4.57	3.05	6.11	3.60	3.98	5.27	4.62
Tb	0.63	0.39	0.92	0.51	0.54	0.72	0.62
Dy	3.97	2.61	6.14	2.98	3.39	4.23	3.62
Ho	0.72	0.51	1.19	0.50	0.63	0.74	0.62
Er	2.03	1.67	3.65	1.32	1.83	2.07	1.74
Tm	0.25	0.24	0.48	0.16	0.23	0.25	0.22
Yb	1.69	1.79	3.38	1.04	1.45	1.70	1.48
Lu	0.23	0.27	0.51	0.14	0.19	0.23	0.21
Sc	6.93	6.59	20.35	4.45	0.67	10.89	12.98
V	54.26	52.67	110.21	24.43	11.05	84.21	110.90
Cr	64.45	32.98	112.67	43.26	9.10	105.90	55.11
Со	8.33	7.02	14.13	6.26	2.05	15.41	21.28
Ni	45.72	35.78	104.93	44.50	30.65	160.35	69.67
Pb	10.37	14.35	10.52	8.03	18.73	11.58	16.14
Cs	1.32	5.31	3.41	1.04	0.57	5.87	5.70
Th/Sc	0.48	1.08	0.25	0.69	1.72	0.54	0.50
Zr/Sc	3.76	10.37	7.12	3.97	20.24	4.11	5.10
La/Th	5.76	3.23	3.45	4.77	18.12	3.32	2.81
Th/Yb	1.96	3.97	1.50	2.95	0.80	3.43	4.35
⁸⁷ Sr/ ⁸⁶ Sr	0.70749	0.71196	0.70576	0.70703	0.70732	0.70952	0.71021
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512551	0.512630	0.512729	0.512510	0.512169	0.512547	0.512371
ε _{Nd} (0)	-1.7	-0.2	1.8	-2.5	-9.1	-1.8	-5.2
T _{DM} (Ga)	1.01	0.68	0.82	1.06	1.42	1.05	1.30

Table 4.3: Continued

	AND-09-49	AND-09-52	AND-09-54	AND-09-55	PB-08-08	PB-08-11	PB-08-13
(ppm)							
Rb	46.61	22.93	22.9	12.24			34.07
Sr	40.74	152.28	152.3	101.35			75.16
Ba	205.79	285.61	285.6	210.45			62.99
Y	17.20	19.32	19.3	16.85			11.27
Zr	86.11	73.52	73.5	43.30			32.21
Hf	2.88	2.82	2.8	1.68			1.49
Nb	6.81	4.13	4.1	3.26			4.27
Та	0.47	0.24	0.2	0.19			0.33
Th	5.96	2.47	2.5	1.73			3.77
U	1.07	0.80	0.8	0.49			0.90
La	14.96	11.32	11.3	10.03			14.05
Ce	38.85	25.70	25.7	20.73			29.79
Pr	4.47	3.52	3.5	2.98			3.77
Nd	18.52	15.46	15.5	12.86			15.15
Sm	4.23	3.94	3.9	3.20			3.30
Eu	1.07	1.44	1.4	1.11			0.98
Gd	4.18	4.26	4.3	3.42			3.35
Tb	0.59	0.67	0.7	0.52			0.46
Dy	3.88	4.63	4.6	3.48			2.97
Но	0.74	0.91	0.9	0.67			0.54
Er	2.29	2.86	2.9	2.05			1.62
Tm	0.31	0.39	0.4	0.28			0.21
Yb	2.20	2.84	2.8	1.92			1.41
Lu	0.32	0.44	0.4	0.28			0.20
Sc	17.91	18.26	18.3	17.29			5.79
V	89.67	87.34	87.3	79.07			41.82
Cr	54.88	9.54	9.5	69.00			115.20
Co	20.93	11.38	11.4	12.05			12.18
Ni	100.18	4.92	4.9	40.93			118.91
Pb	8.03	10.23	10.2	6.74			9.08
Cs	3.68	3.00	3.0	0.71			2.10
Th/Sc	0.33	0.14	0.14	0.10			0.65
Zr/Sc	4.81	4.03	4.03	2.50			5.56
La/Th	2.51	4.58	4.58	5.79			3.72
Th/Yb	2.71	0.87	0.87	0.90			2.67
⁸⁷ Sr/ ⁸⁶ Sr	0.71255	0.70856	0.70497	0.70554	0.71092	0.71130	0.70657
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512564	0.512667	0.512833	0.512796	0.512237	0.512278	0.512604
ε _{Nd} (0)	-1.4	0.6	3.8	3.1	-7.8	-7.0	-0.7
T _{DM} (Ga)	1.00		0.64	0.68			0.86

Table 4.3: Continued

	AND-09-04	AND-09-12	AND-09-14	AND-09-36	AND-09-43	AND-09-61	PB-08-09
(ppm)							
Rb	89.68	75.09	81.26	90.63		90.51	
Sr	78.16	17.24	27.29	73.37		66.78	
Ba	233.74	272.72	175.35	312.93		271.06	
Y	12.56	8.61	8.79	10.55		17.75	
Zr	22.11	27.46	58.69	21.47		31.37	
Hf	1.26	1.32	2.00	1.08		1.36	
Nb	8.96	7.56	6.50	6.67		9.20	
Та	0.61	0.50	0.43	0.55		0.67	
Th	10.62	12.35	7.09	7.58		9.90	
U	1.65	1.49	1.49	1.11		1.41	
La	29.91	26.26	20.64	22.81		38.33	
Ce	63.59	55.74	43.29	47.57		85.41	
Pr	7.53	6.58	5.16	5.79		10.37	
Nd	29.06	24.83	18.60	22.48		40.11	
Sm	5.97	4.78	3.06	4.57		8.82	
Eu	1.30	0.97	0.69	1.10		1.97	
Gd	5.45	4.11	2.75	4.11		7.73	
Tb	0.67	0.47	0.34	0.50		1.00	
Dy	3.73	2.55	2.18	2.80		5.55	
Ho	0.63	0.44	0.44	0.49		0.90	
Er	1.76	1.26	1.42	1.36		2.51	
Tm	0.22	0.16	0.20	0.17		0.31	
Yb	1.48	1.07	1.49	1.17		2.07	
Lu	0.19	0.14	0.22	0.15		0.28	
Sc	6.26	1.57	7.05	6.24		6.85	
V	49.19	15.10	90.08	48.40		59.00	
Cr	35.24	0.00	28.04	40.14		41.44	
Со	10.09	2.67	1.12	7.45		20.47	
Ni	51.17	11.58	33.49	51.99		71.95	
Pb	20.09	13.24	6.17	12.71		17.84	
Cs	5.18	3.41	5.04	4.80		4.49	
Th/Sc	1.70	7.89	1.01	1.21		1.45	
Zr/Sc	3.53	17.54	8.32	3.44		4.58	
La/Th	2.82	2.13	2.91	3.01		3.87	
Th/Yb	7.20	11.52	4.76	6.48		4.78	
⁸⁷ Sr/ ⁸⁶ Sr	0.71548	0.73049	0.71554	0.71837	0.71605	0.72323	0.72105
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512142	0.511987	0.512473	0.512157	0.512059	0.512089	0.512007
ε _{Nd} (0)	-9.7	-12.7	-3.2	-9.4	-11.3	-10.7	-12.3
T _{DM} (Ga)	1.52	1.63	0.80	1.47		1.75	

Table 4.4: Trace element concentration and isotopic ratio data for siliciclastic sedimentary rocks of the Andaman Flysch Group


Fig. 4.4: Major element oxide versus SiO_2 for samples from the Mithakhari and the Andaman Flysch Groups. Values for PAAS are from Taylor and McLennan (1985).



Fig. 4.5: Primitive Mantle (PM) normalized multi element spidergrams (a&b) and chondrite normalized REE patterns (c&d) for the Mithakhari Group and the Andaman Flysch Group rocks. Data Sources: PM- McDonough and Sun (1995); Chondrite- Sun and McDonough, (1989); Post Archean Average Australian shale (PAAS)-Taylor and McLennan (1985); compositions of Barail Group sediments (average)- Hossain et al., (2010); Andaman Ophiolite/Arc field- Kumar, (2011); Pal et al., (2011); Pederson et al., (2010); Himalayan field- Ahmad et al., (2000); Islam et al., (2011); Miller et al., (2001) and Sachan et al., (2010).

- 6. The primitive mantle-normalized trace element patterns for Andaman siliciclastics (Fig. 4.5a & b) are similar to PAAS, however, show more pronounced depletions in Ba, Nb, Ta, Sr, and P. The amounts of these depletions are much higher in rocks of the Mithakhari Group compared to those of the Andaman Flysch Group (Fig. 4.5a & b).
- 7. The contents of Large Ion Lithophile Elements (LILE) in the Andaman Flysch rocks are generally higher than that in the Mithakhari rocks and plot closer to the pattern of PAAS (Fig. 4.5b).
- 8. The trace element patterns of the Mithakhari rocks overlap with the field of the Andaman Ophiolites, whereas those of the Andaman Flysch rocks plot well within the field defined for the Himalaya (Fig. 4.5a & b).
- 9. The chondrite-normalized REE patterns (Fig. 4.5c & d) for samples from the Mithakhari Group show fairly uniform, moderately enriched light REE patterns ($La_N/Sm_N= 1.4-3.6$), fairly flat HREE patterns ($Gd_N/Yb_N= 1.4-2.9$) and moderate negative Eu anomalies (Eu/Eu* = 0.6-1.2) whereas those from the Andaman Flysch rocks, although, have similar features, have much higher light REE enrichment and much closer to PAAS.
- 10. Like other trace elements, the REE patterns for the Mithakhari rocks overlap with the field of the Andaman Ophiolites, whereas the Andaman Flysch rocks plot well within the field of Himalayan rocks (Fig. 4.5a & b).

(C) Radiogenic isotope geochemistry

The ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd isotopic ratios were measured on selected whole rock samples from both the Mithakhari and Andaman Flysch groups. These data are presented in Tables 4.3 and 4.4 and in histograms in Fig. 4.6. Since the Mithakhari Group rocks are better exposed and easier to access, our database is somewhat biased towards this group. Although the isotopic ratio variations in these groups

overlap with each other, there exists a clear difference between the two with the Mithakhari rocks showing lower 87 Sr/ 86 Sr and higher ϵ_{Nd} (Tables 4.3 & 4.4, Fig. 4.6). The range of 87 Sr/ 86 Sr variations in the Mithakhari and Andaman Flysch groups are 0.70497 to 0.71554 and 0.71548 to 0.73049, whereas for ϵ_{Nd} these are -8.1 to +4.2 and -12.7 to -9.4, respectively. Isotopic data on these rocks from the work of Allen et al. (2007) fall well within our observed variations.



Fig. 4.6: Stacked histograms of ${}^{87}Sr/{}^{86}Sr$ and \mathcal{E}_{Nd} (0) distributions in the sedimentary rocks of the Andaman Islands. Also shown are the ranges for Andaman Ophiolites.

The ¹⁴³Nd/¹⁴⁴Nd isotopic ratios are used for calculation of model ages for the sedimentary rocks with respect to depleted mantle. These ages represent timings of mantle extraction of the original igneous source rock from which the sediments were derived. However, because of multiple source contributions and sediment recycling, the calculated model ages of sedimentary rocks often do not reveal the true ages of the source rocks. Usually, the ages of older terrigenous sediments decrease on getting mixed with younger sources (Faure, 2005). These ages, however, can become very helpful in understanding the provenances when used in conjunction with other tracers. A histogram for the calculated depleted mantle ages (T_{DM}) for the Mithakhari Group and the Andaman Flysch Group rocks are shown in Fig. 4.7. The T_{DM} values for the Mithakhari Group are older. Our database, though not exhaustive, shows a prominent mode at ~1.1 Ga and a less pronounced mode at ~0.7 Ga (Fig. 4.7).



Fig. 4.7: Stacked histogram showing frequency distributions of T_{DM} ages for the sedimentary sequences of the Andaman Islands.

4.4 Discussion and Implications

4.4.1 Weathering and Sedimentation

Al₂O₃-CaO*+Na₂O-K₂O (A–CN–K) plot and Chemical Index of Alteration (CIA) together can be used for quantification of weathering and diagenetic history of a sedimentary rock in addition to deciphering climatic conditions during their

deposition (Nesbitt and Young, 1982, 1984). As the intensity of weathering increases, the concentration of Ca, Na, and K decreases in the bulk-rock chemistry due to alteration of feldspars (and glass for volcanic rocks) into clay minerals (e.g., Duzgoren-Aydin et al., 2002; McLennan, 1993) and therefore, CIA value increases. In Fig. 4.8, all our data points plot parallel to the A-CN line suggesting variable degree of weathering of the source rocks. The CIA values for the Mithakhari Group are comparatively lower than the Andaman Flysch Group. This hint at comparatively less weathering in the sources of the Mithakhari sediments while high CIA values for the Andaman Flysch suggests high degree of weathering in their sources. Figure 4.8 also reveals out that the sediments of the Mithakhari Group were derived from predominantly mafic igneous sources whereas plagioclase was dominant mineral and that they did not get transported long distance before deposition- which is reflected in their smectite rich clay content. In contrast, the sediments of the Andaman Flysch appear to have been derived from mixed sources with dominance of felsic igneous sources. The presence of higher amount of illite suggests that the sediments would have been transported a large distance before being deposited.



Fig. 4.8: Al_2O_3 -(CaO*+ Na_2O)- K_2O diagram with the Chemical Index of Alteration (CIA) of Nesbitt et al. (1996) on which samples from the Andaman sedimentary rocks are plotted. Also shown on the diagram are the positions of characterstic minerals, post-Archean Australian Average Shale (PAAS) (Taylor and McLennan, 1985), and average Archean upper crust (Taylor and McLennan, 1985). The solid and dashed lines with CIA values for shales have CIA values to the range of 70-75. Arrows emanating from plagioclase-Kfeldspar join show the weathering trends for basalt, tonalite, granodiorite and granite (Nesbitt and Young, 1984, 1989).

Th/Sc is a good indicator of igneous differentiation process and Zr/Sc ratio gives a measure of mineral sorting and recycling (Hassan et al., 1999), therefore, a plot of Th/Sc versus Zr/Sc in sediments can provide clues about the nature of their sources and processes of removal of material out of them (McLennan, 1993). In such a plot (Fig. 4.9) the Mithakhari Group rocks show low Th/Sc and Zr/Sc ratio suggesting that sediment have not undergone much recycling and sorting. In comparison, the Andaman Flysch Group rocks have relatively higher Th/Sc and Zr/Sc ratios with values above the average value of the upper continental crust (Fig 4.9) suggesting that they have undergone substantial recycling and sorting, and have concentrated zircons during transport.



Fig. 4.9: Plot of Th/Sc versus Zr/Sc for Andaman Island samples. Also plotted are ratios for upper crust (Taylor and McLennan, 1985) and Andaman Trench sediments (Planck and Langmuir, 1998).



Fig. 4.10: (a) K_2O/Na_2O versus SiO_2 discrimination diagram of Roser and Korsch (1986) for sediment of the Mithakhari Group and the Andaman Flysch Group. Fields shown are for passive continental margin (PM), an active continental margin (ACM) and island arc (IA) (b) La-Th-Sc discrimination diagram of Bhatia and Crook, (1986) for the same sediments. Fields shown are for oceanic island arc (OIA), continental island arc (CIA), passive continental margin (PM) and active continental margin (ACM). Values for PAAS are from Taylor and McLennan, (1985).

The chemical composition of sedimentary rocks show close link to tectonic environments and provenance as long as it is not disturbed/modified during post-

depositional processes. The tectonic and sedimentary processes impart a distinctive geochemical signature to sediments and different tectonic environments have distinctive provenance characteristics (Bhatia and Crook, 1986). In two such tectonic discrimination plots of K_2O/Na_2O versus SiO₂ and La-Th-Sc ternary diagram, the data from the Andaman sedimentary formations predominantly fall in the fields of island arc and active continental margin (Fig. 4.10). While the Mithakhari Group rocks, barring a few, show chemical signatures of island arc setting, the Andaman Flysch rocks clearly were deposited in a predominantly active continental margin setting (Fig. 4.10).



Fig. 4.11: La/Th and Th/Yb variations in sandstones and shales from the Mithakhari Group and the Andaman Flysch Group. The Andaman Ophiolite field is drawn using data from Pal et al. (2011) and Pederson et al. (2010) while Himalayan field is drawn using those from Ahmad et al. (2000); Islam et al. (2011); Miller et al. (2001) and Sachan et al. (2010); Andaman Trench sediments-Planck and Langmuir (1998).

4.4.2 Provenance

Chemical compositions of siliciclastic sediments have been widely used as indicators of provenances (e.g., McLennan et al., 1995, 2003). The elemental abundances (and ratios) of relatively immobile elements (REE, Th, Nb, Sc & Zr) remain unchanged during weathering and diagenesis processes. Therefore, these elements get transferred quantitatively from parent rocks to the clastic sediments

produced from them (Taylor and McLennan, 1985; Condie, 1991). In the primitive mantle-normalized multi element spidergrams and chondrite-normalized REE plots (Fig. 4.5) the Mithakhari Group samples show larger contribution of sediments derived from the suprasubduction ophiolites of the Andaman Subduction Zone while the Andaman Flysch sediments with their relatively enriched LREE patterns appear to have been derived primarily from the Himalayan rocks, having crustal signatures.

Because of their relative incompatibility during differentiation of mafic and felsic igneous rocks, the concentration of elements La, Th and Sc vary widely in various magmatic rocks and therefore, can become very useful in determining original igneous sources for sediments. The ratios like La/Th, La/Sc, Th/Yb, Sm/Nd and Th/Sc have been successfully used as tracers of provenance (McLennan et al., 1980; Taylor and McLennan, 1985; Wang et al., 1986; McLennan and Hemming, 1992). In a bivariate plot between La/Th versus Th/Yb (Fig. 4.11) sedimentary rocks from the Andamans plot in-between the fields for the Ophiolite Group of the Andamans and rocks of the Himalayan mountain belt, which suggest that these sediments are a mixture of materials derived from these two sources. As observed in the A-CN-K plot (Fig. 4.8), the majority of the Mithakhari Group formations follow a trend that suggest their derivation from mafic igneous sources (basaltic), while rocks from the Andaman Flysch Group and a few sandstone and shale formations of the Mithakhari Group follow a granodiorite trend, suggesting dominance of continental/felsic material over in their sources.

 143 Nd/ 144 Nd is a robust tracer for determining sediment source in comparison to 87 Sr/ 86 Sr because it is not susceptible to alteration during weathering and diagenesis (Patchett, 2003). In addition, the parent-daughter ratio of Sm-Nd systematics is highly resistant during diagenesis unlike mobile Rb and Sr. However, considering that the rocks of the Andaman Islands are young (< 100 Ma) and that the samples studied for isotopic composition are devoid of carbonate cements, we believe that their 87 Sr/ 86 Sr too can help us trace the sediment sources.



Fig. 4.12: (a) Plot of $\mathcal{E}_{Nd}(0)$ ranges observed in the Mithakhari Group and the Andaman Flysch Group compared with data from possible source regions (as given in Allen et al., 2007; Allen et al., 2008; Najman et al., 2008). (b) Plot of $\mathcal{E}_{Nd}(0)$ versus ⁸⁷Sr/⁸⁶Sr for our samples and sediments from the Barail Group of Bengal Basin compared with two component mixing model curves that assume three different end-member pairs for Arc and Continental sources. The tick marks on the mixing curves represent fraction of Arc source material in the mixture. Data sources: Andaman Island ophiolites: Kumar (2011), Barail Group sediment: Najman et al., (2008). All data were normalized to NBS987 (0.710250) and La Jolla (0.511858) for ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd isotopic ratios, respectively.

Comparing the observed ε_{Nd} variations in Andaman sedimentary rocks with that of the possible sources (Fig. 4.12a) we make the following observations. The Mithakhari Group rocks show large variation in ε_{Nd} which supports our earlier inference that their sediments have variable mixing from arc and continental sources. We also observe that there is a progressive lowering of ε_{Nd} with younging, which ends up with more negative values for the Andaman Flysch rocks; suggesting predominant contributions from continental sources in younger rocks. To quantify the amounts of contributions from these above two sources (i.e. magmatic Arc/ophilitic and continental crust) we utilized both Sr and Nd isotopic ratios.

In a ε_{Nd} versus ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ diagram (Fig. 4.12b) we plot our data and compare them with the potential sources: the rocks of the Andaman Ophiolite Group and the magmatic arc, and rocks of the upper continental sources (e.g., in our case these could have been the rocks of the Indian Shield and the Himalayas). As can be seen in the Fig. 4.12b the rocks of the Mithakhari Group, with their higher ϵ_{Nd} and lower 87 Sr/ 86 Sr, plot closer to the magmatic arc sources, whereas the rocks of the Andaman Flysch Group, with their lower ϵ_{Nd} and higher $^{87}\mathrm{Sr}/^{86}\mathrm{Sr}$ plot away from it but are closer to values for upper continental crustal material. All the sedimentary rocks appear to show a mixing trend (hyperbola) between the above two end member sources. Considering that the most probable upper continental crustal end-member in this case is the sources in the Himalayas and the magmatic arc member is local (Andaman ophiolites), we draw simple binary mixing curves, using three different end-member pairs, to explain the observed data. The results of mixing calculations reveal that the local arc sources possibly contributed >80% of sediments to the Mithakhari Group, whereas the same sources contributed about 60-80 % to the Andaman Flysch. The maximum amount of continental contribution to the rocks of the Andaman forearc basin could not have exceeded 30% of the total. Interestingly, we also find that the sediments of the Barail Group in the Bengal Basin too had received significant amount of sediments ($\leq 30\%$) from the arc sources, probably those located in Myanmar.

Presence of mafic-ultramafic clasts in the basal conglomerate and interbedded tuff deposits, and results of our geochemical study all indicate that the local ophiolites along with volcanic arc acted as the major sources for sediments to the Mithakhari Group. The suprasubduction ophiolites present in the Andamans were probably subaerially exposed during the early to middle Eocene and had undergone substantial erosion. This process produced clasts with weakly weathered coarse sediments which were locally deposited within the trench-forearc basin. The tectonic plate configuration at this time possibly allowed transport of additional sediments from the Himalayan/Indian Shield sources into this basin.

Results from petrography (this study; Pal et al., 2003; Allen et al., 2007), geochemical and isotopic studies indicate that distant felsic continental rocks acted as the main sources of sediments to the Andaman Flysch Group. Presence of Archean, Proterozoic, and Paleozoic age zircons corroborates this (Allen et al., 2007). These sediments apparently have a complex weathering and transportation history. All evidences support our inferences that these sources were located either in the rising Himalayan mountain belt or in the eastern Myanmar region. Although, it is difficult to discriminate between the two, the absence of major drainage system in eastern Myanmar, capable of transporting large amounts of sediments led us to believe that the Himalayan- Transhimalayan sources were the primary provenance for the Oligocene forearc deposits.

4.4.3 Evolution of the Andaman region

Based on the inferences made above on the provenance of the Andaman sediments and our understanding of the paleogeography and paleodrainage systems that might have existed in the continents surrounding the forearc during the Paleogene, we propose certain tectonic configurations of SE Asia at ~50 Ma and ~30 Ma in Fig. 4.13. We believe that the terminal phase of continental collision between the Indian and Eurasian plates happened at ~50 Ma (Rowley, 1996; Hodges, 2000). A remnant of the Neo-Tethys Sea probably existed at this time north of the Himalayan suture zone (Fig. 4.13). On further movement of the Indian plate to the north during the Early Eocene, this narrow sea might have started closing from the west. The Trans-Himalayan-Sunda magmatic arc existed parallel to the collision-subduction



Fig. 4.13: Schematic maps showing the Paleogene configuration of the Indian Plate and other tectonic blocks of SE Asia at 50 Ma and 30 Ma (modified after Replumaz and Tapponnier, 2003). The Greater India moved northward throughout the Paleogene with crustal shortening at its northern margin. The sketch also illustrates the location of volcanic arc (triangles) and continental sources and paleodrainage directions (arrows) to the forearc and foreland basins.

boundary. The first ~12 Myr of post-collisional sedimentary record (i.e. until about 38 Ma) from the foreland basin, Bengal basin and the Andaman forearc basin show negligible input of sediments from continental sources but have substantial contributions from magmatic arc/suture zone sources (Najman and Garzanti, 2000; Najman et al, 2008). Figure 4.13a illustrates the paleogeographic configuration at ~50 Ma and shows the direction of sediment transportation from arc/suture zone sources into the Bengal basin and Andaman forearc through paleodrainage systems. The dominance of arc/suture zone sources suggests that until that time, the Himalayas had not attained the critical height for large scale weathering. In addition, the SW monsoon system did not exist, which otherwise would have aided the erosion and transportation of sediments from the Indian Shield and nascent Himalaya.

We believe that at ~ 40 Ma, major thrusting events occurred along the northern continental margin of the Indian plate. This probably provided sufficient topographic barrier to the moisture from the south and resulted in the development of the first monsoon system. The monsoon enhanced the weathering and erosion of the exhumed continental crust and resulted in exponential increase in the sediment input to the adjacent basins at the start of the Oligocene (Metivier et al., 2002). From the Oligocene onwards the major input of Himalayan detritus was deposited in the Bengal Basin and also in the Andaman forearc basin. The volume of sediments deposited suggests that the exhumation and erosion of the Himalaya that occurred during the Oligocene was rather more intense and covered a larger spatial extent than previously thought. This suggests that the some major river systems had already developed and was supplying sediments to these basins from the north or northeast and west. We believe that these sediments were supplied by the paleodrainage system of the Ganga draining into the foreland basin through the southern slopes of the rising Himalaya and that of Yarlung-Tsangpo River (Tibet) developed in the Himalayan arc and suture zone with the closing of the Neo-Tethys (Fig. 4.13b). The now exposed Indo-Burman ranges and Andamans were part of the same forearc basin. With continued subduction in its western part, these trenchforearc deposits of the Indo-Burman block started uplifting during the Oligocene to form the Indo-Burman Ranges (Mitchell, 1993). This upliftment of Indo-Burman

Ranges must have acted as a barrier to the drainage system and divided it into two parts- the Brahmaputra River system carrying sediments to the Indian plate and Irrawaddy River system on the Burmese plate. The Ganga and Brahmaputra river systems supplied sediments to the Bengal basin and Bay of Bengal while the Irrawaddy river system supplied sediments to basins to the further east.

Tectonic activities continued the upliftment of trench-forearc sediments which progressed from the north to the south. This first exposed the Indo-Burman Ranges and later the (islands of) Andamans. A major episode of uplift at ~20 Ma, believed to be a result of major tectonic events (Allen et al., 2007; Mountain and Prell, 1990; Acharyya et al., 1989), terminated the deposition of the Andaman Flysch Group and probably shifted the centre of deposition further west to the Bay of Bengal. To the east of the raised ophiolite and trench-forearc sequences deposition of the Archipelago Group sediments started in deep marine conditions (Pal et al., 2005; Singh et al., 2000). Allen et al. (2007) suggested that the present topography of the Andaman Islands is resulted from a major regional upliftment event that occurred during ~ 10 to 5 Ma. The subduction continued in the west of the Andaman Islands and the Andaman Sea opened in the east during the Late Miocene to Pliocene (Curray, 2005).

4.5 Provenance of sediments of the Andaman Sea

4.5.1 Background and Earlier work

The Andaman Sea receives more than 360 million tonnes of sediment from the surrounding landmasses annually (Meade, 1996; Milliman and Meade, 1983). Earlier studies have shown that a large portion of this sediment supply is delivered by the tropical/subtropical river systems of Myanmar; notably the Irrawaddy, Sittang and Salween with the first being the dominant of all (Rodolfo, 1969) (Fig. 4.14). The Irrawaddy River is the fifth largest in the world in terms of sediment discharge (Rao et al., 2005). The reason for such a high influx of sediment is higher monsoonal rainfall and resultant erosion in the drainage basins. Along with the Indian subcontinent, Myanmar is affected by South Asian Monsoon, which comprises the South West (summer) and the North East (winter) monsoons. Both these monsoons play an important role in the erosion in the catchments of these rivers. The sediment

load in the Irrawaddy River system is derived from: the Himalayas in far north, Indo-Burman mountain ranges in the west and Shan Plateau in the east (Fig. 4.14) (Robinson et al., 2007). Although the exact amounts of sediment loads contributed by each of these sources are unknown, the study by Allen et al. (2007) suggests these to be a complex mixture of sediments derived from multiple sources. In Table 4.1, we describe important physical and chemical characteristics of these sediment sources. The Indo-Burman sources are similar to those occur on the Andaman Islands and mainly are ophiolites, volcaniclastics and siliciclastic sedimentary rocks of Cretaceous to Oligocene age (Allen et al., 2008; Colin, 2006).

The Andaman Sea is characterized by a seasonally reversing surface circulation pattern linked to the monsoons. The sediment load brought in by the rivers is supplied to it from the north, of which most get deposited on the broad continental shelf in the gulf of Martaban while rest are carried to the east and south along the continental shelf of Myanmar and Thailand by the coastal currents driven by SW monsoonal winds (Rodolfo, 1969). Only a small part of the sediment supplied finally reaches to the deeper parts of the Andaman Sea. These coastal currents reverse their direction during the NE monsoon and carry sediments westward towards the Bay of Bengal (Ramaswamy et al., 2004; Rao et al., 2005) (Fig. 4.15). Apart from the major sediment contribution from north, there is also some minor contribution from Malay Peninsula, however, its deposition is limited only to the inner shelf (Rodolfo, 1969). The sediments deposited in the south and south-east Andaman Sea, which are probably derived from sources in Malay Peninsula and Sumatra, are carried by the Malacca currents into the Bay of Bengal (Rodolfo, 1969) (Fig. 4.15).

The intensity of the monsoons has been known to have varied in millennial scale during the late Pleistocene and Holocene (Duplessy, 1982; Herzschuh, 2006; Prell and Kutzbach, 1987; Sarkar et al., 1990; Tiwari et al., 2005). Such fluctuations are mainly controlled by the low latitude solar insolation, which varied with precession and eccentricity of the Earth's orbit (Clemens and Prell, 1990; Clemens et al., 1991; Colin et al., 1998; Duplessy, 1982). Several studies have shown that past variations



Fig. 4.14: Map showing (sediment) source regions and major rivers of Myanmar (source: <u>http://en.wikipedia.org/wiki/File: Myanmar_relief_location_map.jpg</u>).



Fig. 4.15: Map of the Andaman Sea (source: <u>http://www.internalwaveatlas.</u> <u>com/Atlas PDF/IWAtlas Pg207_AndamanSea.PDF</u>) showing bathymetry and location of sediment cores SK-234-60, SK 168/GC-1 MD77-169, MD77-171 and MD77-176. Blue arrows are directions of North-East (winter) monsoon currents while orange arrows are South-West (summer) monsoon currents (Rodolfo, 1969), Thick yellow arrows show direction of sediments carried by currents.

in the South-East Asian monsoonal intensity have influenced erosion and caused major changes in the sediment supply (e.g., Tripathy et al., 2011). These also affected sea-water circulation and sediment dispersal pattern in the Indian Ocean (Hashimi et al., 1995; Goodbred and Kuehl, 2000; Colin et al., 1999). Therefore, it is expected that temporal changes in monsoon intensity in the past could have brought major changes to the weathering-erosion pattern in the source regions (of Myanmar) for the sediments to the Andaman Sea. Unfortunately, due to lack of

detailed studies in the S-E Asia, we know very little about how sediment contribution to the Andaman Sea varied with time as a result of climatic fluctuations. In this work, we have made an effort to understand the relationship between variations in provenance of terrigenous sediments to the Andaman Sea and monsoon aided erosion in the source regions in the South-East Asia and Indian shield, during the late Pleistocene and Holocene, with the help of geochemical and isotopic tracers in a sediment core collected from the Andaman Sea.

4.5.2 Results

The sediment core (SK-234-60) which was raised from the western Andaman Sea from the location N12°16'40", E93 ° 51'30" to study the history of volcanism of Barren Island was utilized for this work as well. The sampling and chronological details of the core have already been discussed in Chapter-2 and Chapter-3, respectively. The geochemical and Sr-Nd isotopic data for the carbonate free siliciclastic sediments are presented in Table 4.5.

(A) C-14 ages and sedimentation rates

The AMS C-14 ages obtained for different sediment layers in the core (Tables 3.1) were used for calculation of sedimentation rates (Fig. 3.3). Our calculations yielded the following rates of sedimentation: 6.1 ± 0.4 cm/kyr during 2.2-3.8 ka, 7.3 ± 0.7 cm/kyr during 3.8-5.2 ka, 8.3 ± 0.8 cm/kyr during 5.2-6.4 ka, 7.5 ± 0.4 cm/kyr during 6.4-8.0 ka, 3.7 ± 0.1 cm/kyr during 8.0-12.3 ka, 3.0 ± 0.1 cm/kyr during 12.3-17.3 ka, 2.4 ± 0.1 cm/kyr during 17.3-23.5 ka, 3.5 ± 0.1 cm/kyr during 23.5-34.4 ka, and 5.5 ± 0.7 cm/kyr prior to 39.5 ka (Fig. 3.4). The errors, at 1 σ level, on the above rates were determined by propagating the errors on the ages. The average sedimentation rate in the core estimated to be 5.3 cm/kyr, is much lower than earlier estimates of average sedimentation rate in the Andaman Sea: 15 cm/kyr (Frerichs, 1968), 10cm/kyr (Core MD77-169 from the Sewell seamount region; Colin et al., 2006) and 7.8 cm/kyr (from Alcock Seamount region; Sijinkumar et al., 2010) (Fig. 4.15).

The lower sedimentation rate at our core site could be due to the unusual local bathymetry and easterly flowing surface currents that disperse sediments more towards eastern continental margin than to straight south (Fig. 4.15). Linear

extrapolation of calculated rate of sedimentation from bottommost dated band gives the age of bottom of the core to be \sim 74 ka (Fig 3.4). The depth-age model for the core is used to infer about the temporal variation of the sources supplying sediments to the core location.

(B) Geochemical Data

Based on major and trace element contents of the siliciclastic sediments from various layers of the core we make the following observations.

- 1. The SiO₂ contents of the analysed sediments varied between 48 and 61% and Al_2O_3 concentrations were always >15%.
- 2. The K_2O/Na_2O contents varied between 0.25 and 2.44.
- 3. The abundances of CaO in the samples varied between 1 and 6%.
- 4. The primitive mantle normalized trace element concentration data for the sediments (Fig. 4.16) show strongly depleted Nb and Ta patterns. Depletions in Sr and Zr are also observed in many samples. Enrichments in Th, U and Pb are prominent.
- 5. The concentrations of most of the elements, except for Nb, Ta and Pb, in the sediments of the core overlap with that observed in sediments in the Irrawaddy and Ganga-Brahmaputra river systems. Nb, Ta and Pb contents in sediments from top layers of the core mimic the pattern observed in island arc lavas from Barren Island (Fig. 4.16a).
- 6. The chondrite-normalized REE patterns for sediment layers in the core show enriched light REE patterns (Fig. 4.17), which tend to become more enriched towards the older half of the core (Fig. 4.17a-d). The total LREE contents in the top layers are much lower than that in the rest and in the sediments of the Irrawaddy and G-B river systems (Fig. 4.17a). The HREE patterns are relatively flat and all layers show moderate negative Eu anomalies (Eu/Eu* = 0.7-1.1).
- In both the plots, Fig. 4.16 and Fig. 4.17, the trace element contents in the sediments from the core overlap with the fields drawn for the Indo-Burman (IB) and Himalayan sources.

Sample code	SL-1 (0-5 cm)	SL-2 (15-20)	SL-3 (20-25)	SL-4 (30-35)	SL-32 (40-45)	SL-5 (55-60)
Mean Depth (cm)	2.5	17.5	22.5	32.5	42.5	57.5
(wt. %)						
SiO ₂				60.79		61.66
TiO ₂				0.81		0.86
Al ₂ O ₃				17.66		17.31
Fe ₂ O ₂ ^T				7.16		6.67
MnO				0.07		0.07
MgO				3.87		3.93
CaO				5.20		4.91
Na ₂ O				2.99		3.46
K-0				1.37		1.07
P.O.				0.07		0.07
				64.87		64 69
K.O/Na.O				0.46		0.31
1,000				00		
(ppm)	20.10	12.25	10.50	100.40		70.00
Rb	39.10	43.35	49.56	108.42		78.09
Sr	150.63	162.16	133.28	03.85		81.21
Ba	154.64	226.51	259.98	351.85		281.98
Y	15.80	16.01	14.07	8.49		12.94
Zr	53.90	51.63	50.53	43.40		65.66
Hf	1.88	1.74	1.74	1.94		2.25
Nb	1.44	1./1	2.11	4.98		4.37
Ta	0.14	0.16	0.20	0.40		0.43
Th	3.32	3.82	4.35	9.44		0.45
U	0.55	0.57	0.02	1.00		1.04
	0.55	7.50	0.19	15.50		22.41
Ce D-	14./2	2.10	2 10	23.05		22.41
rr Nd	1.91 9.22	2.10	2.19	5.09 11 57		2.70
INU S	0.55	0.91	0.97	2.25		2.25
Sm Fu	2.21	2.23	2.10	2.33		2.33
Eu	2.63	2.60	0.78	0.70		0.70
Gu Th	0.43	2.00	0.38	0.33		0.37
	3 10	3.05	2.78	2.35		2.65
Но	0.66	0.62	0.57	0.49		0.55
Fr	2.17	2.02	1.88	1.61		1 79
Tm	0.30	0.29	0.26	0.23		0.25
Vh	2 21	2.05	1.90	1.70		1.88
Lu	0.34	0.32	0.29	0.26		0.28
Sc	24.44	24.62	21.07	16.22		22.10
V	216.75	210.45	179.86	118.18		171.81
Cr	70.91	79.77	72.69	66.38		72.07
Со	14.51	16.31	12.27	6.44		6.88
Ni	38.67	49.68	47.44	47.73		55.55
Pb	5.94	10.58	6.26	7.15		7.92
Cs	3.34	3.77	4.39	10.19		7.36
La/Th	1.97	1.96	1.88	1.44		1.75
Th/Yb	1.50	1.86	2.29	5.55		3.43
Th/Sc	0.14	0.16	0.21	0.58		0.29
Th/U	6.28	6.73	6.97	9.45		6.18
Sm/Nd	0.26	0.25	0.24	0.20		0.22
Eu/Eu*	1.05	1.02	1.00	0.89		0.86
⁸⁷ Sr/ ⁸⁶ Sr	0.70531	0.70600	0.70629	0.70944	0.70967	0.70791
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512691	0.512629	0.512566	0.512354	0.512234	0.512435
$\varepsilon_{\rm Nd}(0)$	1.0	-0.2	-1.4	-5.5	-7.9	-4.0
T _{nu} (Ga)	1.04	1.07	1 11	1 17		1 16

Table 4.5: Geochemical and isotopic data for sediments from core SK-234-60

Major element oxide data are normalized to 100% on a volatile free basis. $CIA = [Al_2O_3/(AI_2O_3 + K_2O + Na_2O + CaO^*]$ 100 where CaO* represents CaO in silicate fractions. SL: Sediment Layer.

Sample code	SL-6 (65-70)	SL-7 (75-80)	SL-8 (85-90)	SL-9 (95-100)	SL-10 (105-110)
Mean Depth (cm)	67.5	77.5	87.5	97.5	107.5
(wt. %)					
SiO ₂	60.28	61.05	63.00		62.58
TiO ₂	0.93	0.82	0.92		0.90
Al ₂ O ₃	17.02	17.67	19.93		18.56
Fe ₂ O ₂ ^T	6.95	5.95	6.09		6.43
MnO	0.07	0.06	0.05		0.08
ΜσΟ	2.85	3.77	3.21		3.10
CaO	5.09	5.67	2.94		4.57
Na ₂ O	3.98	3.71	1.85		2.18
K ₂ 0	1.13	1.21	1.93		1.55
P ₂ O ₅	1.70	0.08	0.06		0.05
CIA	62.54	62.51	74.77		69.09
K2O/Na2O	0.28	0.33	1.05		0.71
<u>2</u> 22					
(ppm)	20.80	90.54	124 74	54.07	02.07
Rb	30.89	89.54	124.74	54.97	92.96
Sr	303.29	09.40	05.70	180.37	95.40
Ba	133.33	201.02	347.39	1/9.20	12 26
1	7 20	9.22 50.03	9.97	10.77	67.58
Zľ	0.60	1.87	2 11	1.87	2 15
III Nb	0.00	1.67	6.90	2.49	5.08
NU Ta	0.82	0.35	1.67	0.21	0.37
Ta Th	5 55	7.04	9.85	4 55	6.68
III II	6 78	1.04	1 49	2 74	1.05
La	11.05	12.10	16 30	11 34	12.28
Ce	24.98	23.30	31.06	25.59	24.35
Pr	2.83	2.81	3.64	3.19	2.93
Nd	11.17	10.53	13.35	13.44	11.08
Sm	2.56	2.13	2.51	3.38	2.33
Eu	0.61	0.65	0.66	1.00	0.70
Gd	2.63	2.08	2.30	3.77	2.37
Tb	0.38	0.30	0.31	0.58	0.35
Dy	2.49	2.10	2.11	3.96	2.47
Но	0.48	0.43	0.44	0.79	0.51
Er	1.48	1.41	1.45	2.47	1.68
Tm	0.20	0.20	0.21	0.33	0.25
Yb	1.42	1.47	1.54	2.39	1.75
Lu	0.21	0.22	0.23	0.36	0.26
Sc	10.55	19.59	21.01	22.00	21.56
V	13.60	154.54	158.25	211.06	159.67
Cr	0.00	71.21	98.26	44.12	75.11
Co	0.21	7.44	10.77	21.52	6.81
Ni	36.18	56.91	82.38	67.79	44.86
Pb	31.35	7.93	9.50	11.32	7.77
Cs	3.12	8.35	11.81	4.98	8.96
La/Th Th	1.99	1.72	1.65	2.49	1.84
Th/Yb	3.90	4.77	6.40	1.91	3.82
1 ft/SC	0.55	0.30	0.4/	0.21	0.31
1 fl/U Sm/NJ	0.82	0.75	0.39	1.00	0.33
SIII/INU Fu/Fu*	0.23	0.20	0.19	0.23	0.21
ЕЧ/ЕЧ" 87 _{Ст/} 86 _{Ст}	0.70	0.71	0.01	0.03	0.07
Sr/ Sr 14357 14457 5	0.70947	0.70889	0./1132	0.70758	0.70801
Nd/ ^T Nd	0.512353	0.512361	0.512286	0.512521	0.512367
ε _{Nd} (U) T	-3.0	-3.4	-0.9	-2.3	-3.3
T _{DM} (Ga)	1.39	1.15	1.16	1.30	1.20

Table 4.5: Continued

Sample code	SL-11 (115-120)	SL-12 (125-130)	SL-13 (135-140)	SL-14 (145-150)	SL-15 (155-160)
Mean Depth (cm)	117.5	127.5	137.5	147.5	157.5
(wt. %)					
SiO ₂	62.66	61.96	61.88		61.56
TiO ₂	0.83	0.95	0.86		0.89
Al ₂ O ₃	18.46	19.27	19.89		19.27
Fe ₂ O ₃ ^T	5.85	9.18	8.17		8.71
MnO	0.04	0.06	0.06		0.06
MgQ	3.09	3.03	3.20		2.91
CaO	3.55	2.06	2.23		2.10
Na ₂ O	3.64	1.13	1.50		1.48
K ₂ O	1.83	2.27	2.16		2.25
P ₂ O ₂	0.04	0.07	0.05		0.77
	67.16	77 93	77.16		76 78
K.O/Na.O	0 50	2 00	1 44		1 51
1120/11120	0.20	2.00	1		1.01
(ppm)					
Rb	122.95	129.65	140.85	131.36	115.66
Sr	70.63	56.01	60.59	69.60	213.64
Ba	291.39	258.92	381.89	376.87	340.36
Y	8.46	8.46	9.19	11.04	20.19
Zr	63.80	59.10	76.76	62.36	41.44
Hf	2.11	2.09	2.20	2.01	1.86
Nb	6.78	6.84	7.48	6.60	6.24
Та	0.48	0.48	0.51	0.49	0.51
Th	9.09	10.74	10.63	10.42	12.58
U	1.22	1.79	1.60	2.03	8.02
La	15.51	17.65	17.58	18.25	26.93
Ce	29.82	33.93	33.64	36.52	59.09
Pr	3.45	3.93	3.89	4.20	6.77
Nd	12.53	14.28	14.14	15.52	26.02
Sm	2.33	2.58	2.52	2.99	5.64
Eu	0.63	0.64	0.60	0.74	1.32
Gđ	2.11	2.33	2.24	2.73	5.4/
Tb	0.29	0.30	0.30	0.36	0.73
Dy	1.88	1.97	1.85	2.32	4.52
Ho	0.39	0.40	0.39	0.48	0.83
Er T	1.55	1.54	1.50	1.55	2.49
l m Vh	0.19	0.19	0.19	0.21	0.33
YD L	1.39	1.45	1.41	1.55	2.33
Lu	0.20	0.21	0.20	0.23	0.55
SC V	10.44	120.20	164.85	157 56	115.68
v Cr	71 70	87.80	104.85	85 33	50.44
	2.14	11 70	101.89	15 18	12.28
Ni	2.14	76.32	94.60	07.37	97.04
Dh	9 1/	11 56	10.21	92.32 11 5 4	40 5 9
	12.14	12.55	13.65	12.61	11.83
Cs I a/Th	1 71	1 64	1.65	1 75	2 14
Th/Vh	6.52	7 52	7 52	6.73	5 36
Th/Sc	0.55	0.59	0.49	0.50	0.96
Th/I	7.46	5.99	6.63	5.14	1.57
Sm/Nd	0.19	0.18	0.18	0.19	0.22
En/En*	0.84	0.77	0.75	0.77	0.70
⁸⁷ Sr/ ⁸⁶ Sr	0.71160	0.71222	0.71343	0.71155	0.71088
143 _{NLJ} /144 _{NLJ}	0.71107	0.71222	0.71343	0.71133	0.71000
1NU/1NU	0.312241	0.31219/	0.312184	0.312238	0.312207
ε _{Nd} (U)	-/./	-0.0	-0.7	-7.4	-1.2
I _{DM} (Ga)	1.21	1.24	1.24	1.23	1.42

Table 4.5: Continued

Sample code	SL-33 (165-170)	SL-16 (175-180)	SL-17 (195-200)	SL-18 (210-215)	SL-19 (225-230)
Mean Depth (cm)	167.5	177.5	197.5	212.5	227.5
(wt. %)					
SiO ₂	62.77	61.63	63.78	57.01	54.01
TiO	0.97	0.89	0.95	0.77	0.71
ALO.	19.86	19 44	18 90	17.81	16 78
	0.10	9.7	9.40	9.00	7.50
Fe ₂ O ₃	8.18	8.0/	8.49	8.00	7.59
MnO	0.06	0.06	0.05	0.06	0.18
MgO	2.73	3.10	2.44	2.86	2.83
CaO	2.01	2.15	1.52	2.01	2.62
Na ₂ O	1.10	1.52	1.10	8.51	8.02
K ₂ O	2.24	2.38	2.25	2.38	2.60
P_2O_5	0.06	0.16	0.53	0.59	4.68
CIA	78.78	76.26	79.51	58.00	55.91
K ₂ O/Na ₂ O	2.03	1.56	2.04	0.28	0.32
(ppm)		140.20	120.52	141.25	1 (0.10
Rb		148.29	139.52	141.35	168.18
Sr		69.86	64.86	184.53	26.00
Ba		351.04	310.22	270.50	307.88
Y		8.55	5.55	16.50	9.73
Zr		62.77	48.64	31.75	57.13
Hf		2.38	1.98	1.62	2.39
Nb		8.36	8.54	7.16	9.03
Та		0.60	0.66	0.52	0.63
Th		11.26	9.54	13.58	12.70
U		1.91	1.33	4.62	2.46
La		19.42	18.94	28.02	23.61
Ce		37.93	37.18	62.78	47.75
Pr		4.45	4.40	6.80	5.52
Nd		16.42	16.18	25.85	20.25
Sm		3.01	3.00	5.40	3.81
Eu		0.74	0.71	1.24	0.86
Gd		2.71	2.57	5.12	3.35
Tb		0.34	0.31	0.67	0.43
Dv		2.20	1.91	3.97	2.63
Ho		0.44	0.36	0.71	0.52
Er		1 46	1 10	2 13	1 64
Tm		0.21	0.16	0.27	0.24
Yh		1 49	1 10	1.98	1.71
In		0.21	0.15	0.28	0.25
Sc		18 55	17.28	13.80	18 19
V		143.00	17.20	130.93	120.05
Cr.		89.22	89 56	76 75	120.05
		11 14	4 21	18.25	14.86
Ni		01 33	64.63	118 11	122 52
INI Dh		9 95	7 20	20.50	122.52
ru Ca		0.05	12.41	14.65	16.44
US La/Th		14.05	12.41	2.06	10.44
La/10 Th/X/h		1./5	1.99 9.70	2.00	1.80
1 11/ Y D Th/So		1.34	0.70	0.02	/.43 0.70
1 11/SC TL/U		0.01	0.33	0.98	0.70
		5.90 0.19	/.10	2.94	3.1/ 0.10
Sm/Nd		0.18	0.19	0.21	0.19
Eu/Eu*		0.//	0.76	0.70	0./1
°′Sr/°°Sr	0.71135	0.71349	0.71299	0.71140	0.71680
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512265	0.512204	0.512210	0.512321	0.512206
$\varepsilon_{\rm Nd}(0)$	-7.3	-8.5	-8.3	-6.2	-8.4
T _{DM} (Ga)		1.25	1.25	1.26	1.28

Table 4.5: Continued

Table 4.5:	Continued
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Sample code	SL-20 (240-245)	SL-21 (255-260)	SL-34 (265-270)	SL-22 (275-280)	SL-23 (285-290)
Mean Depth (cm)	242.5	257.5	267.5	277.5	287.5
(wt. %)					
SiO ₂	60.61	59.37	58.25	62.40	57.76
TiO ₂	0.87	0.86	0.93	0.91	0.76
Al ₂ O ₃	18.41	17.96	18.17	19.11	17.21
Fe ₂ O ₃ ^T	8.73	8.16	8.77	6.74	6.02
MnO	0.07	0.09	0.06	0.08	0.09
MgO	3.36	3.25	2.86	3.19	3.16
CaO	2.14	3.76	0.82	1.96	5.29
Na ₂ O	3.10	2.93	7.73	2.20	7.71
K ₂ O	2.07	1.97	2.34	2.37	1.90
P_2O_5	0.64	1.66	0.06	1.01	0.09
CIA	71.58	67.47	62.52	74.51	53.60
K ₂ O/Na ₂ O	0.67	0.67	0.30	1.08	0.25
(nnm)					
(ppm) Rh	139 50	48 52		104 46	121 79
Sr	51.67	366.31		141.60	103.36
Ba	344.96	139.44		200.82	330.29
Y	8.77	35.61		16.80	12.41
Zr	61.86	21.00		39.09	85.22
Hf	2.24	1.37		2.00	2.68
Nb	8.24	3.44		6.23	7.68
Та	0.60	0.29		0.44	0.53
Th	9.98	18.37		11.26	9.51
U	1.60	5.91		4.70	1.55
La	19.64	36.12		23.71	18.18
Ce	38.74	94.25		55.66	37.08
Pr	4.59	9.60		5.94	4.43
Nd	16.96	38.43		23.11	16.77
Sm	3.17	9.12		5.03	3.35
Eu	0.79	2.18		1.19	0.88
Gd	2.82	9.31		4.89	3.09
Tb	0.36	1.32		0.68	0.42
Dy	2.28	8.13		4.19	2.74
Но	0.45	1.45		0.77	0.55
Er	1.49	4.26		2.31	1.77
Tm	0.21	0.55		0.31	0.24
Yb	1.53	3.74		2.22	1.78
Lu	0.22	0.54		0.32	0.26
Sc	18.73	11.06		12.47	20.39
V	134.16	23.78		103.58	160.63
Cr	118.29	-0.09		52.20	120.64
Co	10.63	3.32		15.40	20.54
NI	103.44	59.51 22.69		113.88	116.20
Pb	/.0/	52.08		33.48	9.4/
US Lo/Th	12.08	3.33		10.71	10.98
La/III Th/Vh	1.97	1.97		2.11	5.25
Th/10	0.54	т.71 166		0.00	5.55 0.47
Th/J	6.22	3.11		2.40	6.13
Sm/Nd	0.19	0.24		0.22	0.15
En/En*	0.19	0.24		0.22	0.20
⁸⁷ Sr/ ⁸⁶ Sr	0.71317	0.70	0 71753	0.71060	0.71073
143 _{NJJ} /144 _{NJJ}	0.71317	0.70934	0.71733	0.71007	0.71075
ra/ra	0.312243	-8.0	0.312214	-7.0	-6.4
	- / . /	-0.0	-0.5	-7.0	-0. 1 1 21
$\mathbf{D}_{\mathbf{D}\mathbf{M}}(\mathbf{G}\mathbf{a})$	1.22	1./2		1.40	1.41

Sample code	SL-24 (295-300)	SL-35 (305-310)	SL-25 (315-320)	SL-26 (325-330)	SL-27 (340-345)
Mean Depth (cm)	297.5	307.5	317.5	327.5	342.5
(wt. %)					
SiO ₂	61.10		64.87	61.41	63.28
TiO ₂	0.86		0.94	0.76	0.91
Al ₂ O ₃	19.08		20.10	17.65	19.79
Fe ₂ O ₃ ^T	6.97		5.64	7.16	7.68
MnO	0.08		0.04	0.04	0.04
MgO	3.56		2.84	3.45	2.96
CaO	2.75		1.08	3.80	1.48
Na ₂ O	2.63		1.35	2.86	1.08
K,0	2.35		2.58	2.29	2.63
P ₂ O ₅	0.64		0.54	0.58	0.13
CIA	71.18		80.04	66.36	79.22
K ₂ O/Na ₂ O	0.89		1.90	0.80	2.44
(
(ppm) Dh	103 42		05.63	153 47	120.20
KU Sm	103.42		201 42	75.05	201.62
Ba	49.00 253 76		201.42	328.61	364.38
Da V	6.64		201.24	9 50	18 85
1 7r	0.04 46 38		36.05	59.64	57 30
Hf	1 66		1 37	1 91	2.05
Nh	6.75		5.15	8 79	6.48
Та	0.46		0.39	0.81	0.48
Th	7.66		12.23	11.70	13.26
U	1.13		3.52	1.66	9.71
La	15.17		25.42	19.64	26.98
Ce	30.22		58.75	38.08	57.59
Pr	3.56		6.31	4.44	6.56
Nd	13.29		24.20	15.79	24.75
Sm	2.51		5.23	2.86	5.09
Eu	0.61		1.16	0.64	1.16
Gd	2.21		5.09	2.49	4.79
Tb	0.28		0.69	0.32	0.65
Dy	1.79		4.20	1.98	4.00
Но	0.33		0.77	0.39	0.78
Er	1.03		2.32	1.29	2.35
Tm	0.15		0.31	0.18	0.32
Yb	1.06		2.11	1.37	2.31
Lu	0.15		0.31	0.19	0.33
Sc	14.44		13.28	16.57	15.29
V	101.89		106.97	130.74	140.71
Cr	82.48		59.08	111.04	80.01
Со	13.22		14.27	10.76	13.79
Ni	80.70		79.45	83.85	112.53
Pb	6.18		35.99	8.51	46.16
Cs L - /Th	9.22		9.60	13.76	13.50
	1.98		2.08	1.08	2.04
1 11/ 1 1) Th/So	1.2 4 0.52		J.17 0.02	0.33	J. /4 0.97
т II/SC Ть/П	6.80		2.48	7.05	1.37
s II/U Sm/Nd	0.00		0.72	0.18	0.21
5m/ivu Fu/Fu*	0.19		0.22	0.10	0.21
87 Sr/86 Sr	0.71466	0 71220	0.71063	0.71186	0.71071
143 _{NL4/} 144 _{NL4}	0.71400	0.71330	0.71003	0.71100	0.71071
ru/ru	-8 1	-7 5	-0.8	-0.0	-8 1
$c_{\rm Nd}(v)$	-0.4 1 20	-1.5	- 2.0	-9.0	-0.1
$I_{DM}(Ga)$	1.20		1.05	1.2/	1.37

0 1 1	SI 26 (255 260)	CI 20 (2(5.250)	CI 20 (255 200)	CI 20 (207 200)	CI 21 (205 400)
Sample code	SL-36 (355-360)	SL-28 (365-370)	SL-29 (375-380)	SL-30 (385-390)	SL-31 (395-400)
Mean Depth (cm)	357.5	367.5	377.5	387.5	397.5
(wt. %)	<- 1 -		<1 - ((2.1)	
SiO ₂	65.17		61.76	62.44	
TiO ₂	0.96		0.89	0.90	
Al_2O_3	20.21		18.81	19.13	
Fe ₂ O ₃ ^T	5.75		7.67	7.12	
MnO	0.05		0.07	0.07	
MgO	2.57		3.21	3.14	
CaO	1.48		3.68	2.77	
Na ₂ O	1.12		1.72	1.42	
K ₂ O	2.52		2.12	2.17	
P_2O_5	0.16		0.10	0.86	
CIA	79.79		71.44	75.04	
K ₂ O/Na ₂ O	2.24		1.23	1.53	
(ppm)	12(20		112.02	129.04	120.46
KD S	130.30		113.92	128.04	120.46
Sr	72.95		111./4	95.15	508.28
Ва У	2/4.00		319.55	28/.90	508.28
Y 7	9.91		13.08	11.08	15.40
	109.10		95.09	/ 3.49	/ 3.88
HI	2.92		2.40	2.20	2.20
ND T-	0.62		7.04	7.83	/.04
1 A Th	0.03		0.38	0.34	0.34
	9.30		0.44 1.26	8.90 1.41	9.45
U	1.30		1.50	1.41	2.03
La	19.12		22.00	10.39	19.70
Ce Du	30.90		32.09	30.03	59.// 1.67
rr Na	4.29		3.02 14.22	4.24	4.07
Nu Sm	13.02		14.33	2.02	2 29
Sill Fu	2.77		2.79	2.95	0.85
Eu	2.40		0.74	2.64	3.15
Gu Th	2.40		0.38	0.35	0.42
TU Du	2.02		2.56	0.35	0.42
Dy Uo	2.02		2.50	0.47	2.07
no Fr	0.42		1.78	1.54	1.69
Tm	0.20		0.25	0.22	0.24
Tm Vh	1.45		1.77	1.58	1.76
In	0.21		0.27	0.23	0.26
Eu Se	18.65		23 31	20.83	20.13
V	144 38		205.05	153 32	149 79
Cr	116.70		132.94	119.87	126.60
Co	8.79		13.76	12.11	18.29
Ni	77.91		87.27	87.00	116.97
Ph	7.87		6.95	8.47	17.78
Cs	12.65		10.31	12.03	11.19
La/Th	2.06		1.90	2.07	2.09
Th/Yb	6.42		4.76	5.63	5.37
Th/Sc	0.50		0.36	0.43	0.47
Th/U	5.97		6.22	6.31	4.61
Sm/Nd	0.18		0.19	0.19	0.19
Eu/Eu*	0.76		0.80	0.78	0.77
⁸⁷ Sr/ ⁸⁶ Sr	0.71048	0.71155	0.71436	0.71644	0.71132
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512203	0.512280	0.512352	0.512279	0.512270
ε(0)	-8 5	-7.0	-5.6	-7 0	-70
	1.21	7.0	1.11	1.17	1.22
IDM (Ga)	1.41		1.11	1.1/	1.22

Table 4.5: Continued



Fig. 4.16: Primitive mantle (PM) normalized multi element spidergrams for the sediments in the core. Data Sources: average compositions of Irrawaddy sediments (Kurian et al., 2008); Barren Island volcanic (BI) (Luhr and Haldar, 2006; Pal et al., 2010); fields for sediments from Indo-Burman (IB) sources (this study, determined from rocks of Andaman Islands); sediments from Himlayan sources brought by Ganga-Brahmaputra (GB, dashed line) (Stummeyer et al., 2002).



Fig. 4.17: Chondrite normalized REE patterns for the sediments in the core. Fields are same as in Fig. 4.16.

(C) Sr-Nd Isotopic ratios

Sr–Nd isotopic ratio analyses were also carried out on the decarbonated sliciclastic sediments from selected layers in the core. These data are presented in Table 4.5 and in histograms in Fig. 4.18. Neglecting the top disturbed part of the core, the data reveal that 87 Sr/ 86 Sr of sediments varies between 0.707 and 0.718 and ϵ_{Nd} values range from -9.8 to -2.3. The modes of 87 Sr/ 86 Sr and ϵ_{Nd} distributions in our core are 0.711 and -7, respectively (Fig. 4.18a and b). The lower part of the core between 120 cm and 400 cm, contains more radiogenic Sr (and non-radiogenic Nd) bearing sediments, while sediments from top 120 cm of the core have comparatively lower 87 Sr/ 86 Sr and higher ϵ_{Nd} . The lowest 87 Sr/ 86 Sr value (and the highest ϵ_{Nd} value) is observed at ~100 cm depth.



The ε_{Nd} versus 87 Sr/ 86 Sr compositions of core sediments are plotted in Fig. 4.21. Overall, our data show values with lower 87 Sr/ 86 Sr and higher ε_{Nd} as compared to the studies done on sediments from the other parts of the Andaman Sea (Colin et al., 1999). The histogram for the depleted mantle ages (T_{DM}) for these sediments is shown in Fig. 4.18c. The T_{DM} ages range from 1.11 Ga to 1.72 Ga and show a prominent mode at ~1.25 Ga (Fig. 4.18c).



Fig. 4.18: Histograms showing frequency distributions of (a) ${}^{87}Sr/{}^{86}Sr$, (b) ε_{Nd} and (c) TDM ages for siliciclastic sediment layers in the core SK-234-60.

4.5.3 Discussion

(A) Weathering in the source regions

Chemical Index of Alteration (CIA) and $AI_2O_3 + (CaO^* + Na_2O) + K_2O$ (A-CN-K) ternary diagram provide valuable information about the weathering conditions at the sources at the time of derivation of these sediments in our core. Figure 4.19 shows our data plotted in an A–CN–K diagram with CIA scale on the left. Our data points trend parallel to the A–CN join suggesting variable degree of weathering of source rocks. The CIA value of these sediments varies from 53 to 80, which again hint at low to moderate degree of weathering of their sources. The sources for the sediments are compositionally similar to tonalities suggesting higher contribution from lower crust and/or juvenile mafic igneous rocks compared to sediments in Bay

of Bengal or to PAAS. We suspect that this juvenile material contribution could have come from the Indo-Burman mountain ranges of Myanmar.



Fig. 4.19: A-CN-K diagram with the Chemical Index of Alteration (CIA) of Nesbitt et al, (1996) on which core sediments are plotted. Also shown on the diagram are the positions of minerals, post-Archean Australian Average Shale (PAAS), and average Archean upper crust. Arrows emanating from plagioclase-K-feldspar join show the weathering trends for basalt, tonalite, granodiorite and granite (Nesbitt and Young, 1984, 1989).

Th/U ratios in sediments are controlled by weathering-erosion-diagenesis cycle (Condie, 1993). In Fig. 4.20, Th/U ratios of most of our samples are higher than the upper crustal value of 3.8 (Taylor and McLennan, 1985) suggesting derivation from recycled crustal material, which has lost U. Two of our samples plot within the field for depleted mantle indicating their link to magmatic arc volcanoes. Some samples show lower Th/U ratios and higher Th values. These samples probably suffered U enrichment under reducing conditions of deposition.



Fig. 4.20: Plot of Th/U versus Th (after McLennan et al., 1995) for the core sediment samples from the core.

(B) Provenance

Significant variations observed in the Sr and Nd isotopic compositions of the sediments probably indicate variable contributions from multiple sources to the core site. In the absence of Sr-Nd isotopic data on likely continental sources, we used published data on sediments from major river systems that drain into the Bay of Bengal and Andaman Sea and compared them with our data. As suggested by earlier studies, the main source of sediments to the Andaman Sea is the sediments discharged by the Irrawaddy river system (Colin et al., 1999; Ramaswamy et al., 2004). However, considering that there exists a major easterly flowing surface current during the period of South-West monsoon (Rodolfo, 1969), we expect significant sediment contribution of Himalayan derived Ganga-Brahmaputra (G-B) sediments from the Bay of Bengal to the Andaman Sea by this current. Comparing $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} data from our core (SK-234), with the available data from other cores studied in the Andaman Sea and NE Bay of Bengal (Fig. 4.21), we observe that isotopic ratios in our core (${}^{87}\text{Sr}/{}^{86}\text{Sr}=$ 0.707–0.718; ϵ_{Nd} = -9.8 to -2.3) are less radiogenic as compared to the sediments of the eastern Andaman Sea (⁸⁷Sr/⁸⁶Sr= 0.712 to 0.719; ϵ_{Nd} = -10.8 to -9.3, Core-MD77-169, Colin et al., 2006) and sediments from Irrawaddy (${}^{87}Sr/{}^{86}Sr$ = 0.713; ϵ_{Nd} = -10.7, Colin et al., 1999) (Fig.

4.21). This observation points to the involvement of sources other than Irrawaddy which were also actively supplying sediments to the core site (Fig. 4.15). While most of the isotopic data from our core overlap with those from the cores MD77-171 and MD77-176, a general shift towards lower 87 Sr/ 86 Sr and higher ϵ_{Nd} values is observed. Such a trend appears to be a result of contributions from a third source. We believe that Cretaceous-Oligocene sedimentary rocks, volcanics and ophiolitic rocks present in the Indo-Burman Ranges (Allen et al., 2008; Colin et al., 2006) represent this third source.



Fig. 4.21: Plot of ε_{Nd} versus ⁸⁷Sr/⁸⁶Sr for sediments from our core (SK-234) and cores in the region studied by others. Data sources: sediments of Irrawaddy, sediments of the cores from Andaman Sea and NE-Bay of Bengal: Colin et al., (1999), Colin et al., (2006); Indo-Burman Ranges: Allen et al., (2008), this study, determined from rocks of Andaman Islands; Ganga-Brahmaputra (G-B): Ahmad et al., (2005) and references therein. All data were normalized to NBS-987 (0.710250) and La Jolla (0.511858) for ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd isotopic ratios, respectively.

From the above discussion it is apparent that our site received sediments from three different sources drainage systems; viz: Myanmar continental sediments through Irrawaddy river system, Indo-Burman mountain ranges and Himalaya derived Ganga-Brahmaputra (G-B) sediments from the Bay of Bengal. A three componentmixing calculation using these three sources suggest that the Irrawaddy and Indo-Burman ranges are the main contributors of sediments to our core site; however, there has been significant contribution of G-B sediments from the north-east Bay of Bengal as well (Fig. 4.22). A closer look at these contributions through time reveals that the sediments deposited during the Last Glacial Maximum (LGM), with lower ⁸⁷Sr/⁸⁶Sr and high ε_{Nd} values, have relatively larger (upto 98 %) contribution from the Indo-Burman sources (Fig. 4.21 and 4.22).



Fig. 4.22: Plot of ε_{Nd} vs. ⁸⁷Sr/⁸⁶Sr of sediments from our core compared with model curves of a three components mixing involving sediments from Indo-Burman Ranges, Irrawaddy river system and Ganga-Brahmaputra (G-B) river system. The end-member compositions used in these calculations are: Indo-Burman (Sr= 200 ppm; Nd= 24 ppm; ⁸⁷Sr/⁸⁶Sr= 0.707; $\varepsilon_{Nd} = -2$), Irrawaddy (Sr= 100 ppm; Nd= 50 ppm; ⁸⁷Sr/⁸⁶Sr= 0.712; $\varepsilon_{Nd} = -10.7$) and Ganga-Brahmaputra (Sr= 400 ppm; Nd= 15 ppm; ⁸⁷Sr/⁸⁶Sr= 0.726; $\varepsilon_{Nd} = -12.6$). The source for the end-member compositions are: Indo-Burman Ranges: Allen et al., (2008), this study, determined from rocks of Andaman Islands, Irrawaddy & G-B: Colin et al., (1999), Ahmad et al., (2005).

The results from isotopic data are further supported by the REE distributions in the samples from the core. The lower REE contents in these compared to that in Irrawaddy sediments suggest that there exists an additional source, possibly of mafic-magmatic in nature. As mentioned above such mafic sources, which can contribute sediments to the Andaman Sea, are extensively present only in Indo-Burman Ranges (Allen et al., 2008; Colin et al., 2006). Except for the top disturbed part of the core (showing patterns similar to Barren Island volcano), all our data fall within the field drawn for the Indo-Burman sources (Fig. 4.17). Samples from
further down the core show relatively enriched LREE pattern suggesting more contribution from felsic continental sources in the past (Fig. 4.17). These can be due to increased contribution from Himalyan derived sediments either carried through Irrawaddy or Ganga-Brahmaputra systems. Although the A-CN-K diagram (Fig. 4.19) suggests an average source rock composition for the core sediments to be tonalitic, considering the above results from isotopic ratios we believe that the evolutionary trend shown by core sediments reflects mixing of sediments from multiple sources (one basaltic and other granitic/granodioritic).

(C) Climatic Implications

The Sr-Nd isotopic ratios of sediments in the Andaman Sea show striking shift as a function of time (depth) (Fig. 4.23). These excursions apparently reflect changes in provenance and most likely are governed by changing monsoonal strength. In the following discussion we discuss climatic implications of the chemical variations observed in the core in three time periods: pre-LGM, during LGM and post-LGM. By doing so, we look into the sedimentary record to search for evidences for changes in the sea-level, oceanic circulation patterns, sediment transport and monsoonal intensity.

The down core record prior to the LGM (at 18 kyr) shows several excursions towards lower ⁸⁷Sr/⁸⁶Sr and higher ε_{Nd} at ~36 kyr, ~44 kyr, ~52 kyr and ~58 kyr (Fig. 4.23). We believe that these excursions, except for the ~52 kyr event, are correlated to the Heinrich events H4, H5 and H6, respectively, which occurred during the last glacial period (Hemming, 2004). The excursion at ~52 kyr might also be related to some event similar to it. Earlier studies have shown that during the Heinrich global cold events the intensity of the Asian summer monsoon had remained very weak (Porter and Zhisheng, 1995; Colin et al., 1998). The penetration of the monsoon in the interior parts of the continental source areas largely depends on its intensity. As rivers in South Asia are mainly fed by sediments eroded by rainfall during summer monsoons, the weak summer monsoons during these cold periods probably restricted the penetration of monsoonal rainfall into the inner parts



Fig. 4.23: Depth profiles of Sr-Nd isotopic compositions, trace element ratios and CIA values of sediments in our core SK-234-60. HE1-HE6 on left are Heinrich cold events (Hemming, 2004) and numbers 1 to 20 on right are Dansgaard-Oeschger warm events (Schulz, 2002). Curve for mean effective moisture is from Herszchuh, (2006). The arrows mark significant excusions (purple: cold events and green warm events). Irr: Irrawaddy; IB: Indo-Burman Ranges; G-B: Ganga-Brahmaputra; MIS: Marine Isotopic Stage.

of Myanmar and thus reducing the supply of high ⁸⁷Sr/⁸⁶Sr bearing sediments from the glaciated regions of the Himalaya. It is highly likely that during these events the relative contribution of sediments derived from the Indo-Burman Ranges, through the seaways between Bay of Bengal and Andaman Sea, increased to our core site. The weak summer monsoon conditions at ~36 kyr, ~47 kyr, and ~58 kyr BP have also been observed in a core studied from the Bay of Bengal by Colin et al. (1998), in which, increase in size of magnetic grains and decrease in CIA value have been correlated to Heinrich events. In our core, we also observe excursions in some key elemental ratios (La/Th, Th/Yb, Sm/Nd and Th/Sc) at the same places where isotopic shifts have been observed. These apparently are results of changes in provenance and relative contribution of sediments from felsic Himalayan-Transhimalayan sources and mafic Indo-Burman sources. The correlateable excursions observed in the isotopic data and geochemical indicators confirm that these variations are related to change in contribution of sediments from various sources, and not to processes like particle sorting/weathering (Colin et al., 2006; Tutken et al., 2002; Walter et al., 2000). As the sources are located in the intertropical regions, wet summer monsoons cause relative increase in chemical weathering of sediments in river plains, and loss of the most mobile elements (Na, K, and Ca) from the detrital minerals (feldspars and glass from volcanic rocks) producing pedogenic clays minerals (e.g., smectite; kaolinite) leading to increase in CIA value. In contrast, dry periods are characterized by enhanced physical erosion in the higher reaches which results in supply of high volumes of unaltered minerals that reduces the CIA value of bulk sediments. The lowering of CIA value in our core at ~44 kyr and ~ 58 kyr further support the hypothesis of weak summer monsoon conditions (related to Heinrich events) during these times. In addition to the above, we also observe variations that show significant increase in ⁸⁷Sr/⁸⁶Sr and decrease in ε_{Nd} at ~46 kyr, ~54 kyr, ~60 kyr and ~72 kyr in the core (Fig. 4.23). We postulate that these are associated with the Dansgaard–Oeschger warm events like DO-3, 8, 9, 10, 13 and 16 during which the South Asian summer monsoon had regained its strength that caused higher input from the Himalayan sources through Irrawaddy river system. Support for such an inference comes from an earlier study by Herzschuh (2006), which reported intensification of SW monsoon at ~47 ka and ~54 ka.

Many studies have shown that sediments discharged by the global rivers were greatly reduced during the LGM (Duplessy, 1982; Van Campo, 1986), when the sea level was lower by 100-125 m (Chappell and Shackleton, 1986). Lower sedimentation rates during the LGM have also been reported from cores from the Bay of Bengal (Tripathy et al., 2011, Kessarkar et al., 2005, Galy et al., 2008). In fact, the lowest rate of the sedimentation observed by us in our core occurred during ~17-23.5 ka. This reduced supply of sediments to the Andaman Sea during the LGM could to be attributed to slower continental erosion. The Sr-Nd isotopic composition observed in the core during the LGM support this hypothesis. The observed $^{87}\mathrm{Sr}/^{86}\mathrm{Sr}$ ratio during the LGM (~95-110 cm) is ~0.708 compared to its value of ~0.711 in sediments deposited before and after the LGM (Fig. 4.23). ε_{Nd} shows a higher value of -2.3 during the LGM compared to pre- and post-LGM values of -7 (Fig. 4.23). Similar trends of lower ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ and higher ϵ_{Nd} during the LGM have also been observed in the sedimentary record of the Bay of Bengal (Tripathy et al., 2011). Such trends can be attributed to lowering of contribution from older continental sources particularly from the Himalayas. During the LGM mountain glaciers had covered much of the higher reaches of Tibet and Himalaya (Yokoyama, 2000; Tripathy et al., 2011) and there was a weak SW monsoon (Herzschuh, 2006; Prell and Kutzbach, 1987) and an intensified NE monsoon (Duplessy, 1982; Prell et al., 1980; Sarkar et al., 1990; Tiwari et al., 2005). It appears that the weakening of SW monsoonal rainfall is the primary reason for reduced erosion in the higher Himalayas and continental Myanmar which led to reduced sediment input from these sources. However, such a weakening probably did not affect the sediment derivation and input from the Indo-Burman Ranges, which explains the isotopic shifts. Such an inference is also supported by the work of Colin et al. (1999), who observed increase in ε_{Nd} during the LGM in cores from NE Bay of Bengal, known to have dominant contribution of sediments from the Indo-Burman Ranges. La/Th, Th/Sc, Sm/Nd and Th/Yb ratios of sediments in our core also show excursions at ~100 cm depth corresponding to the LGM indicating increase in contributions from mafic igneous sources (Fig. 4.23).



Fig. 4.24: Temporal variations of Sr and Nd isotopic compositions of sediments in core SK-234-60 compared with other cores MD77-169 (Sewell seamount), MD77-176 from the Andaman Sea. Also shown δO^{18} (‰) variations of core 125KL from Bay of Bengal (Kudrass et al., 2001). Numbers 1 to 20 are Dansgaard-Oeschger warm events while HE0-HE6 are Heinrich cold events (Hemming, 2004). YD: Younger Dryas event, B/A: Bølling-Allerød event, MIS: Marine Isotopic Stage.



Fig. 4.25: (a) Present-day N-S bathymetric profile of Andaman and Nicobar ridge showing various channels through which Andaman Sea exchanges water with Bay of Bengal (b) The coastlines around the Andaman sea (shown in green shade) during the LGM with a 100m drop in sea-level. The South Preparis Channel was the only open passage between the Bay of Bengal and Andaman Sea then.

The sedimentary record subsequent to the LGM shows an increase in 87 Sr/ 86 Sr and decrease in ϵ_{Nd} at ~15 kyr, ~10 kyr and at ~6 kyr BP (Fig. 4.23). The first two excursions are probably related to events of intensification of the SW monsoon during warm and moist Bølling-Allerød (B/A) event that happened somewhere between 14.7 and 12.7 kyr (Cronin, 1999) and during the Pleistocene-Holocene transition (Bookhagen et al., 2005). The third excursion at ~6 kyr followed a weak summer monsoon event at ~8.2 kyr (Alley et al., 1997; Wang et al., 2005). The B/A event initiated the end of the cold period of the LGM and resulted in worldwide rise in the sea level to more than 100 m due to melting of glaciers. The excursions observed in 87 Sr/ 86 Sr and ϵ_{Nd} are correlatable to excursions seen in various elemental ratios and CIA (Fig. 4.23). The drop in CIA value at ~10 ka was probably due to sudden increase in supply of unweathered sediments during the Holocene transition. The chemical variations observed in the top 20 cm of the core have been neglected from discussion as it was disturbed and consisted of sediments mixed with inseparable volcanic ash.

(D) Implications for ocean circulation

For a better understanding of the provenance and climate signals in our core (SK-234-60), we compared our record with other such studies in the Andaman Sea and Bay of Bengal (Fig 4.22). The major climatic fluctuations observed in our core correlate well with that in the δO^{18} record in the core 125KL from the Bay of Bengal (Kudrass et al., 2001), which has been well correlated with major climatic variations in the northern hemisphere. The overall trend of deposition of non-radiogenic Sr (and radiogenic Nd) sediments in the core since the LGM suggests that mafic igneous sources, most likely located in the Indo-Burman Ranges, contributed significant amounts of sediments to the core site (Fig. 4.24). Similar trends, although with lower amplitudes, are also observed in several other cores from the Andaman Sea (e.g., MD77-176, MD77-169) (Fig. 4.24). We believe that these trends are related to the increase in the sea level after the LGM (Hashimi et al., 1995; Goodbred and Kuehl, 2000). During and before the LGM, the continental margins surrounding the Andaman Sea and Bay of Bengal consisted of exposed uplands and rivers such as Ganga, Brahmaputra, Irrawaddy and Salween incised into the continental shelf- pouring sediments directly into the deep sea. The low sea level must have influenced the water circulation pattern and sediment transport restricting deposition to the deeper parts of the seas. It is highly likely that prior to and during the LGM the exchange of water and sediments between the Andaman Sea and Bay of Bengal was completely restricted to the narrow "Preparis South Channel" (Fig. 4.25a), as the "Preparis North Channel" was non-existence because of lower sea level (Fig. 4.25b). After the LGM, major changes occurred in the supply and deposition of sediments, eustatic sea level and surface circulation patterns in the north-eastern Indian Ocean. These were related to increase in precipitation and discharge by rivers (Goodbred and Kuehl, 2000). Subsequent rise in sea level flooded the continental margin and most of the discharged sediment load possibly got trapped in the inner shelf of Myanmar continental margin. The rise in sea-level might have opened up the broad "Preparis North Channel" and which in turn resulted in strengthening of the surface current in the north-western Andaman Sea and brought in substantial quantity of sediments from the NE Bay of Bengal. We envisage that the observed increase in contribution of the Indo-Burman Range sediments to our core site could have occurred either by removal of already deposited sediments from the western shelf of Myanmar in the NE Bay of Bengal (Fig. 4.25) by the easterly flowing surface currents, or by direct sediment supply from the Indo-Burman Ranges to the core site through the Irrawaddy river system, or by both. This contribution from the Indo-Burman Ranges, however, becomes less pronounced in the sediments deposited in the eastern and southern Andaman Sea (Fig. 4.24). The isotopic compositions of the sediments in the core MD77-169 from the southern Andaman Sea have been interpreted to represent the more radiogenic Sr bearing sediments of the Irrawaddy (Colin et al., 1999). We, therefore, rule out the possibility that the Irrawaddy river system ever carried isotopically non-radiogenic (Sr) sediments derived from the Indo-Burman Ranges into the Andaman Sea. Arguably, the lowering of the sea level during the last glacial period had an important control over the dispersal of sediments in this region.

Chapter-5

Chronology of major terrace forming events in the Andaman Islands during the last 40 kyrs

5.1 Introduction

Subduction zones are known for their high seismicity, which varies both spatially and temporally. Seismic activities in such zones are result of movements along numerous thrusts faults, including the decollement, present between the trench and the end of the forearc. The movements along these faults play an important role in the development of various morphological features on both subducting and overriding plates (Ruff and Kanamori, 1980; Kanamori and McNally, 1982; Byrne et al., 1992; Nanayama et al., 2003). The recent 2004 earthquake of Sumatra and 2011 earthquake of Japan were results of such type of movements along the faults. These earthquakes and related tsunamis, which caused large-scale devastations, will remain in our memories for a very long time. Such types of events reaffirm the need of thorough mapping of faults and understanding of tectonic processes those lead to such catastrophic natural events. The main difficulty in studying such faults in subduction zones is that these rarely have surface expressions and therefore, are inaccessible for direct geological investigations. In such a scenario, the only way one can infer about the tectono-geomorphic evolution of such regions, is by studying the displacements along secondary or tertiary faults as recorded in the resultant morphology on surface.

Andaman subduction zone is a part of the same Sunda-Banda subduction zone, which has witnessed some of the high magnitude earthquakes (M > 7.0) of recent times including the 2004 Sumatran earthquake of M9.1. The Andaman and Nicobar Islands, represents the outer-forearc portion of the subduction zone, provide a unique opportunity to study and understand many of the tectono-geomorphic features developed in a convergent margin. These islands are structurally made up of several eastward dipping thrust and strike-slip faults those run parallel to the Andaman trench and large scale folds (Fig. 5.1a) (Allen et al., 2007). However, except for some preliminary work (Roy and Chopra, 1987; Roy, 1992) little is known about

seismicity of these faults. As per current understanding some of these faults were responsible for those changed the surface morphology of the islands, details of which are summarized in Rajendran et al. (2007) and are shown in Fig. 5.1(b).



Fig. 5.1: (a) Map of the Andaman Islands showing the study areas-1 and 2 which are Lamiya Bay in North Andaman and Radhanagar Beach on Havelock Island and major faults, EMF- East Margin Fault; DF- Diligent Fault; WAF- West Andaman Fault (modified after Curray 2005). The solid red line is pivot line of Meltzner et al., (2006); (b) Map showing known historic large magnitude (M > 6.0) earthquakes in the Andaman region with their magnitudes (data from Rajendran et al., 2007). Ms: surface wave magnitude; Mw: moment magnitude scale.

The record shows that many a times these events have resulted in upliftment and/or subsidence of land in the forearc region (Oldham, 1883; Jhingran, 1953; Rajendran et al., 2007). While the chronology of most historical events has been established, that for the prehistoric events remains largely unknown. Considering that such events are important in the study of geomorphology of this region it is imperative that their ages are accurately determined. In such an effort, we have studied morphological changes on the coastlines of the North Andaman and Havelock Islands (Fig. 5.1a) and dated dead coral reefs from raised marine coastal terraces.

Since these terraces are local in nature (100 - 200 m in length) and have variable heights on smooth coastlines, we believe that they are generated by tectonic forces, and not by the changes in the sea level. It is also possible that some of these terraces could have formed by multiple uplift events; in such a scenario the determined age would likely to represent the oldest event. By comparing our results from the above two localities with the existing information on other places in the Andaman and Nicobar Islands (Kayanne et al., 2005; Rajendran et al., 2007 and 2008; Malik et al., 2011) we have made an attempt to reconstruct the seismic history of the region during the last ~40 kyrs. Corals are sensitive to many oceanographic parameters such as temperature, salinity and pH and therefore, the causal link between the death of coral reefs on the raised terraces and the upliftment events may not be unequivocal. However, the fact that we have been able to correlate the formation of terraces studied in this work with several others in the region makes our inferences robust.



Fig. 5.2: Photographs showing (a) submerged forest at Chidiyatapu, South Andaman after 2004 earthquake and tsunami (b) uplifted beach with dead coral reef at Kalipur in North Andaman. The arrow shows present level of sea during low tide.

The M9.1 Sumatran earthquake of 2004 resulted in major morphological changes in the Andaman and Nicobar islands and provided an opportunity to observe the effects of high magnitude earthquakes on the geometry of accretionary wedge (Rajendran et al., 2007). Many studies reported upliftment (1-2 m) in the Middle and North Andaman Islands, and subsidence in the south (Kayanne et al., 2005; Malik and Murty, 2005; Thakkar, 2005; Ramesh et al., 2006; Rajendran et al., 2007) (Fig. 5.2a & b). Rajendran et al., (2007) give a summary of field observations on coseismic changes observable from coastal morphology. The upliftments are manifested in

form of elevated shorelines/coral beds/mangrove swamps. Som et al. (2009) calculated an uplift of 31.21 cm from coral microatolls on Kalipur Beach, which is consistent with the recorded uplift elsewhere in North and Middle Andaman Islands (Kayanne et al., 2007; Rajendran et al., 2007; Ray and Acharya, 2007). We too have observed these raised coral microatolls in Kalipur beach during our field trips in 2009. Interestingly, according to the residents of Kalipur these raised terraces were present even before the 2004 earthquake, which led us to suspect that either the inferences of Som et al (2009) were erroneous or that these terraces had seen multiple upliftment events. Further inland, many such dead coral reefs have also been found to form the terraces, on which thick vegetation has grown. Similarly, we have observed exposed coral reefs along the west coast of Havelock Island (Fig. 5.1a), which prior to this work, have not been linked to any past seismic event.

5.2 Coastal Terraces of Andaman and Nicobar Islands

The morphological changes that occurred during the great 2004 earthquake inspired many workers to search for evidences of such events in the geologic records of Andaman and Nicobar islands. In such an effort, Rajendran et al. (2007) identified numerous steplike older terraces in many islands in the Andaman and Nicobar along several profiles. They studied terraces in Interview Island, Avis Island, Car Nicobar and in the east coast of Hut Bay (Rajendran et al., 2007 and 2008). Based on C-14 ages of corals, shells, dead tree trunks from raised terraces and shallow pits it was observed that the Andaman region had experienced many major seismic events during the last 40 kyrs. The oldest evidence of upliftment comes from Car Nicobar Island where the topmost coral terrace is dated to ~cal yr BP 41,000. Evidences for younger upliftments come from Interview Island, Avis Island and Hut Bay (Table 5.1). Earthquakes not only resulted in upliftment of coastlines, but also subsidence, which is mainly observed in coastline surrounding Port Blair (Fig. 5.2a) (Rajendran et al., 2007; Table 5.1). Studying the coastal stratigraphy in excavated trenches near Port Blair, Malik et al. (2011) found evidences for two major earthquakes during the last 400 years with one of them being associated with a tsunami akin to that occurred in 2004. On basis of the tectonic history of the region, it is quite reasonable to assume that similar episodes, in the past, might have caused upliftment of coral

reefs and development of coastal terraces observed on the two localities studied in this work (Fig. 5.1).



Fig. 5.3: Elevation profiles of coral terraces showing sample locations and calibrated radiocarbon ages (a) East coast of Lamiya Bay, North Andaman Island (b) West coast of Radhanagar Beach, Havelock Island. Note that the profiles start from high-tide levels.

In the entire chain of islands of Andaman and Nicobar, modern alluvium, raised beaches, coastal terraces, wave cut platforms, coral rags, calcareous tufa, and shelly limestones have been reported from the coastlines (Rajshekhar and Reddy, 2003). In Chapter-2 we have characterized these under the Pleistocene-Holocene sediment deposits of the Nicobar Group. For the present work, we collected coral fragments from exposed reefs on the coastal terraces of Lamiya Bay at Kalipur in the North Andaman and from raised coral reefs at Radhanagar Beach of Havelock Island (Study Area-1 & 2, respectively; Fig. 5.1). Lamiya Bay, which forms a part of the eastern coastline of North Andaman, has well-preserved coastal terraces that we

named as LBT1, LBT2, and LBT3 in order of their heights from the high tide water line (Fig. 5.3a; Fig. 5.4a & b). The terrace LBT3 is about 2.5m above the high tide water line and 8m inland (Fig. 3a), and is covered with thick vegetation. It appeared to be the oldest terrace at this location. LBT2 is a flat pebble beach approximately 2m wide and about 1m above the high tide water line. LBT1 is made up of dead coral microatolls, and is exposed in the intertidal zone (10-100m wide). Further inland beyond LBT3, rocks of the Ophiolite Group and sedimentary formations of the Mithakhari and the Andaman Flysch groups are exposed. Our study area in the North Andaman was little affected by the 2004 tsunami, and hence there has been no report of any tsunami deposits along this coast. However, according to some studies the dead coral reef of LBT1 possibly represents an upliftment related to the December 26, 2004 M9.1 Sumatran earthquake (Som et al., 2009). To establish the chronology of these terraces, we collected three coral samples (LB-10-01, LB-10-02 and LB-10-03) for radiocarbon dating. These coral samples represent the dead reefs present on the terraces, except for LB-10-01 which was from a large coral boulder located on LBT3 and appeared to have been deposited by a large wave/tide action or a tsunami.

Samples of coral fragments were also collected from a beach and an elevated terrace, at Radhanagar, on the west coast of Havelock (Fig. 5.1). Two coral terraces RBT1 and RBT2 were identified at Radhanagar (Fig. 5.3b, 5.4c & d). Unlike the terraces of Lamiya Bay, at Radhanagar the two terraces are separated by an almost vertical fall of ~3.5 m (Fig. 5.3b, 4c). RBT2 is composed of alternating layers of yellowish sandy silt and foraminiferal limestone. These rock formations belong to the Archipelago Group, above which beds of a dead coral reef are present (Fig. 5.4c). A sample of this dead coral was taken for radiocarbon dating (AND-09-68), from about 2.2m above the high tide water line. Modern soil and vegetation occupy the terrace RBT2 beyond ~20cm from the dead reef line.

5.3 Results and Discussion

The calibrated ages obtained from two dated sections are given in Table 5.1. The dating results are also presented in Fig. 5.3(a) and (b) while interpretating these calibrated ages we have used modes of the calibrated probability density curve,

rounded off to the nearest hundred. At Radhanagar beach, Havelock Island (Fig. 5.3b), the terrace RBT2 is dated to ~cal yr BP 7800, while the terrace RBT1, that occurs about 2m below RBT2, is dated to an older age of ~cal yr BP 8600. At Lamiya Bay, North Andaman (Fig. 5.3a) corals from the lowermost terrace LBT1 (Fig. 5.3a and 5.4b) yielded an age of ~cal yr BP 6300. Interestingly, this is the same terrace that is believed to have been formed during the earthquake of December 26, 2004 exposing the coral reef (e.g., Som et al., 2009). The terrace LBT2 (Fig. 5.3a) could not be dated, as it did not contain any dateable material. Two samples from the uppermost terrace (LBT3), within the forest, at a height of 2.5m above the high tide water line yielded ages of ~cal BP yr 8100 and ~cal yr BP 500 (Fig. 5.3a). Repeated analysis of another fraction of the latter yielded an age of ~cal yr BP 600. The presence of much younger coral on the terrace that appears to be at least 8100 years old is quite perplexing.



Fig. 5.4: Photographs of the studied terraces: (a) & (b) LBT1, LBT2 and LBT3 at Lamiya Bay, North Andaman (c) & (d) RBT2 and RBT1 at Radhanagar Beach, Havelock Island. The coral sample at Havelock (c) was collected from a unit exposed above the dotted green line, below which lies the calcareous formation of Archipelago Group. The dead corals on the beach terrace RBT1 (d) are exposed in the intertidal zone.

RBT2 at Havelock and LBT3 at Lamiya Bay are elevated to almost identical heights and have same age (within error) even though they are 135km apart. Both these terraces are co relatable to uplifted terraces of similar age from Little Andaman and Interview Island (Table 5.1). The vertical movements of terraces usually occur along fault planes, and a large part of the coastline of the Andaman and Nicobar Islands is deformed by numerous such faults. Pal et al. (2003) reported N-S trending thrust faults in North Andaman and Curray (2005) reported a fault along the west coast of Havelock Island. It appears that LBT3 and RBT2 (Fig. 5.3) have been developed due to reactivation of these faults, whereas RBT1 and LBT1 probably owe their origin to a couple of blind faults. There is no information on uplift/subsidence caused by these faults in literature, however there have been reports of several seismic events in this region that appear to have concentrated around these faults (USGS, see website: http://earthquake.usgs.gov/earthquakes/ eqarchives/epic/. Som et al. (2009) studied uplifted and highly eroded coral beds covered with moss and algae, at Hut Bay, Little Andaman and inferred that these corals were already dead before their upliftment by the 2004 earthquake. He also observed signatures of palaeoseismicity in the coral reefs of North Andaman and Little Andaman and suggested that the coral reefs of the Andaman and Nicobar Islands have been tectonically disturbed by several cycles of seismicity. The same might be true at our studied site also and perhaps, the corals from our terraces LBT1, LBT3 at Kalipur beach and RBT1, RBT2 at Radhanagar beach, Havelock were already dead due to upliftment caused by past tectonic events between 6 to 9 kyrs BP and further uplifted to current heights by later tectonic events.

The presence of ~500 (or ~600) cal yr BP old coral on ~8100 cal yr BP old terrace cannot be explained by an earthquake or changes in the sea level. Its occurrence on thickly vegetated terrace in form of a large boulder of coral made us believe that it was deposited by a tsunami. Several recent studies in the Andaman and Nicobar Islands have been able to discern evidences for tsunamigenic earthquakes in the recent history (Rajendran et al., 2006, 2007 and 2008; Kunz et al., 2010; Malik et al., 2011).

Table 5.1: Radiocarbon ages of fossil coral terraces of the Andaman and Nicobar Islands.

Sample names	¹⁴ C age (vrs BP)	Calandan aga ^c	Flovation (m) Material datad		
Sample names	$\pm 1\sigma$ (yrs BP) $\pm 1\sigma$		asl		
Uplift					
Interview Island					
IN/TOP/A	$30,880 \pm 300$	34925±239	50	Coral fragments, R(2008)	
IN/TOP/B	25021±280	29397±412	50	Coral fragments, R(2008)	
T2/IN/D/P1/A	$22,890 \pm 120$	27204±338	26	Coral fragments, R(2008)	
T3/IN/D/P1/A	19,894±100	23247±241	18	Coral fragments, R(2008)	
T3/IN/D/P1/C	19,304±220	22588±364	18	Coral fragments, R(2008)	
T3/IN/D/P1/D	$16,849 \pm 130$	19623±194	13	Coral fragments, R(2008)	
IN/St/D	6977±85	7479±78	7	Coral fragments, R(2008)	
Hut Bay (Little Andaman)					
HB/L2/CP	3286±100	7287±133	4	Coral fragments, R(2008)	
T2(B)/HUT/D/P4	6775±135	3105±142	4.5	Coral fragments, R(2008)	
T2/HUT/D/P1	3661±133	3551±168	2.8	Coral fragments, R(2008)	
T2(A)/HUT/D/P3	3123±47	2881±80	2.8	Coral fragments, R(2008)	
T2/HUT/D/P5	2295±62	1897±87	2	Coral fragments, R(2008)	
T2(A)/HUT/D/P1	1904±61	1437±78	2	Coral fragments, R(2008)	
T3/HAR/HUT	1843±77	598 ±71	0	Coral fragments, R(2008)	
HB/L1/CP	1042±90	1380±87	1.7	Coral fragments, R(2008)	
Avis Island					
AV/T2/E	2450±84	2081±123	2.5	Coral fragments, R(2008)	
AV/T2/B	1942±114	1477±133	2.5	Shells, R(2008)	
AV/T2/C	1678±89	1215±97	2.5	Shells, R(2008)	
AV/T2/A	1628±94	1168±101	2.5	Coral fragments, R(2008)	
AV/T2/D2	1364±126	898±137	2.2	Coral fragments, R(2008)	
AV/T2/E	1319±91	839±98	2	Coral fragments, R(2008)	
AV/T2/D	1172±42	706±48	1.8	Coral fragments, R(2008)	
AV/T1/PB	967±104	562±80	1.5	Coral fragments, R(2008)	
AV/T1/A	906±108	522±95	1.5	Shells, R(2008)	
AV/PD/LT/1	623±78	242±113	0.75	Coral fragments, R(2008)	
Lamiya Bay (North Andamar	ı)				
LB-10-01*	870±80	476 ± 78	2.5	Coral fragments	
LB-10-01* (Repeat)	1040 ± 80	599 ± 65	2.5	Coral fragments	
LB-10-02*	7650±100	8086±115	1.8	Coral fragments	
LB-10-03*	5880±100	6290 ± 106	0.6	Coral fragments	
Havelock Island				-	
AND-09-68	7320±100	7774±106	2.2	Coral fragments	
AND-09-69	8120±110	8585±155	0.3	Coral fragments	
Car Nicobar					
CN/TC/1	5420±80	5786±97	13	Coral fragments, R(2008)	
CN/SW/1	2750±130	2490±166		Coral fragments, R(2008)	
CN/SA/1	35,910±1040	40471±1059		Coral fragments, R(2008)	
Subsidence/Palaeotsunami					
Port Blair (South Andaman)					
RG/Wood/1	740±100	666±104		Wood, R(2007)	
	5903±35	6722±49		Plant debris, R(2007)	
	6283±25	6720±56		Shell, R(2007)	
	3070±120	3258±178		Peat, R(2007)	
RG/Peat/2	4320±130	4958±309		Peat, R(2007)	
Hut Bay (Little Andaman)					
GS/HUT/OT	753±35	378±56		Shell, R(2007)	
CS/HUT/OT	5623±35	6006±75		Coral fragments, R(2007)	

rable J.r. commune	Table	5.1:	continued
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Sample names	¹⁴ C age (yrs BP)	Calendar age ^c	Elevation (m) Material dated
-	$\pm 1\sigma$	(yrs BP)±1σ	asl	
East coast India				
MB/AR/UL	955 ± 30	862±62		Charcoal, R(2006)
MB/T1/BL	1581±35	1468 ± 51		Charcoal, R(2006)
MB/AR/BL	1674±30	1573±34		Charcoal, R(2006)
Sumatra				
В	550-660	605±55		M (2008)
С	960-1170	1065±105		M (2008)
Thailand				
В	500-650	575±75		J (2008)
С	2200-2400	2300±100		J (2008)
Red Skin Island				
Hv 25468	840 ± 23	466±32	?0.3	Shell, K (2010)
Hv 25465	2955 ± 40	2726±51	-0.9	Shell, K (2010)
Hv 25466	$3,220 \pm 23$	3015 ± 70	-1.3	Organic material, K (2010)
Hv 25467	$2,935\pm40$	2691±74	-1.3	Shell, K (2010)
North Cinque Island				
Hv 25462	710 ± 23	345±49	?0.5 asl	Shells, corals, K (2010)
Hv 25460	810 ± 43	441±54	?2.6 asl	Corals, K (2010)
Hv 25471	695 ± 23	331±49	?2.6 asl	Conch, K (2010)
Hv 25591	$1,740 \pm 23$	1283±39	?2.3 asl	Corals, K (2010)
Liquification feature,				
Diglipur (North Andaman)				
MN/Sill1/S3	1050±100	934±135		Peat, R(2007)

asl: Above mean Sea Level, 'c' stands for calibrated ¹⁴C ages, Data with asterick '*' are results from this study while other data are **R (2006)**: Rajendran et al. (2006); **R (2007)**: Rajendran et al. (2007); **R (2008)**: Rajendran et al. (2008); **M (2008)**: Monecke et al. (2008); **J (2008)**: Janakaew et al. 2008; **K (2010)**: Kunz et al. (2010). All the data used in this study are calibrated using radiocarbon calibration program CALIB 6.0 (Hughen et al., 2004).

From the existing record of upliftment, subsidence and palaeotsunami in this region (Table 5.1), we have observed that there was indeed an earthquake at \sim 500 (or \sim 600) cal yr BP and that it affected numerous coastlines in the Andaman Sea. We believe that this earthquake generated a large tsunami that was responsible for destruction of coral reefs in the Andamans.

Although a large portion of the Andaman Islands show evidences for coseismic emergence caused by the 2004 earthquake, Port Blair, located on the eastern margin of South Andaman, subsided by ~1 m (e.g., Rajendran et al., 2007; Malik et al., 2011). Earlier studies on dead mangroves and peat layers in core samples from sites near Port Blair have identified four events of past submergence/subsidence at ~cal yr BP 700, 3300, 5000 and 6700 (Table 5.1). These results suggest that there have been overall subsidence of South Andaman Island during the last 6700 years as a result of

seismic activity. Meltzner et al. (2006) defined a pivot line separating the areas of coseismic uplift and subsidence those resulted from the 2004 earthquake (Fig. 5.1a.). On the basis of Geological and Historical data and data from the 2004 earthquake we clearly suggest that most of the past upliftment/subsidence events took place along the same "pivot line" indicating continuous subsidence of parts of South Andaman Island and upliftment of parts of North and Middle Andaman Islands.



Fig. 5.5: (a) Stacked histogram of number of uplift and subsidence events as afunction of time. (b) Relative probability density plot of the same events.

The age data of seismic events in the Andaman region, as inferred from uplift, subsidence and tsunami records, are summarized in stacked histogram and relative probability plots (Fig. 5.5). All the probabilities obtained from calibrated radiocarbon ages, given in the Table 5.1, have been summed up and normalized with

the total number of samples as per the procedure described in CALIB 6.1.1 (Reimer et al., 2009). The resultant curve in Fig. 5.5b represents relative probability distribution of all the events, with total area under the curve as unity. From the probability distribution we recognize 14 major events of seismicity at ~40.5, ~35.5, ~27.3, ~23.3, ~19.5, ~8.5, ~7.5, ~6.0, ~6.7, ~4.9, ~3.4, ~2.7, ~1.1 and 0.6 cal kyr BP. There is a gap in the dataset between ~19.5 cal kyr BP to ~8.5 cal kyrs BP, which could be either due to lack of data or a result of absence of any major seismic event during this period. The frequency of major seismicity appears to have revived at ~8.5 cal kyr BP and continued to increase through to present. The increased activity probably caused the major upliftment of terraces in Lamiya Bay, Havelock Island, Interview Island and Hut Bay. The reason for absence of any record of subsidence older than 7 cal kyr BP could simply be lack of data owing to difficulty in identification of evidences. Similarly, higher frequency observed for the younger events (events 13 and 14 in Fig. 5.5), could be a result of their proper preservation.

The evidences of tsunamis/earthquakes from Sumatra, Thailand, east coast of India and Andaman Islands although correlated tentatively, suggest that increased activity can lead to destructive tsunamis in century scale intervals. Such recurrence adds to the challenge of preparing communities along the northern Indian Ocean shorelines for future. Keeping in mind the destructions caused by December 26, 2004 earthquake and tsunami, we need to further refine our results and search for evidences of past tsunamis/earthquakes elsewhere along the coastal areas and establish chronology of past events.

Chapter-6 Summary and Conclusions

The present study is an attempt to understand the evolutionary history of the Andaman Subduction Zone (ASZ) by unravelling the evidences preserved in its sedimentary record. As mentioned earlier the ASZ and its southward extension to Sumatra, is one of the youngest and seismically most active subduction zones of the planet. It is also well known for its high magnitude earthquakes and large and destructive volcanoes. It preserves most of the tectono-morphological features expected in a typical ocean-ocean subduction zone setting. These have made the ASZ an ideal site to study tectono-sedimentary processes, volcanism and paleoseismicity related to subduction. This thesis work was initiated with the objectives: (1) to decipher the past volcanic activities in the Andaman region; (2) to determine the timing of deposition of various formations on the islands and sedimentation in the Andaman Sea (3) to determine sediment provenances; (4) to understand the role of climate and tectonics on sedimentation and (5) to understand the evolution of the Andaman region as a whole. In order to achieve our objectives we had studied the sedimentary records on the Andaman Islands and that in the Andaman Sea. Detailed fieldwork was carried out on the Andaman Islands and suitable rock samples were collected from various formations. For the study of Andaman Sea sediments, a core (SK-234-60) was raised from the western part of the basin. The major conclusions of this study are listed below with answers to the major objectives of the thesis.

To determine the timing of major volcanic activities in the region, we have studied ash deposits preserved in the sedimentary records. The oldest record of volcanism from the Andaman region comes from the tephra interbedded in the Eocene age sedimentary rocks on the Andaman Islands. There exist numerous other tephra deposits in the Mio-Pliocene age rocks, which suggest enhanced volcanic activity during this time. In absence of evidences for existence of nearby arc volcanoes, the source(s) of these tephra deposits are difficult to interpret. The currently active inner volcanic arc, located east of the Andaman Islands, seems to be quite young and probably came into existence only after the opening of the Andaman Sea in the late Miocene. Towards this, we studied a sediment core (SK-234-60) from the Andaman Sea, in which we discovered seven discrete ash layers. Using isotopic and geochemical tracers we have clearly established that the Barren Island Volcano is the sole source to these ash layers. From our study, it appears that the Barren Island Volcano was the only major eruptive centre in the Andaman Sea during the late Pleistocene and the Holocene and its debris covered an extensive area around the volcano. We reconstructed the eruptive history of this volcano by dating foraminifers (AMS ¹⁴C dating) in sediment layers. The seven ash layers present in the core found to represent major eruptions of the volcano at ~72, 71, 62, 24, 17, 12, and 8 ka. Through the study of these ash layers we made several other findings:

• The ash layers erupted from 72 ka through 17 ka have highly uniform ε_{Nd} composition, which indicates that the Nd isotopic composition of magma of the Barren Island had remained almost constant during this large time period.

• Since the ~12 ka eruption to the present the isotopic composition in magma has been highly variable as also observed in ε_{Nd} of historic and recent eruptions on Barren Island.

• Isotopically correlating the precaldera volcanics exposed on the volcano to the uppermost ash layer (AL-1) in the core, we infer that the caldera of Barren Island volcano is younger than 8 ka.

• We speculate that an eruption at ~ 62 ka, thickest in the entire record, with the coarsest particles, was quite large and during this time the volcano had grown to near sea level or above it.

• The ash layers in the core mostly consist of vesiculated lithic fragments that contained microcrystals of translucent plagioclase (labradorite and bytownite), pyroxene (augite, enstatite and diopside) and green olivine (fosterite), titaniferrous magnetite, spinel and amphibole embedded in a matrix of glass.

• Similar to most of the subaerial lava flows on the Barren Island Volcano, the glass matrix compositions of the lithic fragments represent sub-alkalic volcanic magma which is basaltic to andesitic in composition. The parental magmas during various eruptions in the past were more evolved than the modern flows.

• The glass matrix compositions of lithic fragments in a given layer suggest that those were derived from magma undergoing fractional crystallization during that particular volcanic event.

In order to understand the tectono-sedimentary processes occurring in the subduction zone environment we have studied the records preserved in the sediments from the Andaman Islands (Late Cretaceous to present) and the Andaman Sea (Pliocene to present). The study of sedimentary rocks from the Andaman Islands clearly suggests that the Mithakhari Group sediments, deposited during the early to middle Eocene, were derived predominantly from mafic igneous sources comprising suprasubduction ophiolites and volcanic arc rocks of the ASZ. We also found minor contributions from the Himalayan/Indian Shield sources to the basin during this time. In comparison to the Mithakhari Group sediments, the Andaman Flysch Group sediments deposited during the Oligocene appear to have been derived from mixed sources with dominance of Himalayan sources. Through the study of the sediments from the Andaman Islands we made several other findings as listed below.

• Geochemical results show that the sources for the Mithakhari sediments had undergone less weathering while the sources of sediment for the Andaman Flysch Group were highly weathered.

• The sediments to the Mithakhari Group did not get transported long distances and had not undergone much recycling and sorting before deposition, while the sediments contributed to the Andaman Flysch Group were transported to large distances before being deposited and had undergone substantial recycling and sorting.

• The sedimentary rocks from the Andamans show mixing of sediments derived from the Himalayas and Andaman arc/ophiolite sources. During the deposition of rocks of the Mithakhari Group, the local arc/ophiolite sources possibly contributed >80% of sediments, whereas the same sources contributed about 60-80 % during the deposition of the Andaman Flysch Group.

• We believe that the substantial increase in the sediment input from the rising Himalaya during the deposition of the Andaman Flysch Group was result of large scale weathering, erosion and transportation of sediments through the paleodrainage system developed along arc and suture zone. At this time (~40 Ma), probably major

thrusting events occurred in the Himalaya which provided the essential height to act as topographic barrier to the moisture-laden winds from the south. This resulted in the development of the first monsoon system which eventually led to increase weathering and erosion in the Himalaya.

In order to understand the impact of climate on weathering and erosion, and supply of sediments in the past, sediments in the core (SK-234-60) from the Andaman Sea were studied. From the study of these sediments we have been able to determine that the western Andaman Sea show relatively higher contribution of sediments from mafic sources of the Indo-Burman Ranges (mainly ophiolites), while sediments from the Irrawaddy river system dominate in the sediments deposited in the central and eastern Andaman Sea. The elemental and isotopic compositions of the sediments reveal significant variations in the relative supply of sediments from sources over glacial-interglacial timescale that correlate well with the variability of South-Asian monsoon. The changes observed reflect influence of climate on erosion in source areas and relative supply of sediments to sea. We made several other findings from the study of these sediments, as listed below.

• Significant increases were observed in the relative contribution of sediments from mafic Indo-Burman sources at \sim 8 kyr, \sim 20 kyr (LGM), \sim 36 kyr, \sim 44 kyr, \sim 52 kyr and \sim 58 kyr. We believe that these were related to the weakening of the Asian summer monsoon, which restricted material contribution from the Himalayan sources.

• Higher sediment contributions from higher Himalayas and continental Myanmar sources through Irrawaddy and Ganga-Brahmaputra rivers at ~6 kyr, ~10 kyr, ~15 kyr, ~46 kyr, ~54 kyr and ~60 kyr and ~72 kyr could have been resulted from intensification of Asian summer monsoon, which in turn could be correlated to the global events of warm climate during Pleistocene-Holocene transition, Bølling-Allerød (B/A) and Dansgaard-Oeschger (D-O).

• The increase in overall contribution of sediments derived from the Indo-Burman sources since the LGM is inferred to be related to the strengthening of the surface currents in the north-western Andaman Sea due to increase in the sea level after the LGM. This resulted in reopening of "Preparis North Channel" through which substantial quantity of sediments from the NE Bay of Bengal entered into the Andaman Sea.

To understand the pattern of earthquakes and history of deformation on the Andaman and Nicobar Islands, investigations were carried out along the uplifted coastlines of two islands. Based on radiocarbon ages of exposed coral reefs from studied sections, seismic history of the islands for past 9 kyr was reconstructed. From our study on the tectonically formed coastal terraces we made following conclusions.

- Earlier reports and our results reveal that the Andaman region had experienced a major earthquake and associated tsunami event at \sim 500 (or \sim 600) cal yr BP.
- Combining our data with the available data on such events in this region we have been able to determine that there have been at least 14 major landscape changing seismic events between \sim 40 kyr BP to present, with a hiatus between \sim 19.5 and \sim 8.5 cal kyr BP.
- We propose that in a similar fashion as observed subsequent to the 2004 earthquake, the Andaman Islands have been experiencing tectonic upliftments in the north and subsidences in the south, for the last ~40 kyr, along the so called "pivot line" proposed by Meltzner et al. (2006).

Scope for future work

Although the present work carried out on the sediments deposited in the Andaman forearc and backarc basins reveal many interesting aspects of this subduction zone and its evolution, there exist numerous gaps in our knowledge as our study area was very much restricted to the northern sector only. Also, our record has large time gap between 20 Ma and 70 ka. Therefore, a comprehensive understanding of the time evolution of the region would require studies in Archipelago Group and sediments from Mergui Basin. Besides this, future studies should also address the following topics.

• The stratigraphy of the Andaman and Nicobar Islands is still not well developed. There are many sedimentary units which are misidentified, unclassified or wrongly classified. Efforts should be made to improve this. • There are several field of mud volcanoes on the Andaman Islands constantly emitting gases, water and mud breccias along with clasts derived from subsurface lithological units. These clasts may represent formations not exposed anywhere on the islands. Charcterization of these clasts can improve our overall understanding of the geology of the region.

• Apart from those studied in this work, several other coastal terraces have been observed on different islands in Andamans. These can be further studied to refine the paleoseismic record in the region.

• The rocks of Nicobar Islands have been completely neglected in earlier studies including ours. Their study can throw more light on the evolutionary aspect of the ASZ.

• In order to study volcanic history of the region we need to do more work on the tephra interbedded with sedimentary formations of Andaman Islands.

• Although we made an attempt to reconstruct the evolutionary history of the Barren Island Volcano, using a sediment core from the Andaman Sea, we admit that data from a single core may not be sufficient to reconstruct the complete eruption history of the volcano. It is quite possible that many of the past ash eruptions were simply not recorded at our core site because of their dispersal in other directions. Therefore, more cores need to be studied to reconstruct volcanic histories of the Barren Island Volcano as well as of Narcondam.

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List of Publications

- Awasthi, N., Ray, J. S., Laskar, A. H., Kumar, A., Sudhakar, M., Bhutani, R., Sheth, H. C., Yadava, M. G., (2010). Major ash eruptions of Barren Island volcano (Andaman Sea) during the past 72 Kyr: Clues from a sediment core record. Bulletin of Volcanology 72, 1131-1136
- Sheth HC, Ray JS, Bhutani R, Kumar A, Awasthi N. (2010): The ongoing (2008-09) eruption of Barren Island volcano, and some thoughts on its hazards, logistics, and geotourism aspects, Current Science, 98(5), 620-626
- Sheth HC, Ray JS, Kumar A, Bhutani R, Awasthi N. (2011): Toothpaste lava from the Barren Island volcano (Andaman Sea), Journal of Volcanological and Geothermal Research 202, 73-82
- Awasthi Neeraj, Ray J.S., Laskar A. H., and Yadava M. G. Chronology of major terrace forming events in the Andaman Islands during the last 40 kyrs. Journal of Geological Soceity of India (*in press*)
- Ray J.S., Kumar A., Binusarma P.E., Bhutani R., Balakrishnan S., Awasthi N. and Shukla A.D. Origin of breccia in mud volcanoes of the Andaman accretionary prism: implications for slab contribution to mantle wedge (Under review)
- Ray J.S., Rao D.K., Kumar A., Patil D.J., Sudheer A.K., Deshpande R.D., Awasthi N., Dayal A.M. and Butani R. Origin of fluids in mud volcanoes of Andaman accretionary prism: implications for fluid migration in forearcs (Under review)

Abstracts in International and National Conferences:

- Awasthi Neeraj and Ray J.S. (2010): Eruptive history of the Barren Island Volcano during last ~70 ka, AOGS 2010, Hyderabad. (*Poster presentation*)
- 2. Awasthi Neeraj, Ray J.S., Laskar A. H., and Yadava M. G. (2011) Chronology of major terrace forming events in the Andaman Islands during the last 40 kyrs, ICAMG-7, NIO, Goa. *(Oral presentation)*
- Awasthi Neeraj, Ray J.S. and Rai V.K. (2012) Provenance of sediments in the Andaman Sea: Implications from Sr-Nd isotopic ratios variations, GSC-2012, University of Baroda, Gujarat. (Oral presentation)

 Awasthi Neeraj and Ray J.S. (2012) Provenance of siliciclastic rocks of Andaman accretionary prism: Implications from Sr-Nd isotopic ratio variations, 34th IGC, Brisbane, Australia. (Oral presentation)