Sedimentological and Geotechnical Studies of Coastal Sediments of Central Kerala

Thesis submitted to the

COCHIN UNIVERSITY OF SCIENCE AND TECHNOLOGY

By

Vinu Prakash

in Partial Fulfillment of the

Requirements for the Award of the Degree of

DOCTOR OF PHILOSOPHY

Under the Faculty of Marine Sciences

DEPARTMENT OF MARINE GEOLOGY & GEOPHYSICS

SCHOOL OF MARINE SCIENCES

COCHIN UNIVERSITY OF SCIENCE AND TECHNOLOGY

COCHIN – 682 016, INDIA

November, 2012

CERTIFICATE

This is to certify that the thesis entitled "**Sedimentological and Geotechnical Studies of Coastal Sediments of Central Kerala**" is an authentic record of the research work carried out by Ms. Vinu Prakash under my supervision and guidance at the Department of Marine Geology and Geophysics, Cochin University of Science and Technology, in partial fulfillment of the requirements for the degree of Doctor of Philosophy, under the Faculty of Marine Science, and no part thereof has been presented for the award of any degree in any University/Institute.

November, 2012

A C NARAYANA
Research Supervisor
(Formerly at Department of Marine Geology & Geophysics
Cochin University of Science & Technology)
Centre for Earth & Space Sciences
University of Hyderabad

DECLARATION

I hereby declare that the thesis entitled **"Sedimentological and Geotechnical Studies of Coastal Sediments of Central Kerala"** is an authentic record of research work carried out by me under the supervision and guidance of Prof. A.C. Narayana, formerly from Department of Marine Geology and Geophysics, School of Marine Sciences, Cochin University of Science and Technology, in partial fulfillment of the requirement for the award of Ph.D degree in the Faculty of Marine Science and no part thereof has been presented for the award of any other degree in any University/Institute.

Cochin-682016 **Vinu Prakash**
November, 2012

Acknowledgement

It is with all sincerity and high regards that I express my deep sense of gratitude to Prof. A C Narayana, Director, Centre for Earth and Space Sciences, University of Hyderabad, Hyderabad-500046 (formerly Professor, Department of Marine Geology & Geophysics, School of Marine Sciences, Cochin University of Science & Technology) for his meticulous guidance, pertinent encouragement, inspiring discussions and valuable suggestions throughout this investigation. I will ever remain grateful to him for his wise counsel, unfailing attention and for suggesting me an interesting problem and guiding through its solutions. I express my sincere gratitude for his valuable advices and motivation ever since my M.Sc days in Cochin University of Science and Technology.

*I greatfully acknowledge All India Council of Technical Education for the junior research fellowship through a research project. This work forms a part of the project entitled "**Geotechnical studies of marine sediments of Kerala Coast**" funded by the All India Council of Technical Education, to my research supervisor.*

I express my sincere gratitude to Dr. K Sajan, Head of the Department (in charge), Dr. P. Seralathan, former faculty, Dr. C.G. Nambiar, Dr. M Ravisankar, Dr. K.R Baiju, faculty of Department of Marine Geology & Geophysics, for their help and co-operation. I am extremely thankful to Prof. K Sajan, Dean, School of Marine Sciences, Cochin University of Science and Technology, for the general facilities of the school. Director, School of Marine Sciences is also acknowledged for the facilities provided during the course of this research work.

I record my respectful thanks to Prof. N. Chandramohanakumar, Chemical Oceanography Department, School of Marine Sciences for his valuable advice and Prof. Philip Kurian, Controller of Exams, Cochin University of Science and Technology, for his inspiration and cooperation during the course of this study.

I express my sincere gratitude to Drs. S.K Bera, and C. M. Nautiyal, Birbal Sahni Institute of Palaeobotany, Lucknow for palynological analysis and ^{14}C dating. I also express my sincere gratitude to Drs. V. Balaram, T. Gyaneshwar Rao, P.K. Govil, and Mr. Keshav

Krishna, Geochemistry Division, National Geophysical Research Institute, Hyderabad for providing the ICP-MS facilities. I place on record my heartfelt thanks to Mr. Girish Prabhu, National Institute of Oceanography, for the XRD facility.

I am extremely thankful to the Centre for Earth and Space Sciences, University of Hyderabad for the facilities provided in the final stages of the thesis.

I gratefully acknowledge the help received from Late Sri Shaiju S, Mr. Subeer A. of Cochin University of Science and Technology and Mr. Pawan Kumar Gautam, JRF, Centre for Earth & Space Sciences, University of Hyderabad.

I am indebted to Ms N Sravanti, and Mr. G Shiva Kumar research scholars of the University Centre for Earth and Space Sciences, University of Hyderabad for their invaluable co-operation in the time of need.

I owe deepest gratitude to my organization, Mar Athanasius College of Engineering, Kothamangalam, Kerala for giving permission to do research work and the support of my colleagues in the institution.

This work would not have been accomplished without the tremendous help and assistance rendered by colleagues, doctors. Priju C P, Shinu N, and Sheena V. Dev. I also thank Dr. Nisha N R, Dr. Lasitha S and Mr. Mohammed Mashood P.A for their cooperation and help.

I thank the non-teaching staff of the Department of Marine Geology and Geophysics, Director's office, the staff members in the Central and Marine Science libraries and academic section of Cochin University of Science and Technology for their help.

I wish to thank my parents, my brothers, and my sisters who raised, supported, taught, and loved me. I also wish to thank my in-laws who gave me immense inspiration and moral support.

Last but by no means least, I am immensely indebted to my beloved husband and my daughter who filled me with courage at times of despair and frustration and their love and affection always lighted my path. To them I dedicate this thesis.

Vinu Prakash

Contents

Chapter 1

INTRODUCTION

Contents

1.1 General Introduction

According to the predictions the global sea level will rise between 0.2 and 0.5 m during the coming c 100 yrs as a consequence of global warming (Church et al., 2006; Bindoff et al., 2007). In order to improve the possibilities to predict future sea-level change, knowledge about both amplitudes and durations of eustatic fluctuations in the past is essential. During the last glacial maximum the eustatic sea level was between approx. 150 and 130 m lower than present (e.g. Colonna et al., 1996; Lambeck and Chappel, 2001; Ramsay and Cooper, 2002; Woodroffe and Horton, 2005). In the early Holocene, global sea levels rose rapidly as a response to increasing temperatures, glacial melting and larger volumes of water within the world's oceans. At c 8000-6000 cal BP, the sea reached levels similar to today (e.g. Zwartz et al., 1998). At many sites worldwide, this was followed by a major transgressive trend, with a rise in sea level above modern levels that affected earlier glaciated areas where isostatic uplift had been substantial (e.g. Svendsen and Mangerud, 1987; Risberg et al., 1991; Berglund et al., 2005; Mann and Streveler, 2008).

All over the world, several Quaternary proxy data have been used to reconstruct past sea levels, mainly radiocarbon or OSL dating of exposures of marine facies or shore line indicators (e.g. Carr et al., 2010) as well as paleo-environmental indicators in lagoon or estuary sediments (e.g. Baxter and Meadows, 1999). Estuaries and deltas develop at river mouths during transgressive and regressive phases, respectively (Boyd et al., 1992). In particular, the postglacial Holocene sea-level rise has contributed importantly to the estuary-to-delta transition (Hori et al. 2004). By analyzing radiocarbon ages of the basal or near-basal sediments of the world's deltas, Stanley and Warne (1994) showed that delta initiation occurred on a worldwide scale after about 8500–6500 years BP and concluded that the initiation was controlled principally by the declining rate of the Holocene sea-level rise.

Worldwide there were different regional sea-level changes since the last glacial maximum (LGM) (Irion et al., 2012). Along the northern Canadian coast, for example, sea level has been falling throughout the Holocene due to the glacial rebound of the crust after the last glaciation (Peltier, 1988). This is comparable to the development in Scandinavia (Steffen and Kaufmann, 2005) where sea level drops today. From about Virginia/USA to Mexico there is a constant sea-level rise similar to the Holocene sea-level development of the southern North Sea (e.g. Vink et al., 2007). From the border of Ceará/Rio Grande do Norte down to Patagonia, indicators of Holocene sea level point to a level that was up to 5 m higher than today's mean sea level (Angulo et al., 1999; Martin et al., 2003; Caldas et al., 2006a, b)

The nearshore zone is a highly energetic environment affected by surface gravity waves and mean currents. The nearshore environment is also a transition zone characterised by on-offshore sediment transport between the surf zone and the offshore zone. The exchange of sediment between the inshore and the offshore environment plays a major role on coastal evolution and shoreline stability (Swift, 1976). A good understanding of sediment transport processes in the nearshore zone is important for scientists and coastal zone managers. The reaction of sediments to hydrodynamics processes result in reworking and redistribution of sediments. It is of fundamental and scientific importance to understand how sediments from various internal and external sources on the shoreface are reworked and redistributed by the shoreface hydrodynamic processes (Liu and Zarillo, 1990).

The longshore current transports sediments that have been placed into motion by the waves, and being continuous along extensive stretches of the coast, can potentially move the sediment for many kilometers in the longshore direction. The longshore movement of sand on beaches manifests itself whenever this natural movement is prevented through the construction of jetties, breakwaters and groins. Such structures act as dam to the littoral drift, causing a build of the beach on its updrift side and simultaneous erosion in the downdrift direction. This often has severe consequences in the erosion of coastal property, and for this reason coastal engineers have shown a considerable interest in the quantities of littoral drift and the process responsible for littoral transport is also of interest to the geologist as a sand transporting agent and to the coastal sediment features.

The estuarine/lagoonal and nearshore sediment characteristics such as texture, geotechnical studies, clay mineralogy, organic matter, etc. throws light on various coastal and nearshore processes and these properties help to understand the sediment history during Holocene/Late Quaternary. The present study envisages understanding the sedimentary environment through physical, geotechnical, textural, and clay mineralogical studies of surface and core samples of central Kerala coast. In addition to the sedimentological and geotechnical studies 14C dating and palynological studies were done for the on shore borehole sediments drilled to a depth of 40 m for unraveling the geochronology, depositional history, sea level variation during the Late Quaternary and sediment provenance.

1.2 Objectives of the present study

The study area is the domain of natural sedimentary processes and different depositional environments. The proposed study is taken up to envisage the following aspects:

1. To understand the relationship between organic matter, textural, physical and geotechnical properties.
2. To understand the relationship between the clay minerals and geotechnical properties of sediments
3. To delineate the depositional environments in the coastal and marine settings.
4. To decipher the paleo-environmental conditions along the central Kerala coast.

1.3 Study Area

The study area comprises a part of Vembanad estuary/lagoon and nearshore region of Cochin and the study area is confined between latitudes $9^{o}45'$ - $10^{o}15'$N and longitudes 76^{o} -$76^{o}30'$E (Fig 1.1). The lagoon is connected to the Arabian Sea through a permanent opening, the Cochin bar mouth, which, is about 450m wide and 8-13 m deep. Here, the depth is maintained by dredging as this opening is used for navigational activities. The bar mouth is also responsible for the tidal flux of the Cochin backwater system. Tides are semidiurnal type, showing substantial range and time. The average tidal range

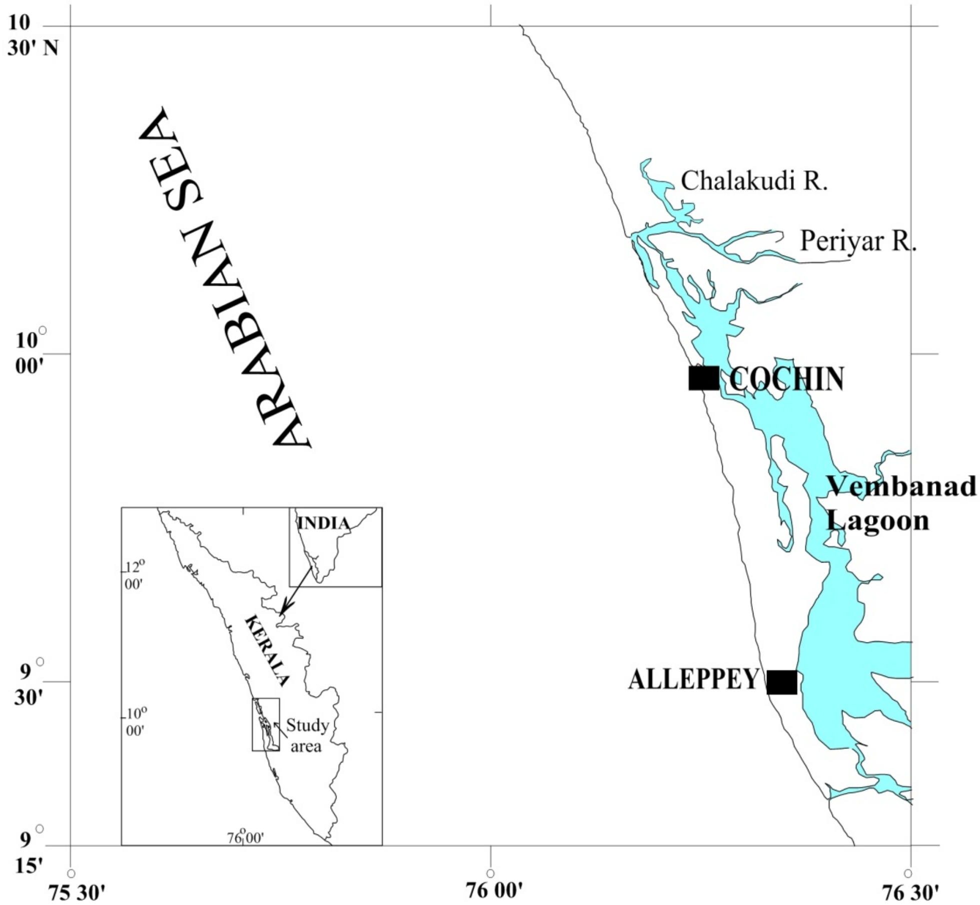


Figure 1.1: Study Area

near the mouth of the estuary is ~ 0.9m. Periyar river which is a part of the Vembanad lagoon has an outlet to the Arabian Sea at Munambam.

Six rivers, Periyar, Muvattupuzha, Chitrapuzha, Pamba and Achankovil debouch into the Vembanad Lake. Of these Periyar is the largest river, Hence Vembanad receives large amount of fresh water and sediments. Cochin backwaters and nearshore areas are separated by the barrier islands such as Vypin. There are a number of islands in the landward side of backwaters and these are Willington Island, Bolghatty etc.

Nearshore region off Cochin receives large flux of both water and sediment during the monsoon. The intense wave activity and longshore current play dominant role in the dispersal and distribution patterns of sediments that are brought from the land through the Vembanad estuary. A number of shoals are observed along the mouth of Cochin inlet.

Geomorphologic studies along the coastal tract between Thrissur in the north and Kollam Quilon) in the south, reveals several morphometric units that have a bearing on the Neogene evolution of this region. Of particular interest is the existence of a palaeodelta in the coastal zone near Munambam, extending from the present shoreline for about 11 km into the inland region along the course of the Periyar River (Narayana et al., 2001a). A borehole drilled to a depth of 40 m from this paleo-delta is also used for the present study.

1.4 Previous Studies

Nair and Sankar (1995) has classified and evaluated the coastal wetlands of Kerala using IRS-1A LISS II data. Nair and Nalinakumar (1997) has discussed about the sediment deposition on a microtidal environment using remote sensing data. Hashimi (1981) has done a comparative study of the moment and graphic size parameters of the sediments of the western continental shelf of India. The studies reveal that the mean size and standard deviation may be calculated either by the moment method or by graphic methods without any significant differences, as the methods is highly correlated. The distribution and textural variation of sediments off West Coast of India have been investigated by various workers (Stewart et al., 1965; Kolla and Kidd, 1982; Hashimi et al., 1978; Nair et al., 1978). Topography and nature of sediments of western continental shelf of India were studied by Nair et al., (1978) and Hashimi et al., (1978). Hashimi and Nair (1981) carried out study on surfacial sediments of Karnataka coast. Studies on sediment transport on the continental shelf of Mangalore by Narayana and Pandarinath (1991) revealed that the transport direction is towards onshore where fluvial influence is absent and the direction is towards both onshore and offshore where the fluvial influence is prevalent.

Veerayya and Murty (1974) have studied the textural characteristics and distribution of bottom sediments of Vembanad Lake from Cochin to Alleppey with a view to understand the depositional processes operating in different parts of the lake. These studies have revealed that (1) In the southern half of the lake, coarser fractions are confined to the western margin of the lake and finer fractions occupy the rest of the area while in the northern half, finer fractions are restricted to the estuarine region and coarser fraction occupy the rest of the area including the rivers, (2) well sorted and negatively skewed sediments are present in the Muvattupuzha and Ittupuzha rivers and in neighborhood of these river mouths and some of the subsidiary channels while poorly sorted and positively skewed sediments are present in the rest of the lake. Seralathan et al (1993) have studied the distribution of sediment and organic carbon in the Cochin harbour area. The study reveals that the mud and the sandy mud are spread over most part of the area with patches of muddy sand and silty sand.

Nair et al (1993) has studied the distribution of sediment characteristics of Cochin estuary in relation to changing hydrography. They have observed a seasonal special grading of particles as sand, silt and clay in the estuary. The contents of organic carbon, phosphorous and iron in sediments were closely studied in relation to hydrographical changes and attempts were made to describe the textural distribution on the above basis. Northern and southern arms of the upper estuary were mainly composed of sand particles. The lower estuarine regions indicated seasonal abundance of sand during monsoons pointing out to bed load movement. Surficial sediments also indicated variations in texture resulting from detritus settlement influenced mixing conditions in the estuary. Priju (2004) has studied landform changes and sedimentological aspects of central Kerala coasts. Badarudeen (1997) has carried out sedimentological and geochemical aspects of mangrove sediments of this region. Narayana et al (2008) has studied the sedimentological and geotechnical aspects of mudbank sediments of Kerala coast. Holland and Elmore (2008) have studied about the heterogenous sediments in different coastal environments.

The sedimentary framework and the distribution of organic carbon have been studied in the rivers and estuaries of Kerala by many workers. Murthy and Veerayya (1972 a, b and 1981) made a preliminary study of organic carbon in the sediments of the Vembanad Lake. Their studies on the sediments of Vembanad lagoon revealed that the distribution of organic matter is strictly following the hydrographic features of the lake. The organic carbon content was found to vary between 0.06-3.48percent. They also reported that silty-clays and clayey silts had higher organic matter content than sand and silty sands. Several attempts were made to study the organic carbon distribution in the various deltas, estuaries and lakes in India. The distribution of organic matter in the Ashtamudi lake sediment was studied by Sajan and Damodran, 1981. Studies on the hydrodynamic sorting and transport of terrestrially derived organic carbon in sediments of the Mississippi and Atchafalaya rivers were done by Bianchi et al (2007). Goni et al (1997) have assessed the sources and importance of land derived organic matter in the surface sediments from the Gulf of Mexico. Eadie et al (1994) studied about the nutrient enhanced coastal productivity in sediments from the Louisiana continental shelf. The terrestrial sources and the export of particulate organic carbon in the Waipaoa

Sedimentary System, New Zealand have been assessed by Blair et al (2010). The transfer of erosion related organic carbon from land to ocean in the Waipaoa Sedimentary System, New Zealand was analysed by Brackley (2006).

Becker et al (2004) studied the geotechnical characteristics of post-glacial organic sediments in Lake Bergsee, southern Black Forest, Germany. Samples of marine deposits retrieved at an onshore old reclamation site on the north coast of Taipa, Macau were analysed for the geotechnical characterisization by Yan and Ma (2010). Soil classification based on the expansivity of different clay minerals has been studied by Sridharan and Prakash (2000). Marine geotechnical properties in the head of Zakynthos canyon, Greece were analysed with the application of computational intelligence tools by Ferentinou et al (2012). These studies provide background information for the assessment of submarine slope stability and also shed light on the sedimentary processes that take place on the seafloor. The changes in late Pleistocene–Holocene sedimentary facies of the Mekong River Delta and the influence of sedimentary environment on geotechnical engineering properties were studied by Truong et al (2011). They discusses the influences of sedimentary environment and conditions on geotechnical properties of the sedimentary facies. Correlation between clay mineralogy and Atterberg limits was done by Schmitz et al (2004). They stated that if a correlation between the Atterberg tests and the clay mineralogy would be available to engineers working in soil mechanics, an estimate of the changes in mechanical properties could be given when the changes in clay mineralogy are known. Depositional and geotechnical properties of marine clays collected from boreholes along onshore sites were studied by Liu et al (2011). Geotechnical properties of the Cassino Beach mud were analysed by Dias and Alves (2009). Hajjaji et al

(2010) have analysed the mineralogy and plasticity in clay sediments from north-east Tunisia.

All fine-grained materials with large surface area are capable of accumulating heavy metal ions at the solid - liquid interface as a result of intermolecular forces (adsorption). The pH values may dominate the adsorption processes of heavy metal cations. The heavy metals are completely released under extremely acidic conditions (Forstner and Whittmann. 1981). Organic carbon plays an important role in the dispersal pattern of many major and trace elements. Mineralogy and geochemistry of sediments of Muvattupuzha and central Vembanad estuary were studied by Padmalal (1992). Sediment characteristics in relation to changing hydrography of Cochin estuary were investigated by Nair et al (1993). Metal concentration in recently deposited sediments in Cochin backwaters were analysed by Nair et al. (1990) and Priju and Narayana (2004). Resmi (2004) has carried out trace element assessment from three aquatic environments such as mangroves, river and estuary of the central Kerala coast. Sarika (2005) has studied the trace element concentrations in the mangroves along the Kerala coast. Geochemical index of trace metals in the surficial sediments from the western continental shelf of India were assessed by Laluraj and Nair (2006). The spatial analysis of trace element contamination in the sediments of Tamiraparani estuary, southeast coast of India was done by Magesh et al (2011). The enhanced preservation of organic matter in fine-grained sediments can lead to trace metal enrichment in their authigenic phase either through complexation reactions or by increasing sorptive capacity of the sediments (Alberic et al., 2000). The heavy metals of the surface sediments of the Portugese continental shelf were assessed by Mil-Homens et al (2006). The distribution and enrichment of trace metals in marine sediments of Bay of Bengal, off Ennore, south-east coast of India was analysed by Raj and

Jayaprakash (2007). The organic carbon, trace element and $CaCO_3$ variations in a sediment core from the Arabian Sea were studied by Shetye et al (2009). The anthropogenic pollution in the surface sediments of Izmit Bay was assessed by Yasar et al (2001).

Physical sorting of river-transported material is a well-known process documented in sediments at the fluvial/marine interface (Mitchell and West, 2002; Hori and Saito, 2007). These processes also occur in clay-mineral suspensions, as the coarse-grained illite and kaolinite particles are preferentially deposited at or near to the fluvial/marine interface, in contrast to the smaller grains of smectite, which may pass into the sea (Meunier, 2005). The composition and origin of clay minerals in Holocene sediments from the south-eastern North Sea was studied by Zuther et al (2000). The varied pathways of river-borne clay minerals in the near-shore region of south-eastern North Sea was analysed by Pache et al (2008). Provenance and distribution of clay minerals in the sediments of the western continental shelf and slope of India was studied by Rao and Rao (1995). The reconstruction of late Quaternary monsoon oscillations based on clay mineral proxies using sediment cores from the western margin of India was done by Thamban et al (2002). The pathways of clay mineral transport in the coastal zone of the Brazilian continental shelf were studied by De Morais et al (2006). The preferential settling of smectite on the Amazon continental shelf was analysed by Patchineelam and De Figueiredo (2000).

Paleooceanic conditions along the Periyar river mouth, Late Quaternary peat deposits in the sediment records of Vembanad lagoon and evolution of central landforms and associates sedimentary environments were discussed in detail (Narayana et al., 2001b; Narayana et al., 2002; Narayana and Priju, 2004). Verma and Mathur (1979) have shown the evidences of palaeoshorelines along west coast of India, which existed several kilometers inland from the present coastline and at much higher levels than the present sea level as evidenced by coral reefs and fossiliferous beach-rocks. The presence of the submerged shorelines was also indicated by submerged forests and occurrence of oolitic limestone at considerable depth from the present sea level. Studies on Late Quaternary sediments and sea level changes of the central Kerala, India has been done by Shajan (1998) with special emphasis on archaeological aspects. Evolution of the coastal wetland systems of SW India during the Holocene were studied by Padmalal et al (2011). Thrivikramji et al (2007) have attributed the evolution of the wetlands of Kerala coast through gradual regression of the sea after the Holocene marine transgression. The occurrence and water resource potential of fresh water lakes in south Kerala and their relation to the Quaternary geologic evolution of the Kerala coast were studied by Soman et al (2002). Joseph and Thrivikramji (2002) assessed the implications of the Kayals of Kerala coastal land to the Quaternary sea level changes.

Changes in Holocene coastal paleo-environment and sea-level variations have been recorded in estuary sediments from Macassa Bay, southern Mozambique (Norström et al., 2012). The Early Holocene history of the Baltic Sea was studied by Berglund et al (2005) from the coastal sediments of south eastern Sweden. The Holocene sea-level fluctuation was inferred from the evolution of depositional environments of the southern Langebaan Lagoon salt marsh, South Africa by Compton (2001). Late Quaternary sea-level changes in South Africa were studied by Ramsay and Cooper (2002). Angulo et al (2006) critically reviewed the mid to late Holocene sea-level fluctuations on the eastern Brazilian coastline. The Late Quaternary evolution of coastal and lowland riverine plains of southeast Asia and northern Australia was studied by Woodroffe (1993).

Stiros *et al.* (2000) opined a seismic coastal uplift in a region of subsidence from their studies on the Holocene raised shorelines of Samos Island, Aegean Sea, Greece. The evidences indicate that the geomorphological evolution of Samos island and of the wider Eastern Aegean is not only due to marine transgression and regional-scale tectonics but also due to earth quakes. Beets *et al*. (1992) studied the Holocene evolution of the coast of Holland. The Holocene regression and the tidal radial sand ridge system formation in the Jiangsu Coastal Zone, East China was examined by Li *et al*. (2001). Based on the facies succession, and the tidal sand units located either in the strand plain they explain that these sand ridges were formed and developed at different time scales. Dalrymple *et al.* (2007) have discussed the morphologic and facies trends through the fluvial–marine transition in tide-dominated depositional systems. Yoon et al (2009) studied a deep sea core from the Drake Passage west Antarctica to reconstruct the paleo-environmental conditions over the last 150 ka.

Mangroves are one among the world's most productive ecosystem and form an important part of the coastal and estuarine environment. Living at the interface between land and sea, the mangrove plants have morphological and physiological adaptations to survive in harsh saline environment. Mangroves produce organic carbon well in excess of the ecosystem requirement and contribute significantly to global carbon cycle. Mangroves flourish on fine alluvial mud composed predominantly of silt and clay particles. Geological history and evidences show that mangroves appeared between Eocene and Oligocene period (30 - 40 million years ago). Plant remains or fossils of major mangrove genera like *Rhizophora, Nypa* and others from the peat sediments provide important clues in this matter (Subramanian, 2002).The World

Conservation Union's report on global status of mangroves lists 61 species (IUCN, 1983). Major mangrove species belong to less than 15 families, but the most frequent occurring mangroves belong to the *Rhizophoraceae, Sonneratiaceae,* and *Avicenlliacea.* The occurrence of peat deposits from various onshore locations along the Kerala coast has been reported earlier. The mangrove vegetation responses to Holocene climate change along Konkan coast of south-western India was studied by Limaye and Kumaran (2012).

****

Chapter 2

GEOLOGY AND GEOMORPHOLOGY OF CENTRAL KERALA COAST

Contents

2.1 Introduction

Kerala has many interesting geologic features in which exposures of exhumed lower continental crust, amphibolite–granulite transition zone; classical exposures of laterite, etc are important ones. The Tertiaries and the recent sediments of Kerala coasts rest directly upon the Archaean crystalline complex consisting of khondalite, leptynite, Charnockite, and mica-hornblende gneisses. The Tertiary formations include Warakalli deposits of variegated sandstones and clays, white plastic clays, carbonaceous clays, associated seams of lignite, and the Quilon formation consisting fossiliferrous limestone intercalated with thick beds of variegated sands and carbonaceous clays (Menon, 1966). Coastal regions are blanketed by the mudflats composed of clays, silty clays, and shell fragments. The most prominent of the mud flat lies in the south eastern side of the Vembanad Lake (Narayana et al., 2008).

The central Kerala region hosts geologic features such as coastal alluvium, Quilon and Warkallai formations, and geomorphic features such as backwater systems backed by lagoons, barrier-island complexes, tidal mudflats etc. The central coast of Kerala is remarkably straight and is believed to have originated as a result of faulting during the late Pliocene (Krishnan, 1968). The central Kerala coast is considered as the 'haffnehrung' type of coast with lagoons and back waters (Ahmad, 1972). The coast contains vast stretches of sandy flats interspressed with lagoons, estuaries, and low lying reclaimed beds. Forty one west flowing rivers, most of them of the type of mountain streams, flow from Western Ghats into backwaters and lagoon that skirts the coasts. Backwaters connected to the sea inlets, are influenced by the movement of sediment straight trending in a NNW-SSE direction. On the landward side of the coast, there is a series of laterite rocks backed by alluvial deposits (King, 1882; Menon, 1974).

2.2 Geology

Kerala State is an integral part of the peninsular shield bounded by the Western Ghats on the east and Arabian Sea on the west. It is mainly occupied by four major rock units - (i) Precambrian crystalline rocks - which include charnockites, garnet biotite gneisses, hornblende gneisses, khondalites, leptynites, cordierite bearing gneisses and other unclassified gneisses which occupy a considerable area of Kerala. (ii) Tertiary sedimentary rocks – which unconformably overlie the Precambrians which extend as a narrow belt along the major part of Kerala coast, comprises continental (Warkalli beds) and marine (Quilon beds) facies. Carbonaceous clays with lignite / coal seams, china clays and friable sandstones forms the continental facies and the marine facies composed of sandstones and carbonaceous clays with thin bands of fossiliferous limestones (Poulose and Narayanaswami, 1968) (iii) Laterites – are the third major litho-unit covering about 60 % of the surface of Kerala and (iv) Recent to sub recent sediments extending from Kasaragod in the north to Capecomorin in the south, which include fringes of coast parallel sand bars, sandy flats, alluvial sands and lacustrine deposits (Brunn and Nayak, 1980).

Tertiary rocks in the Quilon-Warkala area are divided into (i) Quilon beds - consisting of limestone and calcareous clay and (ii) Warakalli beds - made of sandstone and clay with lignite (King, 1882). King also states that the Warakalli beds must then be of the same age, or if different conditions of deposition and apparent overlap go for anything, of perhaps a later age. Jacob and Sastri (1952) found that the calcareous beds exposed in Quilon area continue subsurface into Chavara and their age is Burdigalian. Narayanan (1958) opined that the Warkalli and Quilon beds are different facies of rocks but more or less of same age. Desikachar and Subramanyan (1959) suggest that the Quilon and Warakalli beds continue far into the north up to Kainakari and Ambalapuzha based on their reconnaissance geological mapping of the sedimentary terrain and study of shallow boreholes of coastal Kerala.

Varadarajan and Balakrishnan (1976) suggested that the laterite along the coastal area formed part of the continental shelf and their occurrence as terraces of different elevations could be due to Neogene and Quaternary uplift. Desikachar (1976) identified four formations as (i) Mayyanad Formation, (ii) Azheekal Fromation, (iii) Ambalapuzha Formation and (iv) Kainakari Formation. Azheekal and Ambalapuzha formations are roughly equivalent to Quilon and Warkalli beds. Desikachar (1976) highlighted the hydrocarbon prospects of offshore Kerala Basin. Murty *et al.* (1976) had the opinion of greater sedimentary thickness in northwest of Alleppey and Ponnani on account of the extension of the Achankovil shear zone and Palghat gap faults into the basin. Bose *et al*. (1976) suggested that the deepest part of the sedimentary basin is between Chellanam and Chavara, and around Alleppey the thickness of sediments may be about 600 m. Rao and Datta (1976) opined that the Wrakalli beds could be the continental equivalents of Quilon beds, a view expressed earlier by Narayanan (1958) and doubted by King (1882).

Lineaments and fractures are the major features cutting across the Precambrian terrains of Kerala through which major rivers of Kerala like

Periyar, Achankovil, Pamba are flowing. Nair (1987) reports that among lineaments identified over the Kerala region, the fault lineament parallel to the coast, as well as at right angle to it is found to be neotectonically active and are controlling the configuration of shoreline primarily. The major shearzones of Kerala region are: E-NE trending Nilambur-Attapady-Kozhikode shearzone, E-W trending Achankovil shearzone, Periyar-Thamraparni shearzone, NNW trending Edamalayar shearzone, and SE trending Moyar-Bavali-Mercara shearzone. Minor fracture zones are widely reported all over the state by Geological Survey of India (1995). Central Kerala region consists of basic intrusive rocks, charnockites, khondalites, meta-sediments, Quilon and warakalli formations, laterites, and alluvium (Fig 2.1).

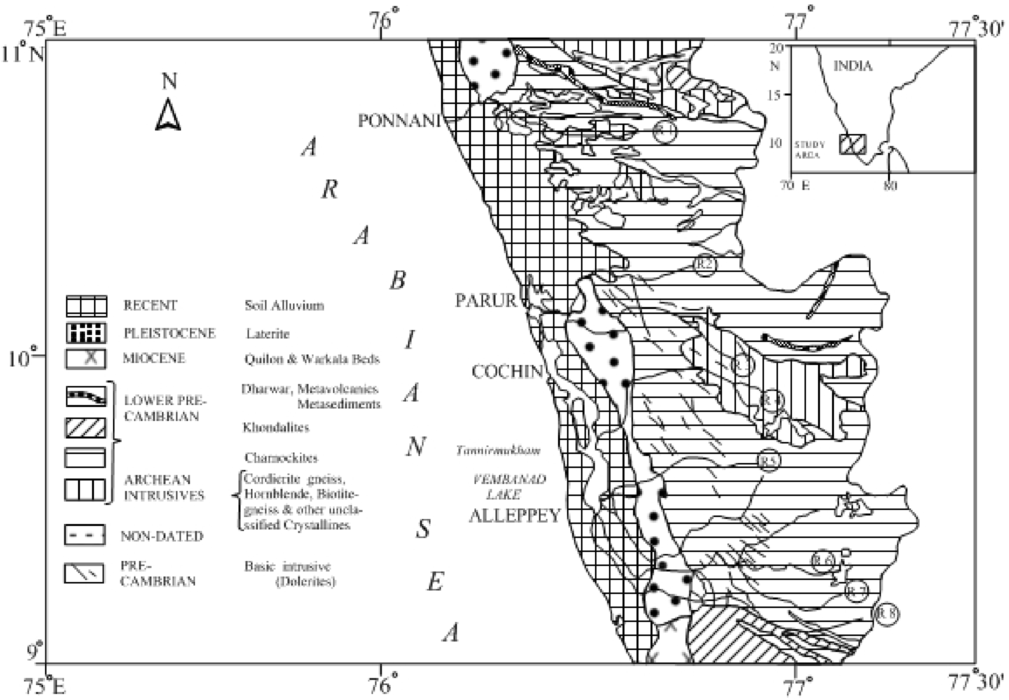


Figure 2.1: Map showing the geology of central Kerala region, and the rivers debouching into the lagoon and the adjoining continental shelf. R1, R2, R3, R4, R5, R6, R7 and R8 represent the rivers Bharatapuzha, Chalakudi, Periyar, Muvattupuzha, Minachil, Manimala, Pamba and Achankovil respectively. Inset shows the map area with reference to its location in India

2.3 Geomorphology and Coastal landforms

Highly irregular, cliffed and wave eroded (Ahmad, 1972) terrains are the geomorphic character of Western Ghats. Kerala has width from 35 to 120 kms with an average of 65 km. Within this small width, the physiographical and topographical features change considerably. It is divided into three geographical regions (1) Highlands (2) Midlands and (3) Lowlands. The Highlands slope down from the Western Ghats which rise to an average height of 900 m with a number of peaks well over 1,800 m in height. The Midlands lying in between the mountains and lowlands too made up of undulating hills and valleys. The lowland or coastal area is made up of river deltas, backwaters and the shore of Arabian Sea.

The characteristic landforms of Kerala coast are the backwaters, lagoons, barrier islands, beach ridges and swales, tidal flats/mudflats, strandlines and floodplains (Narayana and Priju, 2006). They also include lakes (kayals) and inlets, which stretch irregularly along the coast (Joseph and Thrivikramaji, 2002). The biggest one is the Vembanad Lake, with about 230 km^2 in area, opens into the Arabian Sea at Fort Cochin. The other important backwaters are Ashtamudi, Veli, Kadhinam-Kulam, Anjego, Edava, Nadayara, Paravur, Kayamkulam, Kodungallur and Chetuva. Coastal inlets play an important role in the exchange of water between bays/lagoons and ocean. There are about 48 inlets in Kerala, out of which 20 shows permanent nature of opening, whereas the remaining 28 open only during the monsoon season (Nair *et al*., 1993). Munambam inlet is a major permanent inlet just north of Cochin inlet, through which Periyar River joins into the sea. Islands and inlets are major landforms that occur along the lagoon.

Kerala Coastline is 560 kms long in which a cumulative 360 km length of coastline is very dynamic and fluctuates seasonally experiencing a moderate energy with monsoonal-storm-dominated wave climate. Based on vulnerability and dynamism, the shoreline is divided into (1) permeable, gently sloping sandy shoreline (2) semi permeable, cliffed, sedimentary shoreline and (3) impermeable crystalline shoreline arranged in order of intensity of erosion.

The shoreline of Kerala is generally straight, trending NNW-SSE, with minor variations. Even though a straight-line configuration is apparent in a synoptic view, the shoreline is highly irregular and indented especially around promontories comprising of crystallines and sedimentaries. Stretches of shoreline between promontories are usually depositional in nature with sandy beaches and are locally straight. Long shore drift deposits straighten out the shore features over long stretches (Nair, 1987)

The central Kerala coast is described as a submerged coast, falling under the terrigenous coast of primary morphologic disequilibrium. Accordingly the long-term tendency of coastal evolution is to increase the linear extent of erosional coastal segments at the expense of intervening depositional areas and when the process goes to completion, a coast of equilibrium results. The coastal features include beaches, beach cliffs, stacks, islands, shore platforms, spits, bars, beach ridges, estuaries, lagoons, mud flats, tidal flats and deltaic plains (Nair, 1987; Thrivikramaji, 1987). Some of them are briefly described here:

Barrier islands

Several barrier islands occur along the central Kerala coast of which Vypin island is a significant one. It is about 25 km long and acts as a barrier between the Vembanad lagoon and the Arabian Sea. Another wide shore-connected barrier island, separating the lagoon from the Arabian Sea, extends from Cochin to Cherthala.

Beach Ridges and Swales

A number of parallel beach ridges (also called as strandlines) alternating with swales are observed at many places (Nair, 1987). The width of ridges varies from 50 to 150 m and the height varies from 0.5 to 200 m. The width of the swales varies from 50 m to 200 m. These beach ridges represent successive still-stand positions of an advancing shoreline in relation to the sea. Occurrence of strandlines suggests that the coast has undergone a series of marine transgressions and regressions during the Late Quaternary. Significant changes were brought in the coastal configuration and associated landforms since mid-Holocene period. A number of strandlines, running parallel to the coast for a distance of 15 to 25 km occur up to 15 km inland from the present shoreline and about 2-5 m above sea level in the central Kerala region. Width of the strandline varies from 100 to 200m. These strandlines are present on either side of the Vembanad lagoon. Narayana and Priju (2006) have discussed about total obliteration of geomorphic features and sudden abutments of dune ridges with interdunal depressions near Parur, as reached from the tonal differences in the satellite imageries.

The occurrence of these paleo-beach ridges discussed above suggests the progradation of coastal land. This may be either due to fall in sea level or rise in the level of land or both. It is possible that both fall in sea level and uplift of the coast have influenced the formation of cheniers/strandlines along central Kerala coast. The sea level was higher during 3000 yr BP than the present day along this coast (Narayanan and Anirudhan, 2003).

Estuaries and Lagoons

Estuaries and lagoons (also known as kayals) are high in frequency on Kerala's landscape (Narayan and Anirudhan, 2003). These kayals are classified into two types as Type I and Type II. North of Kollam, Kayamkulam Kayal, and Vembanad lagoon, the largest coastal landform of the Holocene on the west coast, are Type I Kayals. The average length to width ratio of type I class kayals is 4.5. Among these kayals, Vembanad lagoon is the largest with a length of about 110 km and the width varies from a few hundred meters to 4.5 km. Depth varies from less than 1 m to 13 m. Type II kayals include kayals of Thiruvananthpuram-Kollam coastal stretch e.g. Vellayani Kayal, Veli Kayal, Kadinamkulam Kayal, Nadayara Kayal, Paravur Kayal, and Asthamudi Kayal. The average length to width ratio of Type II kayals is ~6.5. The evolution of the coastal lagoons has been influenced by the geological and geomorphological history of the coastal area and the sequence of changes in the levels of land and sea, which have resulted in coastal submergence and the formation of inlets and embayments (Bird, 2000). Subsidence of coastal regions may deepen and maintain coastal lagoons, delaying their infilling. The greater depth of the lagoon in the southern part, near Vaikom may indicate subsidence. The shallower depth in the northern part may be related to uplift. The morphology of the lagoon suggests that en-echelon faulting has played a role in its evolution (Valdiya and Narayana, 2007). The Vembanad lagoon at many places runs across the strandlines affecting their continuity.

Paleodelta

A paleodelta feature in the northern part of the study area extending from the present shoreline for about 11 km into inland region along the mouth of Periyar river is evident from satellite imagery (Fig. 2.2). The dull matground terrain devoid of any geomorphic feature is considered as a paleodelta (Narayana et al. 2001a), which encompasses an area of about 50 km^2. Older deltaic plains, formed as a result of coalescence of two deltas of Manimala and Pamba river extend from the coastal sandy plain to the western margin of the lowland region with an elevation of 10-300 m (Narayanan and Anirudhan, 2003).

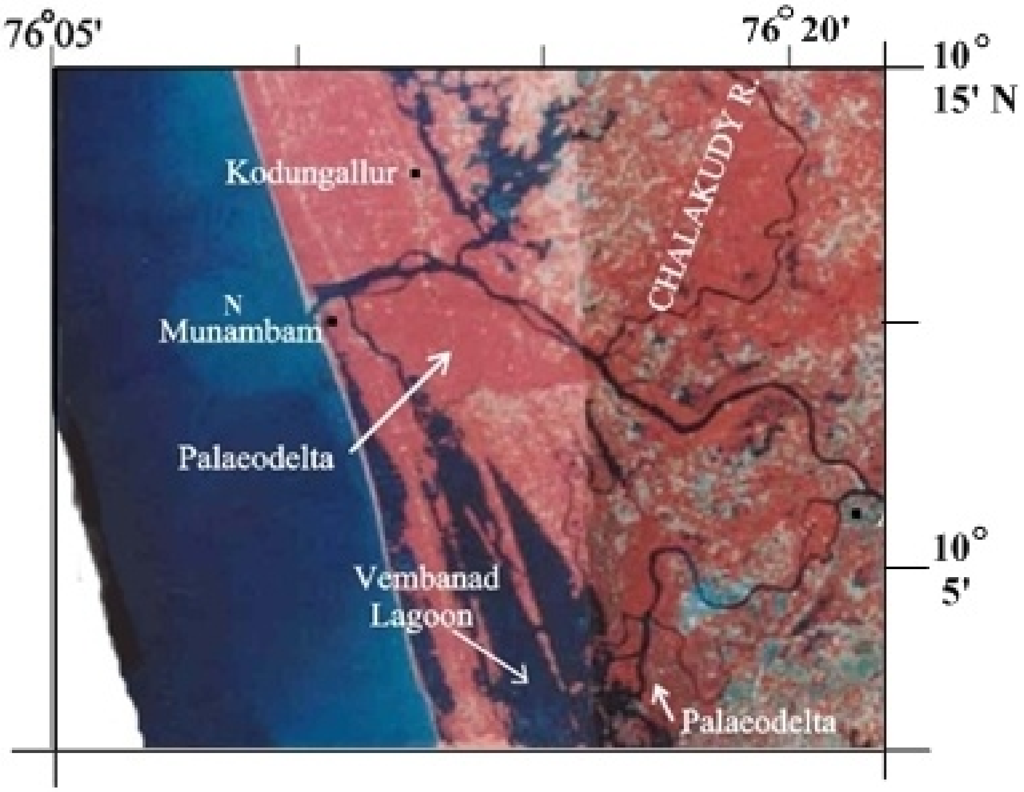


Figure 2.2: Satellite imagery showing the Munambam paleodelta and the associated coastal geomorphic features (from IRS 1B LISS II FCC bands 2,3, and 4 March 1995). New delta 'N' being built up by Periyar river (after Narayana et al., 2001 a)

Mudflats, Tidal Flats and Mangrove Swamps

Extensive tidal and mudflats are observed in the eastern part of Vembanad lagoon, particularly near the mouth of the southern branch of the

Periyar river and in the Chithrapuzha river mouth area. Most of the mud/tidal flats are covered with mangrove vegetation.

Flood plains

Extensive floodplains are seen in the lower reaches of Periyar and Muvattupuzha rivers. The eastern part of the Vembanad lagoon is covered with floodplains.

2.4 Rivers and Drainage system of Central Kerala

The State of Kerala gets drained by 44 rivers out of which 3 flow east-ward. The streams originating from the Western Ghats are short and swift flowing, showing various stages of gradation. Forty-one west flowing rivers drain across Kerala, with innumerable tributaries and branches, but these rivers are comparatively small and being entirely monsoon fed, practically turn into rivulets in summer, especially in the upper areas (*cf.* Resource Atlas of Kerala, 1984). These streams are marked by cascades and waterfalls in the upper reaches, although in the plains they show evidences of maturity of development. Some of these rivers have steep gradients (1/250 or more) in their initial reaches. In the case of Periyar and Chalakudi rivers, this extends for three-fourths of their course, while such gradients are also discernible in the upper reaches of Chaliyar, Valapattanam river, Vamanapuram Ar. Karamana Ar etc. suggesting their youthful stages of development (*cf.* Resource Atlas of Kerala, 1984). West coast faulting and later adjustments could be understood as the main evidence for the youthful behaviour of the rivers in region, while high energy shoreline might have prevented delta formation in the river mouths.

The general drainage pattern observed for the rivers of central Kerala is dendritic. However, trellis, sub-parallel and radial patterns are also noticeable in some places. River courses are mainly straight, indicating structural control due to prominent lineament directions (NW-SE and NE-SW) (Valdiya & Narayana, 2007). Among all the rivers, the Periyar, Pamba and Chalakudi rivers are conspicuous by their length and size of the area they drain.

Periyar is the longest river in Kerala with a length of 244 km and with drainage area of 5398 km^2 (*cf.* Water Resource of Kerala, 1974). The catchment area spreads over the districts of Idukki and Ernakulam. The river originates and flows through a metamorphic terrain consisting of charnockites, garnet-sillimanite-gneiss, garnet-biotite and hornblende-biotite gneiss, besides migmatite and granite. Periyar river branches off into northern and southern arms (Narayana et al.,2001a). The northern arm turns to the northwest and has an almost straight channel, suggestive of its more youthful character, and enters its own palaeodelta located in the south of Krishnankotta and further cuts across palaeodelta to enter directly to the Arabian Sea. The southern arm meanders westward and enters the yazoo zone which is the hydrographic regime connecting Vembanad lake at Elur. At Elur, it branches southwestward into a number of small distibutaries which suggest the existence of remnant deltaic regions named as Varapuzha palaeodelta. Details of the rivers discharging into the Vembanad lagoon is given in Table 2.1

Table 2.1 Details of the rivers discharging into the Vembanad Lagoon (Water Resources of Kerala, 1974; Soman, 2002)

River	Length (Km)	Catchment Area (Km^2)	Annual Run-off (1000 MC ft.)
Chalakudi	130	1704	42.00
Periyar	244	5398	434.00
Muvattupuzha	121	1554	93.68
Minachil	78	1272	96.27
Manimala	90	847	72.67
Pamba	176	2235	222.80
Achankovil	128	1484	76.00

Muvattupuzha is one among the perennial rivers of central Kerala with a length of about 120 km and a catchment area of about 1,554 km^2 (*cf.* Water Resource of Kerala, 1974) and debounches into Vembanad lagoon. The river originates from the Western Ghats and drains mainly through highly lateritised crystalline rocks and finally ends into the Vembanad lagoon near Vaikom. Two major tributaries namely Thodupuzha and Kaliyar join the Muvattupuzha river near Muvattupuzha town. After flowing as a single stream upto Vettukattumukku, the river branches into two distributaries namely Ittupuzha and Murinjapuzha. The river exhibits dendrite drainage pattern. The river discharge ranges from 50 m^3/sec (premonsoon) to 400 m^3/sec (monsoon). Peak discharge is recorded during June to October. Considerable changes have taken place in the flow characteristics of the Muvattupuzha river after the commissioning of the Idukki hydroelectric project in 1976, across the adjoining Periyar river. The tailrace water (19.83-78.5 m^3/sec) was directed into the Thodupuzha tributary from Moolamattom power station. This tailrace water (almost constant) plus surface run-off have not only altered the morphological characteristics of the river considerably but also the sediment dynamics and ecological habitat of the river basin as well.

Pamba is the third longest river (176 km) in Kerala having the fourth largest catchment area of 2235 km^2. It rises in the hill ranges of Pathanamthitta district and the adjoining Pirmed plateau, and is formed by the confluence of Pambiyar Ar, Kakki Ar, Arudai Ar, Kakkad Ar and Kal Ar. Pamba and Kakki are the major reservoirs in the basin. Flowing through the crystalline rocks (pyroxene granulites, charnockite gneiss, khondalites and associated calc-granulite bands), the basin displays dendritic to sub-dendritic and rarely rectangular and trellis drainage patterns.

The river Chalakudy has a length of 130 km and a drainage area of 1704 km^2. Five streams – Parambikulam, Kuriakutty, Sholayar, Karapara and Anakkayam form the Chalakudy river. Of these, Parambikulam and Sholayar rise in the Anaimalai hills at elevations above 1733 m and 1332 m, respectively. The river flows through thick forests and the channel has many waterfalls until it reaches the plains at Kanjirapally. The river debouches into the right arm of the Periyar at Puthenvelikkara. Poringalkuttu, Sholayar and Parambikulam are the major reservoirs in the basin. Charnockites and migmatitic gneisses are encountered along the river course.

Achankovil is also one of the main river of the Kerala with a length of 128 km and a catchment area of 1484 km^2. Achankovil river joins with the river Pamba before entering into the Vembanad lagoon. The Manimala river has a length of 90 km and has a catchment area of 847 km^2 and Minachil river has a length of 78 km and a catchment area of 1272 km^2.

2.5 Climate and Rainfall

Kerala state is diverse which causes the diversity in its climate too. Monsoon is the main feature of tropical climate experienced in Kerala. Subtropical type of climatic regime is seen in certain areas in the eastern part of the state due to the high variation in relief from west coast to the hilly regions of the Western Ghats in the east. Heat waves from the plain of Tamil Nadu enter in Kerala through Palghat Gap and Ariankavu pass. India Meteorological Department has grouped the seasons in Kerala into four seasons:

1. Hot weather period (the pre-monsoon season): March - May
2. Southwest monsoon: June - September

3. Retreating southwest monsoon (i.e., onset of northeast monsoon): October – November

4. Winter: December – February.

The atmospheric temperature reaches a maximum of 37^0C during the pre-monsoon period and from June onwards it gradually comes down due to heavy rainfall. Land and sea breezes influence the coastal area and here the seasonal and diurnal variations of temperature and almost of the same range (5 - 7^0 C). Kerala experiences two monsoons, namely the Southwest (June to September) and northeast (October – December) seasons. The Kerala state receives the highest annual rainfall among the other states of India, which is three times the average rainfall of India. The average annual rainfall varies from 100-400 cm with an average of about 300cm. The northeast monsoon is generally weak along the West Coast and the average annual rainfall is about 60 cm (Pisharody, 1992; Sampath and Vinayak, 1989).

2.6 Waves, Currents and Tides

West coast of India is experiences SW monsoon (rough weather season) between June and September. The wave activity becomes very strong during June and July and low in August and September. The sea remains relatively calm during October to May which is considered as fair weather season. During fair weather season, long swells often mixed with the local wind-seas prevail along the coast. Central Kerala coast is a micro-tidal zone, and falls under the major tectonic and morphologic class of the 'Amerotrailing' edge coast (Inman and Nordstrom, 1971). The dominant energy in the nearhsore is composed of gravity waves and mean currents or circulation. Far infra-gravity wave energy is about two orders of magnitude larger than that of gravity waves and evidence of edge waves in the infra-gravity band was demonstrated by Tatavarti *et al.*

(1996). Far infra-gravity waves may be a common feature in the nearshore oceans in the presence of longshore currents (Bowen and Holman, 1989).

The wind and current systems along the coast play an important role in the dynamics of mudbanks of the southwest coast of India. The important feature of the wind system in the Indian seas is a seasonal reversal of direction associated with two monsoons. Along the west coast of India, during the southwest monsoon the winds blow southwards from May to September and attain a northerly direction during the northeastern monsoon. Thus, the seasonally reversing wind pattern influences the southward littoral drift during the southwest monsoon, while a northward drift occurs during the northeastern monsoon. By the middle of May, the southwest monsoonal winds of oceanic origin are established. These winds continue to increase gradually until June when there is a 'burst' or sudden strengthening of the southwest winds. During July and August, the winds have their highest strength, and in September the wind force decreases ahead of the fall transition, which lasts through October and November (Sharma, 1978).

Estuarine hydrography plays an important role in the sedimentation and geochemical processes of this environment. The quantum, duration, transport and settlement of the particulate sediments depend directly on estuarine hydrography. The hydrography of the Vembanad estuary has been investigated by several researchers (Quasim and Reddy, 1967; Quasim *et al*., 1968; Sankaranarayanan *et al*., 1986; Anirudhan, 1988; Rasheed *et al*., 1995). The distribution of the temperature in the estuary is a function of the input of freshwater from rivers as well as the intrusion of salt water from Lakshadweep sea. Processes like exchange of heat with atmosphere and other localised phenomena are also likely to influence the hydrographic conditions of the system. The temperature of the water in the lagoon varies between 25-31°C.

Low salinity values ranging from 0 to 10 x 10^{-3} at the surface and 0 to 12 x 10^{-3} at the bottom were observed during monsoon. This was brought about by the combined effect of land drainage from the prevailing monsoonal rains causing high freshwater discharge from the river and intrusion of saltwater from the sea. As the season advances to post and premonsoon, higher salinity values ranging from 10 to 22 x 10^{-3} at the surface and 12 to 24 x 10^{-3} at the bottom were observed (Anirudhan, 1988). The estuarine waters considerably get diluted near the Muvattupuzha river confluence. The pH values of the surface and bottom water vary from 6.6 to 7.4 and slight increase is observed seasonally upto postmonsoon period.

****

Chapter 3

METHODOLOGY

Contents

3.1. Introduction:

The nearshore zone is a highly energetic environment affected by surface gravity waves and mean currents. The nearshore environment is also a transition zone characterised by on-offshore sediment transport between the surf zone and the offshore zone. The exchange of sediment between the inshore and the offshore environment plays a major role on coastal evolution and shoreline stability (Swift, 1976). A good understanding of sediment transport processes in the nearshore zone is important for scientists and coastal zone managers.

Vembanad lagoon is the largest backwater system on the Kerala coast extending 80 km in a NW–SE direction from Munambam in the north to Alleppey in the south. The width of the lagoon varies from 500 m to 4 km and the depth from < 1 m to 12 m. The Periyar river discharges into the lagoon in the north and the Muvattupuzha river in the central part. The rivers Meenachil, Manimala, Achanakovil and Pamba discharge into the southern part of the lagoon. Estuaries and deltas develop at river mouths during transgressive and regressive phases, respectively (Boyd et al., 1992). Rivers and its estuaries provide connectivity between terrestrial and marine environments.

The present investigation is scheduled into two phases namely i) sample collection and ii) laboratory methods

3.2. Collection of sediment samples:

3.2.1. Estuarine and nearshore environment:

Nine cores in the Vembanad estuary, eleven gravity cores and twenty two grab samples in the nearshore of Cochin were recovered. The length of the estuary cores vary from 42 to 156 cm and that of the nearshore cores vary from 58 to 190 cm. All the cores were sub sampled at an interval of 2 cm. The subsamples and grab samples were stored in air tight polythene bags. Selected subsamples of the nearshore cores were analyzed for the wet bulk density, moisture content, organic carbon, clay mineralogy, geochemistry, geotechnical and textural parameters. Selected subsamples of the estuary cores were analyzed for the wet bulk density, moisture content, organic carbon, clay mineralogy, textural and geotechnical parameters. Grab samples were analyzed for organic carbon, clay mineralogy, texture and geotechnical properties.

3.2.2. Onshore environment:

A borehole of 40 m depth was drilled at Paravur (Fig: 3.1). The area represents Holocene marine regression environment. The samples were stored in air tight polythene bags. Selected samples of the borehole were analyzed for estimating wet bulk density, moisture content, organic carbon, clay mineralogy, textural and geotechnical parameters, pollen analysis, and C-14 dating.

The locations of the cores and grab samples recovered from the estuary and nearshore are shown in the figure 3.2. and figure 3.3.

3.3. Geotechnical studies of sediment samples:

Sediment samples collected from the borehole, nearshore as well as estuarine area were subjected to geotechnical investigation. The geotechnical parameters analyzed include water content, wet bulk density, shear strength and Atterberg limits. The Atterberg limits comprise of liquid limit, plastic limit, plasticity Index and liquidity Index. The geotechnical parameters of sediments were analyzed according to the procedures as suggested by IS:2720 (Part II)-1964, (ASTM D 4318-2000), Singh and Punmia (1970) and Venkataramaiah (1993).

3.3.1. Water Content

To find out the water content of a sediment sample, oven drying method was used (IS:2720 (Part II)-1964). Sample weighing 10 - 20 grams was kept in a clean container in a thermostatically controlled oven for about 24 hours at 60^{o}C. A clean non-corrodible container was taken and weighed with its lid on an electronic balance. Sample was placed in the container and the lid was replaced. The container and the contents were weighed.

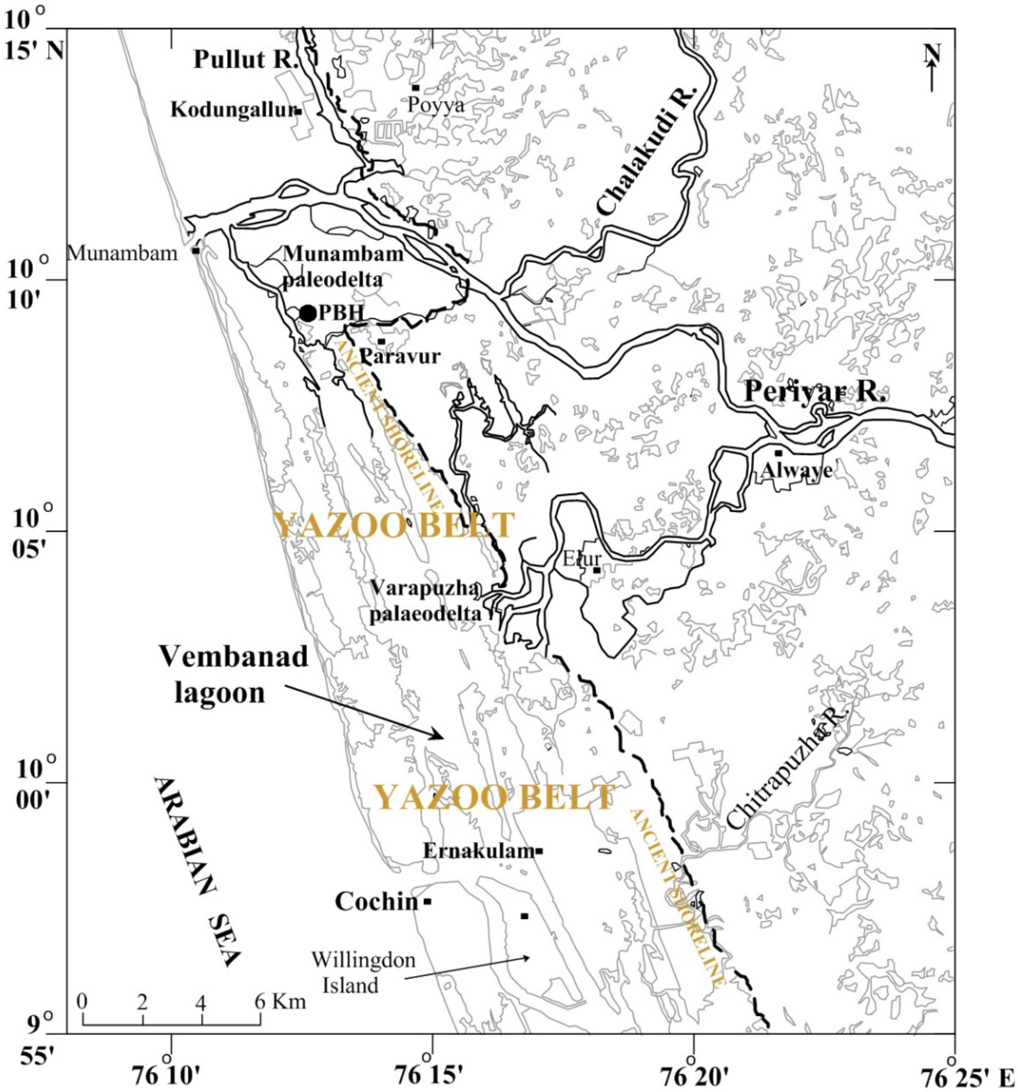


Figure 3.1: Geomorphic units in the Periyar river basin and adjacent areas interpreted from satellite imagery (IRS 1B LISS II FCC bands 2, 3 and 4; March 1995), Narayana et al., 2001). Map shows the location of borehole of 40 m depth shown as a solid circle.

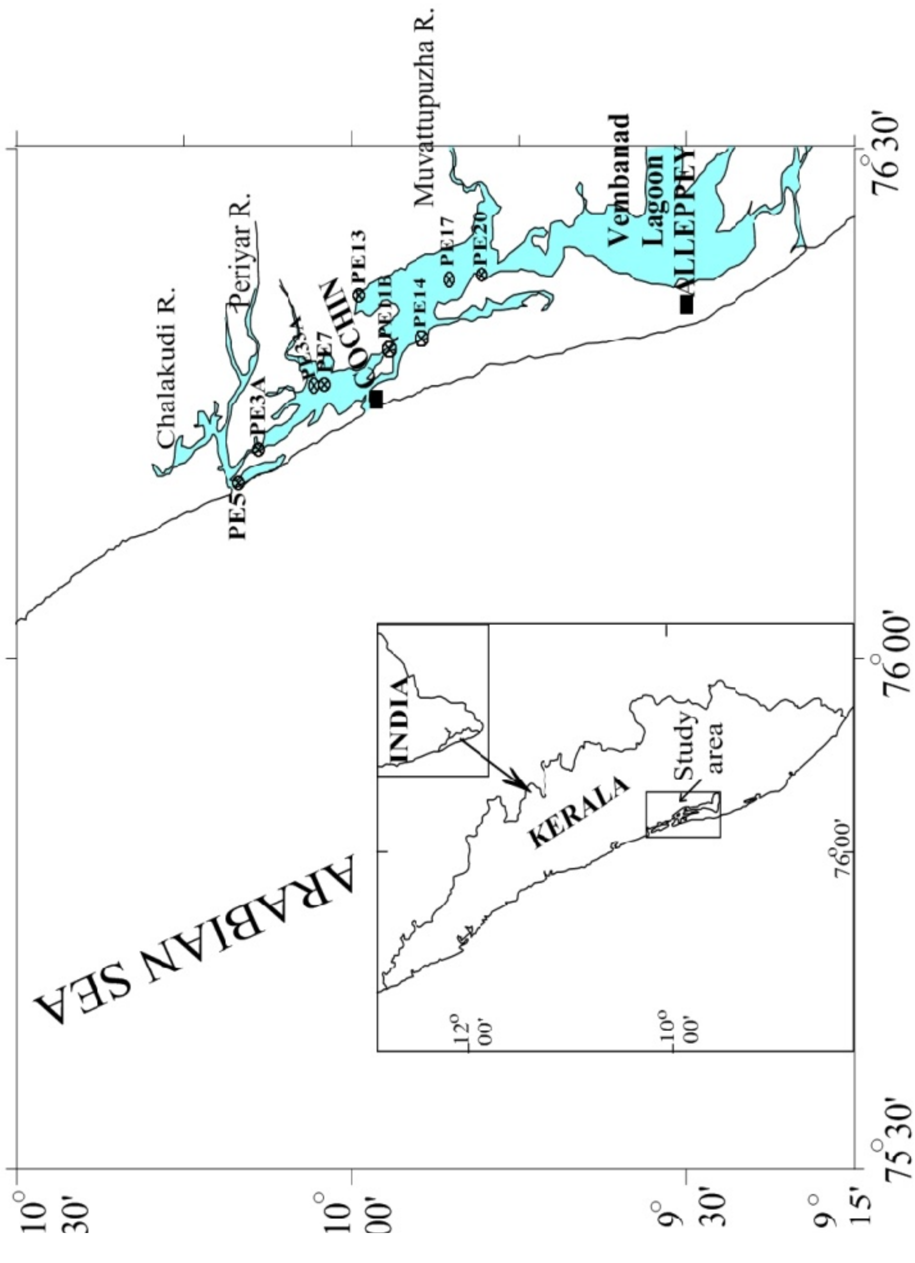
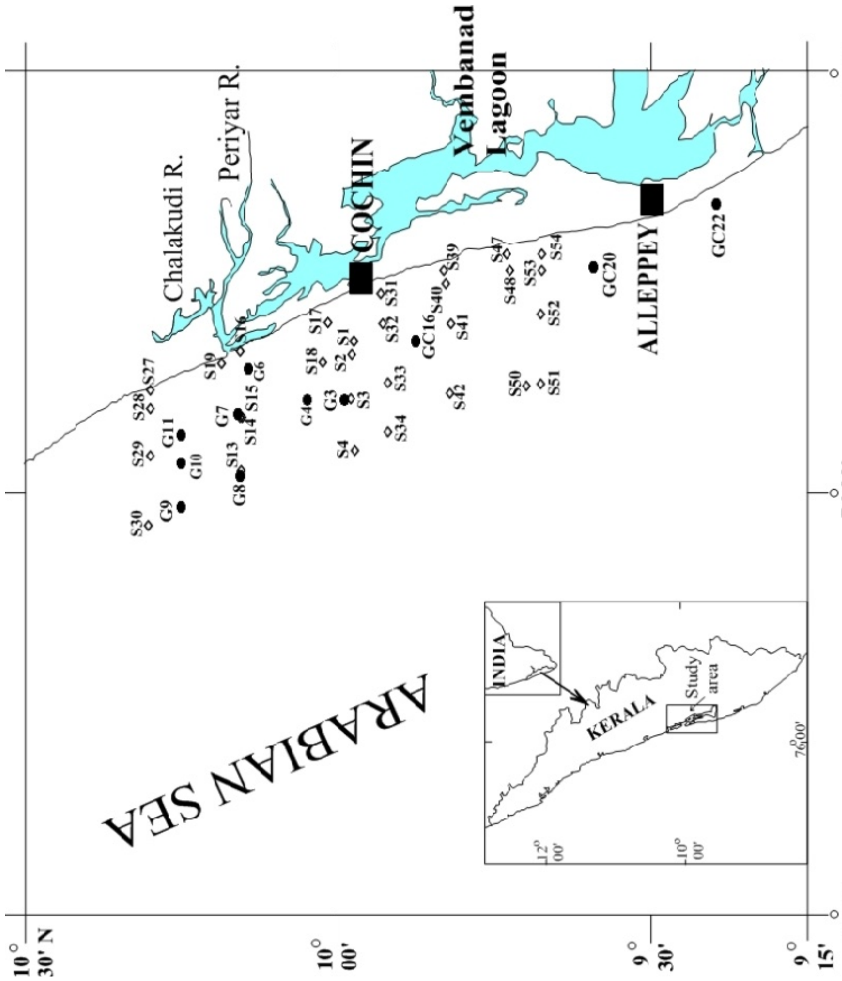


Figure 3.2: Study area an location of core samples in Vembanad lagoon (⊗)



Figures 3.3: Map showing the study area and location of surface (grab) and core samples. (◇)-surface/grab samples; (●) – core samples

Table 3.1: Details of the core samples collected from the Periyar estuary

Estuary core	Water depth (m)	Core length (cm)
PE-3A	1.3	52
PE-5	0.6	72.5
PE-7	1.3	82
PE-11B	1.6	42
PE-13	1	92
PE-14	1.5	108
PE-17	2	72
PE-20	1.5	70
PE-33A	1	156

Table 3.2: Details of the nearshore cores of Central Kerala

Nearshore core	Water depth (m)	Core length (cm)
G-3	20	80
G-4	20	160
G-6	10	80
G-7	20	110
G-8	30	170
G-9	30	58
G-10	20	178
G-11	10	155
GC-16	20	190
GC-20	10	80
GC-22	5	100

Table 3.3: Details of the grab samples collected from the nearshore areas of Central Kerala

Sample No	Water depth (m)
S-27	5
S-28	10
S-29	20
S-30	30
S-13	30
S-14	20
S-15	10
S-16	5
S-31	5
S-32	10
S-33	15
S-34	30
S-17	5
S-18	10
S-39	5
S-40	11
S-41	20
S-42	30
S-51	30
S-52	20
S-53	10
S-54	5

With the lid removed, the container was then placed in the oven for drying. After drying, the container was removed from the oven and allowed to cool in a dessicator. The lid was then replaced, and the container and the dry sediment were weighed. The water content was calculated from the following expression:

$$W = \frac{W_2 - W_3 * 100}{W_3 - W_1}$$

Where, W_1 = Weight of container with lid

W_2 = Weight of container with lid and wet sample

W_3 = Weight of container with lid and dry sample

3.3.2. Wet Bulk Density

In a pre-weighed container of known volume, sample was taken and found out the total weight (W1) of sample and the container. To remove the water content, the container along with the sample was kept in an oven and dried the sample completely. The dried sample and container was weighed to find out the dry weight (W2) of sample and the container. Wet bulk density was found out from the following expression:

$$\text{Wet bulk density} = \frac{W1 - W2}{V}$$

Where, W1 = Weight of wet sample + container

W2 = Weight of dry sample + container

V = Volume of sample

3.3.3. Atterberg limits:

The Atterberg limits are a basic measure of the nature of a fine grained soil, and can be used to distinguish between silt and clay and between different types of silts and clays (Kwon et al., 2011). Evaluation of the Atterberg limits is a soil mechanical test allowing a first insight into the chemical reactivity of clays. Basically, the liquid limit and the plasticity index are highly and mainly influenced by the ability of clay minerals to interact with liquids (Schmitz et al., 2004).

Generally the fabric, physico-chemistry of clay-water system and organic carbon content of sediments influence the Atterberg limits (Rosenquist, 1962; Sodorblom, 1966; Rashid and Brown, 1975; Bennett et al., 1981). Atterberg limits explain the consistency of sediments, which means the relative ease with which sediment can be deformed. This term is mostly used for fine-grained sediments for which the consistency is related to a large extent to the water content. Atterberg limits include liquid limit, plastic limit and plasticity index. Atterberg limits were analysed using the standard procedures (ASTM, D 4318-2000).

3.3.3.1. Liquid limit:

It is the water content corresponding to the arbitrary limit between liquid and plastic state of consistency of sediments. It is defined as the minimum water content at which the sediment is still in the liquid state, but has a small shearing strength against flowing was measured by using standard available procedure (IS 2720 Part-V)-1970).

The liquid limit was determined in the laboratory with the help of the standard liquid limit apparatus designed by Casagrande. The apparatus consists of a hard rubber base of B.S. hardness 21-25, over which a brass cup drops through a desired height. The height of fall of the cup can be adjusted with the help of adjusting screws. Before starting the test, the height of fall of the cup is adjusted to 1 cm. The Casagrande tool cuts a groove of size 2mm wide at the bottom, 11mm wide at the top and 8mm high.

About 120g of the sample was mixed thoroughly with distilled water on a glass plate to form a uniform paste. A portion of the paste was placed in the cup over the spot where the cup rests on the base, squeezed down and spread into position and the groove was cut in the soil pat as shown in the figure 3.4. The handle was rotated about 2 revolutions per second, and the number of blows was counted until the two parts of the sediment sample come into contact at the bottom of the groove along a distance of 12mm. After recording the number of blows, approximately 10 grams of sediment from the closed groove was taken for determination of water content. Since it was difficult to adjust the water content precisely equal to the liquid limit when the groove should close in 25 blows, the liquid limit was determined by plotting a graph between the number of blows at abscissa on algorithmic scale and the corresponding water content as ordinate. It shows that such a graph, known as flow curve, is a straight line having the following equation:

$$W_1 - W_2 = \log_{10} \frac{n_2}{n_1}$$

where, W_1 = water content corresponding to blows n_1

W_2 = water content corresponding to blows n_2

For plotting the flow curve, at least four to five sets of readings in the range of 10 to 40 blows were taken. The water content corresponding to 25 blows is taken as the liquid limit.

(a) (b)

Figure 3.4 – (a) Groove separating two halves of the soil pat, (b) blow counts stopped after the groove closes by about 12mm size.

3.3.3.2. Plastic limit (PL):

It is the water content corresponding to an arbitrary limit between the plastic and the semi-solid states of consistency of sediment. It is defined as the minimum water content at which a sediment mass will just begin to crumble when rolled into a thread approximately 3mm diameter or it is the minimum amount of water necessary to make clay plastic.

To determine the plastic limit, the sample was mixed thoroughly with distilled water until it became plastic enough and moulded easily with fingers. The plastic sediment mass was left for enough time and allowed water to permeate through the sediment mass. A ball was formed with about 10gms of this plastic sediment mass and rolled between the fingers and a glass plate with just sufficient pressure to roll the mass into a thread of uniform diameter throughout its length. When a diameter of 3mm was reached, the sediment was remoulded again into a ball. This process of rolling and remoulding was repeated until the thread starts just crumbling at a diameter of 3mm. The crumbled threads were kept for water content determination. The test was repeated twice more with fresh samples. The plastic limit 'wP' is then taken as the average of three water contents (Singh and Punmia, 1970).

3.3.3.3. Plasticity Index (I_P):

The range of consistency within which a sediment mass exhibits plastic properties is called plastic range and is indicated by plasticity index. The plasticity index is defined as the numerical difference between the liquid limit (wL) and the plastic limit (wP) of sediment mass.

The plasticity index is calculated from the relation:

$$I_P = wL - wP$$

PI and their meanings

Non plastic $I_P = 0$, Slightly plastic $I_P = 1 – 5$, Low plasticity $I_P = 5 – 10$, Medium plasticity $I_P = 10 – 20$, High plasticity $I_P = 20 – 40$, Very high plasticity $I_P > 40$

Low plasticity $w_L = < 35\%$, Intermediate plasticity $w_L = 35 - 50\%$, High plasticity $w_L = 50 - 70\%$, Very high plasticity $w_L = 70 - 90\%$, Extremely high plasticity $w_L = > 90\%$ (w_L Liquid limit) (after Casagrande, 1948).

3.3.3.4. Liquidity index:

The liquidity index (LI) is used for scaling the natural water content of a soil sample to the limits and can be calculated as a ratio of difference between natural water content, plastic limit, and plastic index

$$(LI = (wc - wP)/(wL – wP)$$

where wc is the natural water content (Kwon et al., 2011).

3.3.4. Vane Shear:

The most versatile test to measure compressibility and shear strength parameters of soils is the triaxial test, in which a cylinder – shaped specimen, sealed by a rubber membrane, is submitted to an axisymmetric water pressure and then to an increasing axial loading. If the test is carried out under undrained conditions (no water is allowed to get out of the specimen during the test), the undrained shear strength (su) of the soil can be measured. The triaxial test depends on the preparation of good-quality undisturbed specimens. When the soil is too soft for an adequate preparation of undisturbed specimens (which is usually the case when very soft marine sediments are to be tested), the shear vane test is an alternative for estimating the undrained shear strength. It is determined by applying a rotation to a vane introduced into the soil sample and measuring the maximum torque that the soil can bear before failure (Dias and Alves (2009).

Shear strength is the resistance to shearing stress and a consequent tendency for shear deformation. The shear strength of the sediment sample was determined by vane shear test as this method widely used for soft clays. In the present study shear strength was determined by using Laboratory Vane Shear Apparatus (IS: 2720 (Part XXX), 1980).

The shear strengths of the clay is expressed as:

$$S = T/\pi D2 (H/2 + D/6)$$

Where, T = torque

D = diameter of the vane,

H = height of the vane

3.4. Organic Carbon

Organic carbon in marine sediments is primarily derived from settling biogenic debris from the water column and their abundances serve as indicators of primary productivity (Muller and Suess, 1979), whereas in estuaries and nearshore areas, the organic carbon is contributed by terrestrial processes. The organic carbon concentrations in sediments, especially in the continental margin region, depend upon the depositional environment, sedimentation rate, bottom water redox condition etc. (Canfield, 1994).

For estimation of organic carbon content in sediments, samples were oven dried at 50°C and finely ground in agate mortar for the determination of organic carbon content. The organic carbon content (Corg) of sediment samples were determined by using a modified Elwakeel and Riley (1957) method by Gaudette et al., (1974) which is based on the exothermic heating and oxidation of organic matter. The sample was treated with potassium dichromate and concentrated sulphuric acid, followed by back titration with ferrous ammonium sulphate using diphenyl amine as an indicator. The principle behind this method is based on the oxidation of organic carbon with chromic acid and titrimetric determination of the oxidant consumed.

Powdered sediment sample of 0.5g is taken and treated with 10ml of 1N $K_2Cr_2O_7$ solution and 20 ml of conc: H_2SO_4. It is kept for half an hour and then diluted by adding 170ml of distilled water, followed by adding 10ml of orthophosphoric acid and 0.2g of NaF. About 2 or 3 drops of diphenyl amine indicator is added. The change from turbid green colour to brilliant green indicates the end point. The volume of ferrous ammonium sulphate is noted as T. A blank solution is prepared by taking all the reagents in the same amount as above and titrated against ferrous ammonium sulphate. Volume of ferrous ammonium sulphate used is noted as S. The percentage of organic carbon is calculated using the equation 10(1-T/S). The organic carbon in the bulk sediment was determined in all the surficial samples, nearshore and estuarine core samples and borehole samples employing the above stated method.

3.5. Textural analysis

Sediment grain-size composition has been widely used to characterize sediment sources and transport patterns and to reconstruct depositional environments (Folk and Ward, 1957; Sun et al., 2002). Moreover, the association of organic carbon with mineral surfaces has been established throughout the global oceans (e.g. Hedges and Keil, 1995; Keil et al., 1998; Wakeham et al., 2009), such that grain-size composition is central to understanding the fate of sedimentary organic matter. For example, fine-grained sediments have larger surface area and thus higher organic matter loading capacity than coarse sediments (Hedges and Keil, 1995)

In general the term texture represents the size, shape, roundness, grain surface features and fabric of grains. Because of differential erosion, transportation and deposition sediments laid down in different depositional environments may possess distinct particle size distributions. By determining the particle size distributions it is possible to talk about the environment of deposition and so to utilize this technique as a tool for environmental reconstruction (Lario et al., 2002). Textural parameters comprising sand, silt and clay ratios of the sediment core samples were carried out using standard methods (Carver, 1971; Folk, 1980).

Textural analysis of core samples was carried out at 5cm interval following the standard sieve and pipette techniques. About 15 to 20 g of dried sediment sample was weighed accurately and transferred into a thoroughly cleaned l000ml beaker. The samples were made salt free by repeated washings using distilled water and subsequently treated with H_2O_2 and glacial acetic acid in order to remove the organic matter and carbonates respectively. The samples were wet sieved through a 63µm sieve to separate sand and mud fraction. The sand fraction (>63µm) retained in the sieve was dried and weighed, while the mud fraction was collected in a 1000 ml measuring glass cylinder and subjected to pipette analysis to estimate the silt and clay contents following the method of Folk (1980). Percentage distribution of sand, silt and clay fraction in each sample was determined. Size classification adopted is sand >63µm, silt 63 - 4µm and clay < 4µm. Based on the relative abundance of sand, silt and clay in each sub sample in each size class, sediment texture was determined according to the classification of Folk (1980). The sand-silt-clay ratios were plotted on trilinear diagram for textural classification / nomenclature based on Folk's (1980) textural classification.

3.6. Clay mineral analysis

The clay minerals in the samples were identified and semi-quantified employing the following steps.

3.6.1. Sample preparation:

Dried sediment sample weighing 15-20 g was taken and dissolved salts were removed by washing thoroughly with distilled water. This process has been repeated till the sample was deflocculated. The presence of organic matter in the sample can produce broad X-ray diffraction peaks, increase the background, and inhibit dispersal of other minerals if present in significant amount. Therefore, the organic matter from the sediment samples was expelled by treating with 20-25 ml of 30% H_2O_2. The sample was again washed with distilled water. The sample was also made free of carbonates by treating with 10 ml glacial acetic acid. Excess acid was removed by washing with distilled water. After the sample is disaggregated and deflocculated, >63μm size fraction sediments were separated by wet sieving and the supernatant was collected in a 1000 ml cylinder. The 2μm size fraction of sediment was separated based on the settling velocity (Stoke's law) principle. The resulted suspension was transferred to a settling column and allowed to settle for 3 hrs 27 minutes i.e., until the required size fraction was obtained. The clay water suspension was used for making oriented clay slides of almost equal size and thickness by pipetting equal volume (1ml) on to glass slides. The glass slides were allowed for air-drying at room temperature, while drying, care was taken to avoid contamination by dust and other means (Biscaye, 1964; Gibbs, 1965).

3.6.2. X-Ray diffractogram

X-ray diffraction methods are commonly used for identification of mineral components in soils, sediments and rocks, particularly for clay minerals in sediments and soils. X- ray diffraction studies were carried out on the air dried samples (Gibbs, 1965). The slides were run on X-ray Diffractometer (XRD) from 2 to 30^o 2θ at 1^o 2θ / minute using Ni filtered CuK∞ radiation. The XRD facility at National Institute of Oceanography, Goa (Philips-1840 Model), was availed. The samples were then glycolated by exposing the slides to ethylene glycol vapours at 100°C for one hour and rescanned the slides under same instrumental settings for the confirmation of montmorillonite. The peaks for different clay mineral groups like kaolinite, chlorite, illite and smectite were identified. In order to differentiate the kaolinite and chlorite peaks, the samples were also scanned from 24 to 26° at 0.5° 2θ/minute (Biscaye, 1964).

3.6.3. Semi-quantification of clay minerals

The relative contents of clay minerals were determined using peak areas of smectite (15-17 Å), illite (10 Å), chlorite (7 Å) and kaolinite (7 Å). The peak areas of the spectra of these clay minerals were calculated by using the glycolated X-ray diffractograms and relative weight percentages were calculated following the semiquantitative method of Biscaye (1965). The relative clay mineral contents of smectite, illite, kaolinite and chlorite were determined using ratios of integrated peak areas of their basal reflections, weighted by empirically estimated factors. Accordingly, 17º Å peak area of the smectite is multiplied by 1, the 10 Å peak area of illite by 4, and both the kaolinite and chlorite proportions positioned at 7 Å peak were multiplied by 2. As the 001 plane of kaolinite and 002 plane of chlorite partly overlap each other at 7 Å, the relative proportions of both minerals were first deduced from the areas of 002 kaolinite and 004 chlorite reflections at 3.54 Å and 3.58 Å respectively. The clay minerals quantified in this study are smectite, kaolinite, illite and chlorite.

3.7. Radiocarbon (^{14}C) dating

During the course of geological time, equilibrium has been achieved between the rate of ^{14}C production in the upper atmosphere and the rate of decay of ^{14}C in the global carbon reservoir. This means that the weight of ^{14}C estimated to be produced each year in the upper atmosphere is approximately equal to the weight of ^{14}C lost throughout the world by the radioactive decay of ^{14}C to nitrogen with the release of β-particles (6C14 → 7N14 + β+ neutrino). The total weight of global ^{14}C thus remains constant. Further, plants and animals assimilate a certain amount of ^{14}C into their tissues through photosynthesis and respiration. The ^{14}C content of these tissues is in equilibrium with that of the atmosphere because there is a constant exchange of new ^{14}C once the old cells die. As soon as the organism dies, this exchange and replacement of ^{14}C radioactive clock will be activated (Bradley, 1999).

In most of radiocarbon laboratories, conventional methods -either proportional gas counters or liquid scintillation techniques – are employed. In gas counters, carbon is converted into a gas (methane, CO_2 or acetylene), when it is put into a "proportional counter" capable of detecting β particles. In liquid scintillation counters, the carbon is converted into benzene or some other organic liquid and detects scintillations produced by the interaction of β particle and a phosphor added to the organic liquid. In the present study, radiocarbon dating of sediment core samples was carried out by Gas Proportional Counting method.

In order to understand the geochronology and paleo-environment of the area, the borehole samples (at Paravur) were dated for radio carbon ages. The sediment samples of Paravur borehole were acidified using mild orthophosphoric acid to extract carbon-dioxide under vacuum conditions (Yadava and Ramesh, 1999). The extracted CO_2 is purified by trapping moisture and CO_2 in different traps cooled by various coolants. The purified is reacted with tritium free hydrogen gas in the presence of a catalyst in a stainless steel reaction vessel maintained at 450-500^{o}C. The reaction between CO_2 and hydrogen takes for about 12-15 hours to form methane gas (CH_4). The synthesised methane gas is taken in to a glass vacuum system and purified. The synthesised methane gas is filled in to a gas proportional counter kept in a lead and iron shield. The gas proportional counter is given DC power to develop a potential gradient inside the counter. The charge developed inside the counter gets accelerated by the potential gradient. Finally, the negative charge is collected by the collector electrode. The collected charge gets converted to voltage pulses by the counting electronics. The beta radiation emitted by the carbon-14 atoms is measured in anti-coincidence channel. The net beta count rate is obtained by subtracting the total count rate from the background count rate. The net count rate of the given sample is converted to percent modern carbon upon comparison with the modern standard net count rate. The percent modern carbon value is used to calculate age of the sample based on 5730 year as the half life of carbon-14.

The ^{14}C dating was carried out on sediment layers representing different sections of the Paravur borehole at BSIP, Lucknow.

3.8. Palynological studies of borehole samples

"As cogently noted by Alfred Russel Wallace during his equatorial travels, mangrove forests are crucial occupiers of the boundary between land and sea, being key ecosystems along many tropical and subtropical coastlines" (Alongi, 2009, p. 1). "Mangrove" is an ecological term that refers to a taxonomically diverse assemblage of trees and shrubs that forms dominant plant communities in tidal, saline wetlands along sheltered tropical and subtropical coasts (Blasco et al., 1996). Mangroves prosper mostly in tropical regions as a result of their adaptation strategies, their ecological dynamics being closely linked to changes in sea level. Pollen analysis of mangroves is important for both palaeoecological reconstructions of coastal vegetation and determinations of palaeoenvironment in tropical and subtropical regions (Muller, 1964; Muller and Caratini, 1977; Torricelli et al., 2006; Ellison, 2008; Berkeley et al., 2009).

Palynological analysis of sediments at different sections of the Paravur borehole was done at BSIP, Lucknow in accordance with the standard procedure (as described by Faegri and Iversen, 1992). For palynological study, about 50g of sample was processed and cleaned to free the fossils from extraneous matter. Depending on their lithological composition, the samples were subjected to various chemical treatments (HCl, HF, HNO3, and KOH) for removing undesired organic matter and to clean and concentrate any microfossils. After the completion of chemical reactions, the samples were screen-washed with distilled water using 500 mesh ASTM sieves. The macerated residues were smeared on glass cover slips, mixing with polyvinyl alcohol solution. The dried cover slips were mounted on glass slides using Canada balsam. The slides were oven dried and then examined and photographed under the microscope at high power.

3.9. Trace elements

Prevalent in the clay fraction, metals are predominantly delivered to the coastal waters in association with riverborne particles that act as transporting agents. Their behaviour and fate strongly depend on sedimentary dynamics (Zhang, 1999; Che et al., 2003), but also biological and physico-chemical interactions occurring in the mixing zone between fresh and salt waters can play an important role (Comans and Van Dijk, 1988; Regnier and Wollast, 1993; Eggleton and Thomas, 2004).

The varieties of processes result in trace element enrichments that mirror the specific conditions prevailing at the time of deposition and early diagenetic stages. Consequently trace element abundances in sediments allow us to reconstruct paleodepositional conditions (Werne et al., 2003; Lyons et al., 2003; Rimmer, 2004; Rimmer et al., 2004; Algeo and Maynard, 2004; Algeo, 2004; Nameroff et al., 2004; Tribovillard et al., 2004a). While using trace element concentrations to reconstruct paleoenvironmental conditions, one must assess whether they are relatively enriched or depleted. In general, the degree of enrichment or depletion of a trace element in a sample is evaluated relative to its concentration in a reference that is commonly the average crustal rocks or average shale (Wedepohl, 1971, 1991; Mclennan, 2001). In general, trace elements in fine grained sediments, which are relatively rich in organic matter, are used for paleoenvironmental reconstruction. Rare earth elements are important because their geochemical properties enable them to be powerful tracers of chemical processes. Trace elements are analysed using the following procedure:

3.9.1. Dissolution of sediment samples:

To a 50 mg sample, 10 ml acid mixture of 6 parts HF, 3 parts of HNO3 and 1 part of HCl were added along with 0.5 ml of 10 mg/ml Rh in microwave vessel to act as internal standard. The vessels were then closed and mounted in teflon beaker and heated. This procedure is repeated once again to ensure total dissolution of the total samples. After completion of the heating, the beakers were cooled to room temperature and carefully kept in a fume-hood and the then 1ml of $HClO_4$ is added to each beaker. The solution was evaporated to incipient dryness. The residue was dissolved in 20 ml of 1:1 HNO_3 and brought to a final volume of 250 ml. Clear solutions were obtained in all cases. A procedural blank solution was also prepared (Balaram and Rao, 2003).

Electronic grade HF, analytical reagent (AR) grade $HClO_4$ and distilled HNO_3 and HCl were used in the preparation of samples and standards. Distilled / deionised water was used for all analytical purposes.

3.9.2. Inductively Coupled Plasma-Mass Spectrometer (ICP-MS)

Trace element analyses were carried out on an Inductively Coupled Plasma – Mass Spectrometer (ICP-MS-Perkin Elmer SCIEX, Model ELAN R DRCII) at National Geophysical Research Institute, Hyderabad. The important optimization criteria include adjustment of nebular gas flow, setting of detector and lens voltages, radiofrequency forward power settings and performance of calibration.

The principles of ICP-MS are based on the assumption that, ions may be generated in a suitable ionising source such as an ICP for all elements in the sediment. Ions are physically extracted from the plasma into mass spectrometer and measured using an ion detector. Sample introduction for plasma spectrometry is generally accomplished using solution nebulisation. Sample solutions are aspirated by a nebuliser which shatters the liquid into fine droplets using an Ar stream of ~1L/minute. These droplets are directed into a spray chamber which removes the unsuitable larger material, and allows only the finest spray to pass into the plasma (Balaram, 1997). The ICP-MS offers very low detection levels. The analytical accuracy was checked by analysing international standards (Marine mud / MAG-1).

A total of 32 samples (14 from G11 at 10 m water depth; 9 from G10 at 20 m water depth; 9 from G8 at 30 m water depth) representing different sedimentary column of each core were analysed.

3.9.3. Geoaccumulation Index (Igeo)

The geo-accumulation index is a quantitative measure of the extent of contamination in aquatic sediments (Förstner et al.,1990). To gauge the degree of anthropogenic influence on heavy metals concentration in the sediments geoaccumulation index (Igeo) is used. Generally, the Igeo consists of 7 grades or classes (Table 3.4). The Igeo was calculated using Muller(1969) and Abrahim & Parker (2008) method as follows:

Igeo = log 2([sediment]/ 1.5* [reference sample])

The reference sample is either directly measured in pre-civilization sediments of the area or taken from the literature (average shale value described by Turekian and Wedepohl (1961). The factor 1.5 is introduced to minimize the effect of possible variations in the background values which might be attributed to lithologic variations in the sediments.

Table 3.4. The degree of metal pollution in terms of seven enrichment classes

Igeo Value	Igeo Class	Designation of sediment quality
>5	6	extremely contaminated
4--5	5	strongly to extremely contaminated
3--4	4	strongly contaminated
2--3	3	moderately to strongly contaminated
1--2	2	moderately contaminated
0-1	1	uncontaminated to moderately contaminated
0<	0	Uncontaminated

3.9.4. Enrichment Factor

A common approach to estimate how much the sediment is impacted (naturally and anthropogenically) with heavy metal is to calculate the Enrichment Factor (EF) for metal concentrations above un-contaminated background levels (Huu et al., 2010). According to this technique metal concentrations were normalized to the textural characteristic of sediments. It is standardization of a tested element against a reference. A reference element is the one characterized by low occurrence variability. The selected reference sample is usually an average crust or a local background sample (Chatterjee et al., 2007; Liu et al., 2005; Blaser et al., 2000). The low occurrence element is often taken to be Al (Chatterjee et al., 2007; Sutherland, 2000), Li, Sc, Zr (Blaser et al., 2000) or Ti, and sometimes Fe (Zhang et al., 2007) or Mn (Liu et al., 2005) has been used. In this study Sc was used as the reference element.

The EF is calculated using the following equation (after Feng et al., 2004):

$$EF = \frac{C_a \text{ x } Sc_b}{C_b \text{ x } Sc_a}$$

where C_a and C_b are the examined metal content in the sample and the background reference respectively; Sc_a and Sc_b are the Sc content in the sample and the background reference respectively. Five contamination categories are recognized by Sutherland (2000) on the basis of the enrichment factor as follows: EF < 2 is deficiency to minimal enrichment, EF 2 – 5 is moderate enrichment, EF 5 – 20 is significant enrichment, EF 20 – 40 is very high enrichment, EF > 40 is extremely high enrichment, as the EF values increase, the contributions of the anthropogenic origins also increase (Sutherland, 2000).

3.9.5. Pollution load index (PLI)

Pollution load index (PLI): Pollution load index was evaluated as indicated by Tomilson *et al.* (1980). Pollution load index= (CF1 * CF2 *.....* CFn) 1/n where, n is the number of metals (five in the present study) and CF is the contamination factor. The contamination can be calculated from; Contamination factor (CF) = metal concentration in sediments/Background values of the metal. The PLI value > 1 is polluted whereas PLI value < 1 indicates no pollution (Chakravarty and Patgiri, 2009; Seshan *et al.,* 2010).

****

Chapter 4

SEDIMENT TEXTURE AND GEOTECHNICAL CHARACTERISTICS

Contents

4.1 Introduction

River borne sediments dispersed into the coastal ocean undergo several cycles of transport, deposition, and resuspension before they become part of the long term geological record. Factors such as particle size, rate of sediment supply, oceanographic regime (waves, tides and currents) and the geometry of the shelf influence sediment dispersal and transport pattern. Asian rivers have high sediment yields as a result of high relief and rainfall in their drainage basins (Rao et al., 2005). The sediment influx from the Asian rivers is transported along the shelf by the monsoon currents or carried to the deep-sea floor through submarine canyons. In addition, more than 80% of the sediment and fresh water influx takes place during a time span of 4 months of the southwest monsoon period (May–August). Because of this combination of high tidal energy, monsoon related circulation and episodic sediment and freshwater influx, the sedimentary processes on the Asian shelves are different from those operating in the temperate and polar regions. Understanding the sedimentary processes in these areas is important as 30% of global river discharged sediment to the ocean is from the Asian rivers (Milliman and Meade, 1983). Study of these processes is also very important from the climate change point of view as 85% of the total global organic carbon burial rates occur in deltaic and shelf sediments (Berner, 1982).

The hydrodynamic intensity in shelf seas in monsoon-affected regions is strong in the winter season and weak in the summer season due to seasonal changes in wind strength. The water depth in shelf seas is generally no more than 200 m, and ranges from about 20 m to 100 m in mid-continental shelf and inner-continental shelf areas. The surface sediment deposited in summer on the shallow shelf is frequently resuspended and transported in the winter season by storm waves, resulting in significant seasonal variations in shelf sediment transport (Butman et al., 1979; Drake and Cacchione, 1985; Sternberg, 1986; Schoellhamer, 1996; Dyer and Moffat, 1998; Fettweis et al., 1998; Castaing et al., 1999; Muslim and Jones, 2003). Existing models of estuaries often describe the bottom sediment distribution in terms of tidal dynamics, wave energy, and river flow that control sediment transport in the system (Nichols and Biggs, 1985; Dalrymple et al., 1992; Bird, 2000). The result, e.g. for coastal plain estuaries, is a large scale pattern consisting of sandy sediments near the mouth of the estuary, muddy sediments in the central section and coarser sediments in the fluvial-dominated upstream section of the estuary (Dalrymple et al., 1992). Besides the large-scale pattern, the small-scale, local sediment distribution is influenced by several additional factors including riverbed morphology, bedrock type, tributary input, and human modifications (Nichols and Biggs, 1985).

Texture refers the size, shape, and those features of the grains which on systematic study give the idea of its depositional environment, processes acting, and the energy level of processes. Following Lario et al (2002), particle size distribution can be utilized as the tool for environment of deposition which can further provide a significant relationship between sediment properties and geologic processes or settings. Similarly, geotechnical properties also help in understanding the sedimentary processes such as erosion, transport, deposition, consolidation, etc. many of the alteration that occur after the sediment is deposited, or in turn, controlled by strength properties of the sediment (Silva, 1974). Hence the physical and geotechnical properties like water content, wet bulk density, organic matter, Atterberg limits, vane shear strength, etc, provides a comprehensive information to understand the transport mechanism and energies, and its depositional environments of the sediments.

In this chapter, textural and geotechnical parameters of nine core samples recovered from Vembanad lagoon, and eleven core samples and 22 grab samples in the nearshore sea of Cochin are discussed and the study attempts to understand the modern depositional sedimentary environments.

4.2 Sediment texture

In a system of related environments, sediment texture can be used to identify both the probable source and the probable deposit and, by inference, the net sediment transport paths among sedimentary deposits. Such an analysis provides a rapid understanding of the sedimentary processes, identifies patterns of erosion and accretion, and may suggest transport processes (McLaren, 1981). On the inner continental shelf and coastal environments, grain-size trends make it possible to characterize ancient and modern environments, and to identify net sediment transport patterns, using grain-size distributions. Grain size distributions follow trends that identify the direction of transport and the sedimentary processes of winnowing, selective deposition, and total deposition.

Light grains have a probability of being eroded and transported than heavy grains (McLaren and Bowles, 1985).

Most tide-dominated estuarine and deltaic deposits accumulate in the fluvial-to-marine transition zone, which is one of the most complicated areas on earth, because of the large number of terrestrial and marine processes that interact there. An understanding of how the facies change through this transition is necessary if we are to make correct paleo-environmental and sequence-stratigraphic interpretations of sedimentary successions. The most important process variations in this zone are: a seaward decrease in the intensity of river flow and a seaward increase in the intensity of tidal currents. Together these trends cause a dominance of river currents and a net seaward transport of sediment in the inner part of the transition zone, and a dominance of tidal currents in the seaward part of the transition, with the tendency for the development of a net landward transport of sediment. These transport patterns in turn develop a bedload convergence within the middle portion of all estuaries and in the distributary-mouth-bar area of deltas. The transport pathways also generate grain-size trends in the sand fraction: a seaward decrease in sand size through the entire fluvial–marine transition in deltas, and through the river-dominated, inner part of estuaries, but a landward decrease in sand size in the outer part of estuaries. A turbidity maximum (i.e., a zone of significantly elevated suspended-sediment concentrations) is developed within estuaries and the delta-plain region of deltas as a result of flocculation and density-driven water-circulation patterns. This leads to an area within the estuary or delta plain where the abundance and thickness of the mud drapes are greatest, including the potential for the development of fluid-mud deposits (i.e., structureless mud layers more than 0.5–1 cm thick that were deposited in a single slack-water period) (Dalrymple and Choi, 2007).

Jayalakshmi et al. (2001) reports about ~ 70 m thick Quaternary sediment, along the Kerala coast where oldest part is unfossiliferrous consisting of gray clay with admixture of lithoclasts of laterite derived from adjacent terrestrial regions. Prakash and Prithviraj (1988) report predominance of clayey silts and silty clays in the sediment distribution of the inner shelf zone of the central Kerala coast. He also reports that the clayey silts extend over the inner portion of the shelf adjoining the coastline and estuaries whereas silty clays dominate the central and the seaward portion of the inner shelf. Knowledge of various characteristics of sediment inputs permits the assessment of its state of evolution, its maturity and also a better understanding of its history (Priju, 2004). Veerayya and Murty (1974) have revealed the coarser sand fraction in central and southern parts of Vembanad lagoon and finer fraction in the remaining areas. The textural studies of surface sediments of central Vembanad estuary by Seralathan et al. (1993), and Seralathan and Padmalal (1994) report the muddy sand, sandy mud, sand, mud, clayey sand, and silty sand facies. Sunil and Antony (1994), and Badarudeen et al. (1996) have reported an admixture of sand and silt (silty sand and sandy silt) in the sediments of the mangrove areas in the central and southern parts the Vembanad Lake.

4.2.1. Sediment texture of lagoonal environment

Textural data of core samples of Vembanad lagoon are presented in the ternary diagram of Folk (figure 4.1 & 4.2). Core PE-3A was recovered from the southern arm of the Periyar River whereas core PE-5 was taken from the mouth region of the Periyar River at Munambam. The percentage of sand, silt and clay varies from 51 to 74 %, 7 to 14 % and 16 to 42 % with average values of 68, 11 and 22 % respectively in the core PE-3A. At 40-50 cm depth, the texture of the sediment is clayey sand whereas muddy sand represents the texture of the sediment from surface to 40 cm depth of PE-3A core (Fig 4.2). The water depth at the place was recorded as 1.3 meter. The down core variations of physical and textural parameters are shown in figure 4.3. The percentage of sand, silt and clay varies from 42 to 94 %, 3 to 30 % and 3 to 28 % with average values of 75, 13 and 12 % respectively in core PE-5. According to Folk's classification the sediments of PE-5 core exhibits muddy sand texture (shown in Fig. 4.2). The core was recovered from a water depth of 0.6 meter. Silt and clay have a similar trend showing opposite nature with sand as shown in figure 4.4.

The percentage of sand, silt and clay varies from 30 to 90 %, 3 to 37 % and 6 to 33 %, with average values of 66, 16 and 17 % respectively in the core PE-7. The sediments exhibit sandy mud, muddy sand and sand texture with depth (Fig. 4.2). There is a down core increase in the sand percentage (Fig. 4.5). Core PE-7 is taken from the central part of the mouth region of Chitrapuzha River at a water depth of 1.3 meter.

The percentage of sand, silt and clay varies from 70 to 96 %, 3 to 25 % and 1 to 9 %, with average values of 78, 15 and 5 % respectively in the core PE-11B. The sand percentage shows a decreasing trend and silt percentage is showing an increasing trend with depth (Fig. 4.6). The core is showing muddy sand to silty sand texture (Fig. 4.2). Core PE-11B was recovered from the lagoon near the Southern end of Willingdon Island where the water depth was 1.6 meter.

Sand, silt and clay percentage varies from 1 to 9 %, 16 to 39 % and 52 to 83 %, with average values of 4, 25 and 71 % respectively in the core PE-13. The sand percentage and silt percentage of the core sediments is showing a decreasing trend and the clay percentage is showing an increasing trend with depth (Fig. 4.7). The core sediments are showing mud to clay texture (Fig. 4.1). Core PE-13 was recovered from the eastern edge, in the central part of the lagoon at water depth of 1m.

Core PE-14 (Fig. 4.1) sediments showing muddy sand texture were recovered from the western edge, at the central part of the lagoon. The percentage of sand, silt and clay varies from 20 to 96 %, 0 to 51 % and 0 to 32 %, with average values of 75, 15 and 11 % respectively in the core PE-14 (Fig. 4.8). The water depth at the location of PE-14 core was recorded as 1.5 meter.

The percentage of sand, silt and clay varies from 14 to 34 %, 39 to 61 % and 18 to 38 %, with average values of 24, 50 and 26 % respectively in the core PE-17 (Fig. 4.9). The core sediments are showing sandy mud texture (Fig 4.1). The core was recovered from the central part of the lagoon where the recorded water depth was 2 meter.

The percentage of sand, silt and clay varies from 75 to 88 %, 4 to 12 % and 5 to 17 %, with average values of 82, 8 and 10 % respectively in the core PE-20 (Fig. 4.10). The core sediments are showing muddy sand texture (Fig. 4.1). The core was recovered from the western edge of the lagoon opposite to the mouth region of Muvattupuzha River, where the water depth was 1.5 meter.

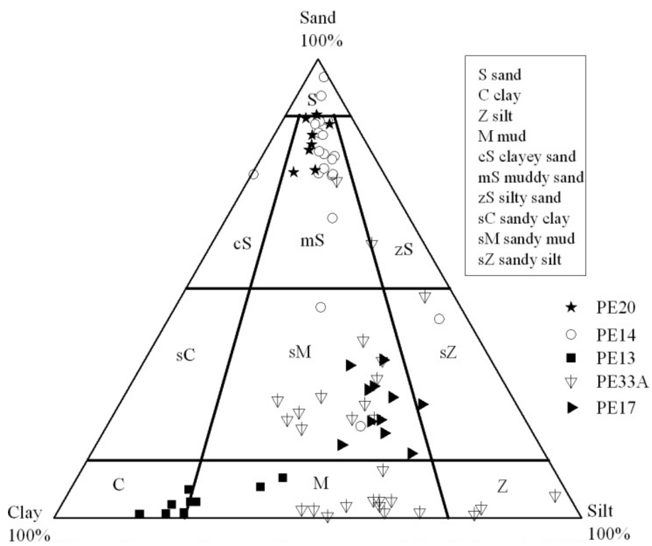


Figure 4.1: Ternary diagram showing the sand silt clay ratios in the Vembanad lagoon cores (after Folk et al., 1980)

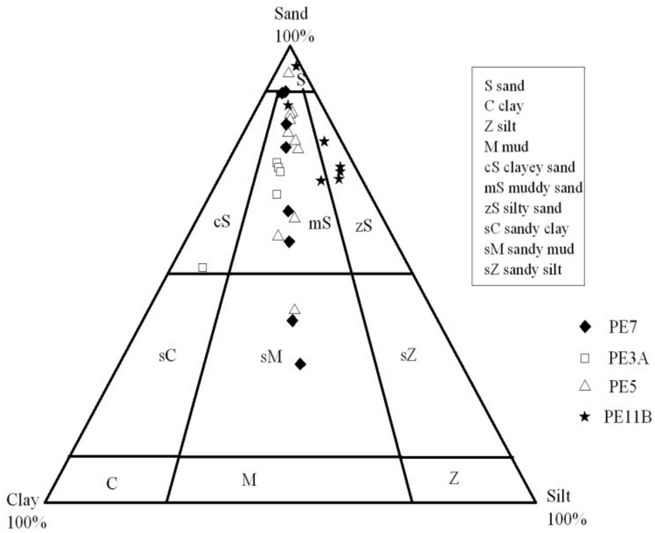


Figure 4.2: Ternary diagram showing the sand silt clay ratios in the Vembanad lagoon cores (after Folk et al., 1980)

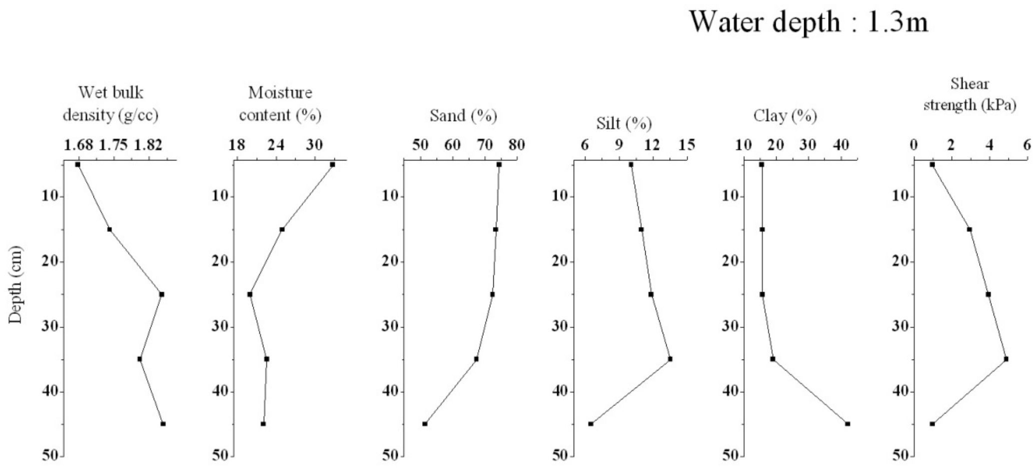


Figure 4.3: Down core variation of textural and geotechnical properties of an estuary core (No. PE3A)

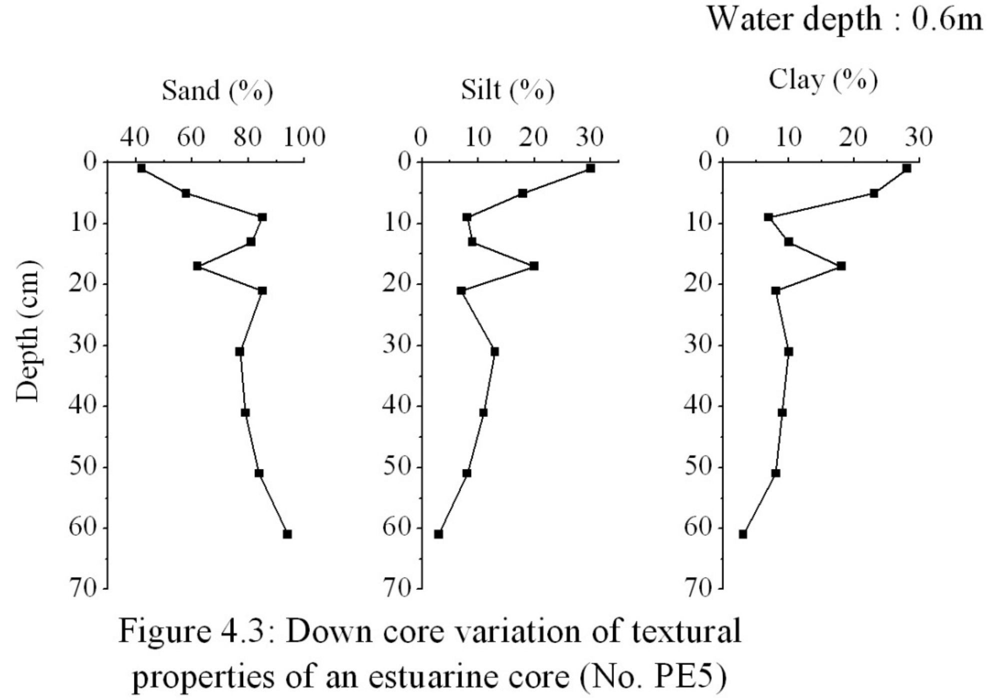


Figure 4.4: Down core variation of textural properties of an estuary core (No. PE5)

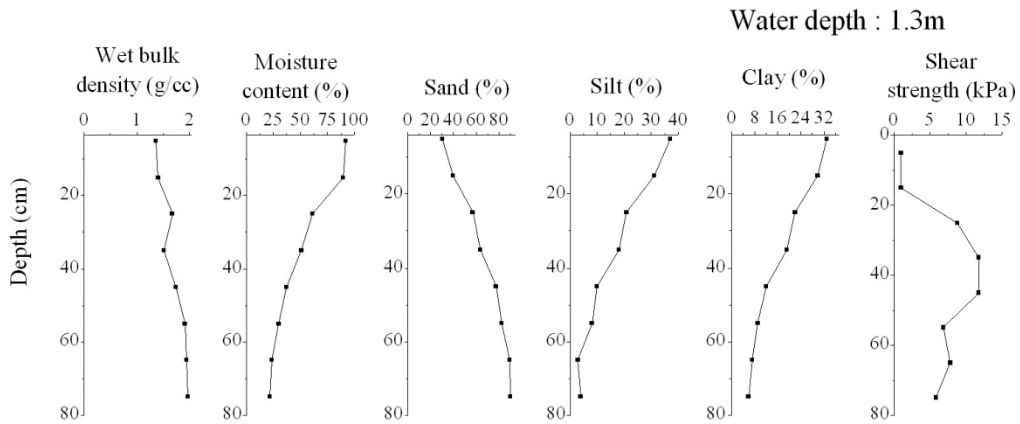


Figure 4.5: Down core variation of textural and geotechnical properties of an estuary core (No. PE7)

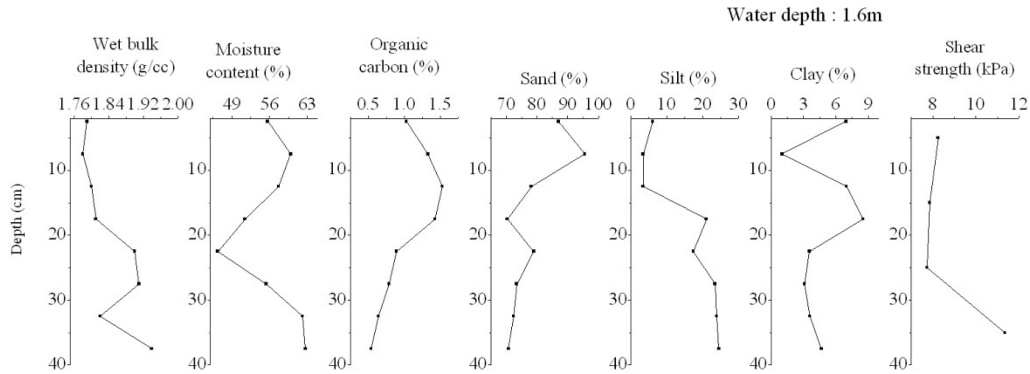


Figure 4.6: Down core variation of textural and geotechnical propertiesof an estuary core (No. PE11B)

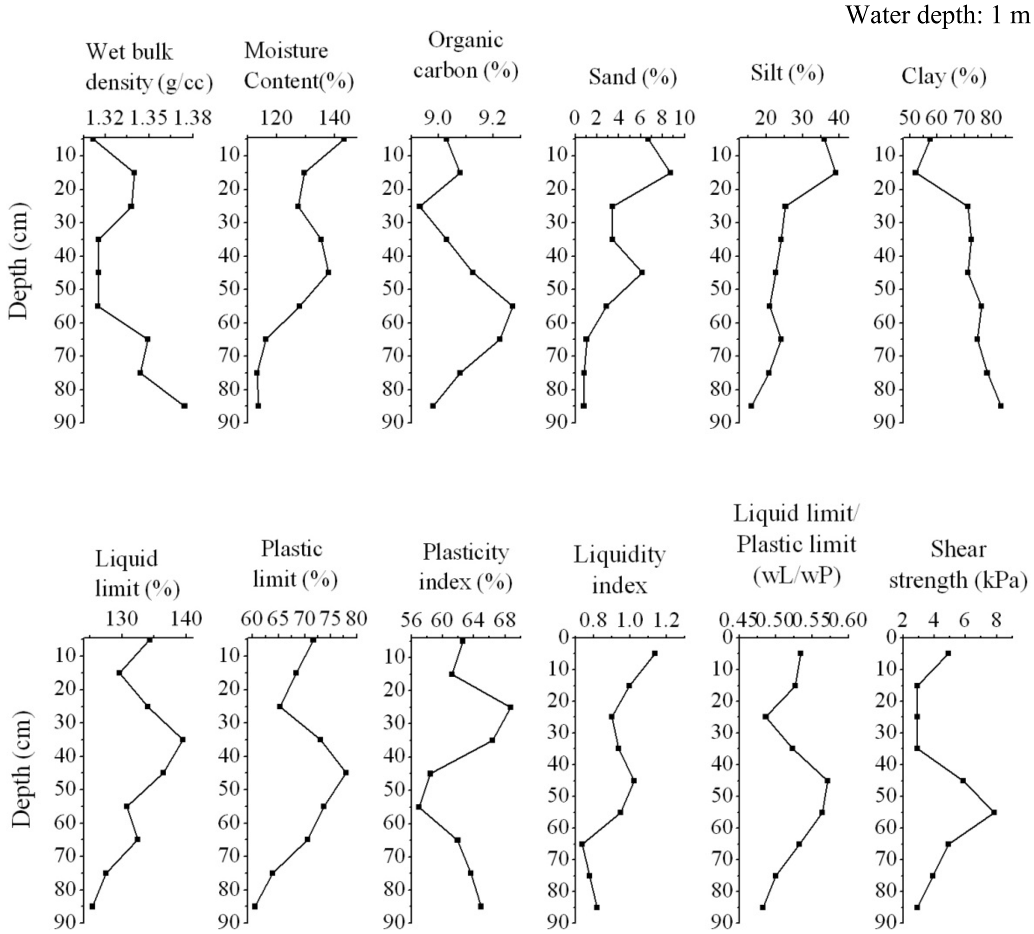


Figure 4.7: Down core variation of textural and geotechnical properties of an estuary core (No. PE13)

Water depth: 1.5 m

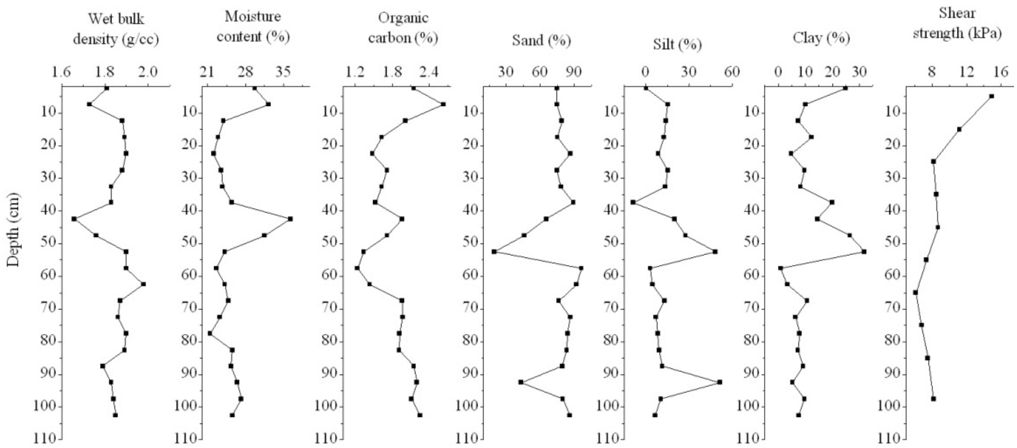


Figure 4.8: Down core variation of textural and geotechnical properties of an estuary core (No. PE14)

Water depth: 2 m

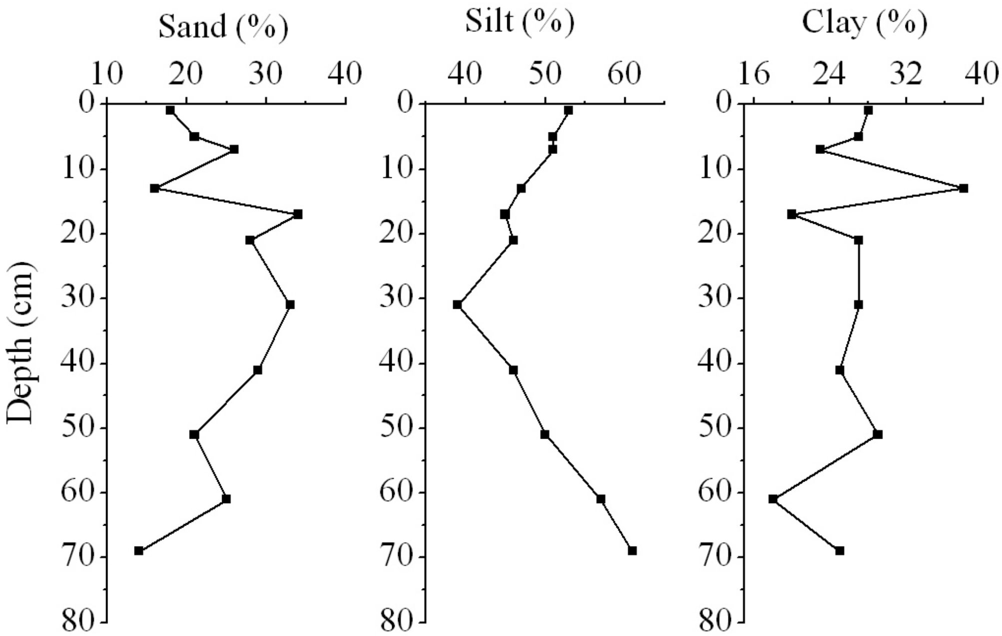


Figure 4.9: Down core variation of textural properties of an estuary core (No. PE17)

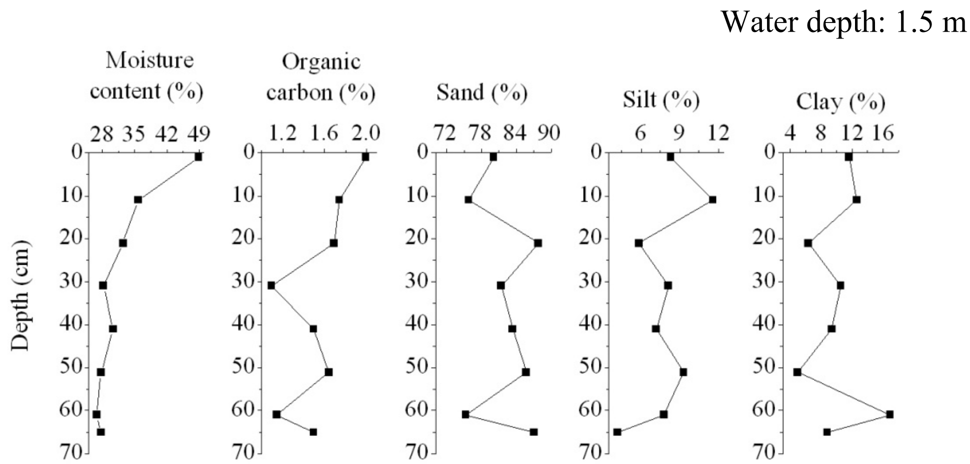


Figure 4.10: Down core variation of physical and textural properties of an estuary core (No. PE20)

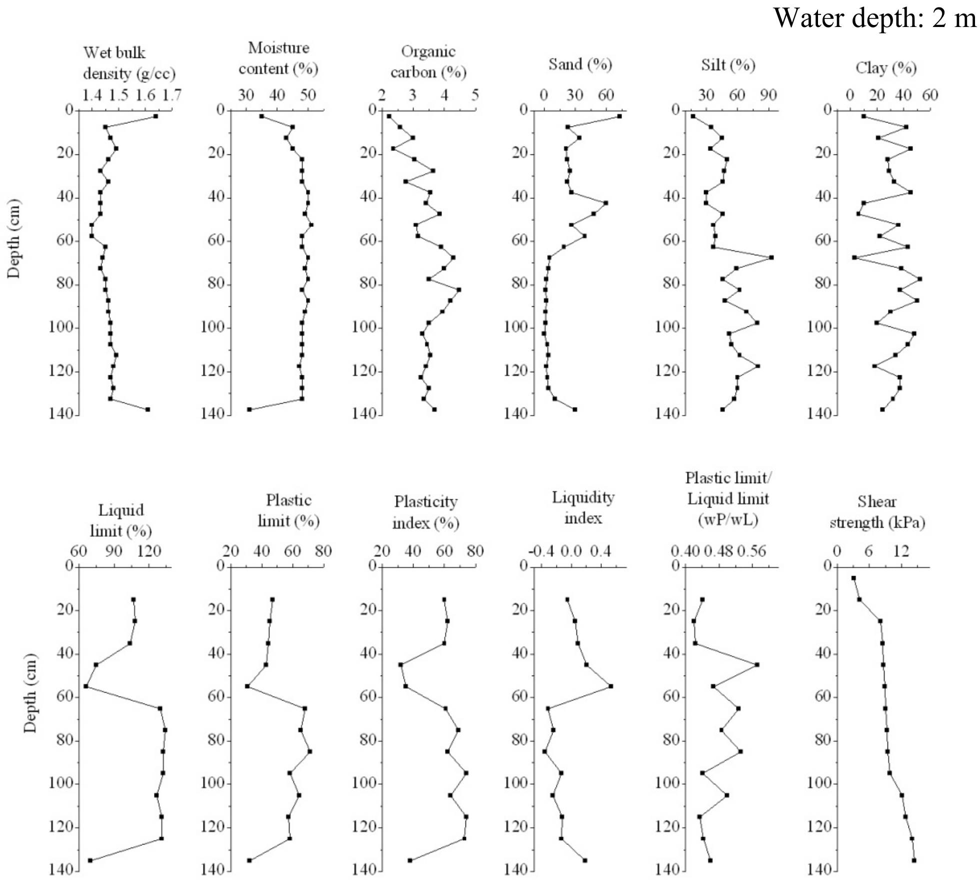


Figure 4.11: Down core variation of textural and geotechnical properties of an estuary core (No. PE33A)

The percentage of sand, silt and clay varies from 0 to 73 %, 17 to 93 % and 3 to 52 %, with average values of 18, 51 and 31 % respectively in the core PE-33A (Fig. 4.11). According to Folk's classification the sediments exhibit muddy sand, sandy mud, mud and silt texture (Fig. 4.1). Only the surface sediments and sediments at a depth of 40-45 cm exhibit muddy sand texture. Upto a depth of 60-65 cm the sediments exhibits a sandy mud texture. Beyond 65 cm depth the sediments are showing muddy and silty texture. The sand percentage decreases from top to bottom of the core. Core PE-33A was recovered from the northern side of the mouth region of Chitrapuzha River having the water depth of 1 meter.

Priju and Narayana (2007) reported that according to suite statistics the northern part of the northern sector of the lagoon reveals a dominant high-energy depositional condition, whereas a moderate to low energy depositional regime in the southern side of this sector. They observed a low to moderate energy regime in the central sector of the lagoon and a mixed energy regime towards the southern sector of the lagoon.

Core PE-3A and PE-5 were recovered from the northern sector of the lagoon and is a high energy environment and this can be attributed to the sediment texture of the cores. Core PE-11B, PE-13, PE-14 and PE-17 were recovered from the central part of the lagoon where the influence of anthropogenic activities is more such as dredging and inputs of industrial/domestic wastes, discharged from the industries situated along the Chitrapuzha River and southern arm of Periyar River. Thus the sediments show a mixed textural facies such as muddy sand, sandy mud, silty sand, mud, and clay texture. Core PE-20 was recovered from the southern sector of the lagoon and can be a high energy environment because of the proximity to the mouth region of the Muvattupuzha River. This can be attributed to the muddy sand texture of the core sediments.

4.2.2. Sediment texture of nearshore environment

The textural nomenclature of nearshore grab sediments based on Folk classifications is shown in figure 4.12. Textural characteristics shown by the nearshore grab samples are variable from mud to sandy mud. The sand % of the grab sample ranges from 1 to 86 % with an average of 32 %, silt ranges from 7 to 53 % with an average of 32 %, and clay ranges from 6 to 73 % with an average of 36 %. The grab samples S-1, S-16, S-34, S-47 and S-48 exhibit a sandy mud texture having sand percentage of 14, 48, 18, 35 and 32 respectively. The samples were recovered from water depths of 5 m, 5 m, 30 m, 5 m and 10 m respectively. Samples S-2, S-3, S-31, S-32 and S-33 are showing mud texture having a sand percentage 1 in S-1 and S-2 samples and 8, 6 and 0 sand percentages in S-31, S-32 and S-33 samples. S-2 and S-32 samples were recovered from 10 m water depth. The water depths of other samples are 20 m, 5 m and 15 m for S-3, S-31 and S-33 respectively. The grab samples S-4, S-17, S-19 and S-50 exhibit muddy sand texture with sand percentages 82, 73, 86 and 71 respectively in the samples. Samples S-4 and S-50 were recovered from a water depth of 30 m, S-17 and S-19 from a water depth of 5 m.

Core G-3 and G4 exhibit muddy texture (Fig.4.13, 4.15 & 4.16) throughout with high silt and clay and low sand percentages. Sand ranges from 1 to 9 % with an average of 2 %, silt from 34 to 69 % with an average of 68 %, and clay from 26 to 65 % with an average of 45 %. The water depth recorded at G-3 & G4 core locations was 20 meter.

The percentage of sand, silt and clay varies from 0 to 32 %, 41 to 52 % and 22 to 58 %, with average values of 5, 46 and 49 % respectively in the core

G-6. The surface sediments up to a depth of 10 cm are showing a sandy mud texture and below that the sediments exhibit mud texture (Fig. 4.13 & 4.17). Here, the water depth recorded was 10 meter.

The percentage of sand, silt and clay varies from 3 to 5 %, 27 to 78 % and 18 to 69 %, with average values of 3, 48 and 49 % respectively in the core G-7 with a water depth of 20 meter. The sediments are exhibiting mud texture (Fig. 4.13 & 4.18). The percentage of sand, silt and clay varies from 0 to 13 %, 21 to 90 % and 9 to 72 %, with average values of 2, 47 and 51 % respectively in the core G-8 with the water depth of 30 meters. The sediments are exhibiting mud texture throughout the core except at 78-80 cm and 0-2 cm showing sandy clay and sandy mud texture respectively indicating episodes of flood event (Fig. 4.13 & Fig. 4.19).

The sand percentage in core G-9 ranges from 81 to 86 %, silt percentage 2 to 9 % and clay percentage 7 to 13 % with average values of 84, 5 and 11 % respectively and the core was recovered from a water depth of 29 meters. The core sediments are having clayey sand to muddy sand texture sequence upward (Fig. 4.14 & Fig. 4.20).

The percentage of sand, silt and clay varies from 0 to 2 %, 43 to 71 % and 28 to 56 %, with average values of 1, 56 and 43 % respectively in the core G-10 with a water depth of 20 meter. According to Folk's classification the sediments are exhibiting mud texture (Fig. 4.14 & Fig. 4.24). G-11 core, having a water depth of 10 meter, shows mud texture with high content of silt and clay and lower sand percentage (Fig. 4.25 & 4.14). The sand ranges from 0 to highest 18 % at the surface level showing a sandy mud texture with an average of 2 %. Silt ranges from 28 to 58 % with an average of 44 %, clay ranges from 31 to 71 % with an average of 53 %.

Core GC-20 shows mud to sandy mud texture (Fig.4.14 & Fig. 4.22) when moved downward. The water depth recorded was 11 meter at core GC-20 where the sand percent ranges from 3 to 26 % with an average of 16 %, silt ranges from 44 to 66 % with an average value of 53 %, and clay ranges from 19 to 53 % with an average of 31 %. GC-22 core shows mud to sandy mud textural downcore variation where sand percent ranges from 3 to 65 % with an average of 14 %, silt ranges from 34 to 82 % with an average of 48 %, and clay ranges from 8 to 61 % with an average of 38 % (Fig. 4.14 & Fig. 4.23).

Muddy texture of the GC-16 core is altered at the depths 10-12 cm, 80-82 cm, and 150-152 cm by silt showing episodes of flood event (Fig 4.14 & Fig. 21). The sand percent ranges from 0-2 % with an average of 2 %, silt ranges from 36 to 80 percent with an average of 48 %, and clay ranges from 19 to 64 % with an average of 51 %. The water depth recorded at GC-16 core location was 19 meter.

Core G-9, G-10 and G-11 were recovered from the northern side of the mouth region of Periyar River at Munambam. Core G-11, G-10 and G-9 were recovered from 10 m, 20 m, and 30 m water depths respectively. Core G-6, G-7 and G-8 were recovered from the mouth region of Periyar River at Munambam from water depths of 10 m, 20 m and 30 m respectively.

In core G-4 and G-3, the sediments exhibit mud texture which was recovered from the mouth region of Cochin outlet of Vembanad lagoon from 20 m water depth. Core GC-16, sediments exhibit silt and mud texture, mostly having a mud texture. The core was recovered further south of the Cochin outlet of Vembanad lagoon from a water depth of 20 m. Core GC-20 and GC-22 exhibit sandy mud and mud texture which were recovered from the southern part of the study area near Alappuzha. Core GC-22 was recovered from a water depth of 5 m and GC-20 from a water depth of 10 m. Except core G-9 all other core samples are having sandy mud to mud texture. The surface sediments along the nearshore region are showing mud, sandy mud and muddy sand texture.

The sediment distribution in the nearshore regions mostly depends on the monsoonal climate, waves and longshore currents. In a typical coastal sediment distribution system, the nearshore sand follows offshore mud (Reineck and Singh, 1975). The nearshore area is blanketed mostly with the sandy mud in the shallower region and muddy sand in the deeper region. The dominance of the muddy sand and sandy mud deposits in the surface sediments of the nearshore region of Cochin is the reflection of the high wave energy condition that is prevalent in the coastal region of Kerala. Earlier workers have also observed high sand content in the surface sediments of nearshore region, particularly up to 5 m water depth, along the Kerala Coast (Manojkumar et al.,1998). Zonality of different sediment types based on size class, sorting and composition may be two dimensional, but more typically occurs along one dimension, usually either in the cross-shore or alongshore direction or with respect to dominant wave or current direction (Holland and Elmore 2008).

The core sediments in the nearshore region are dominated by mud and sandy mud texture which can be attributed to a low energy environment. The dominance of the coarse sediment close to the river mouth reveals the influence of the river circulation and their deposition at the mouth region.

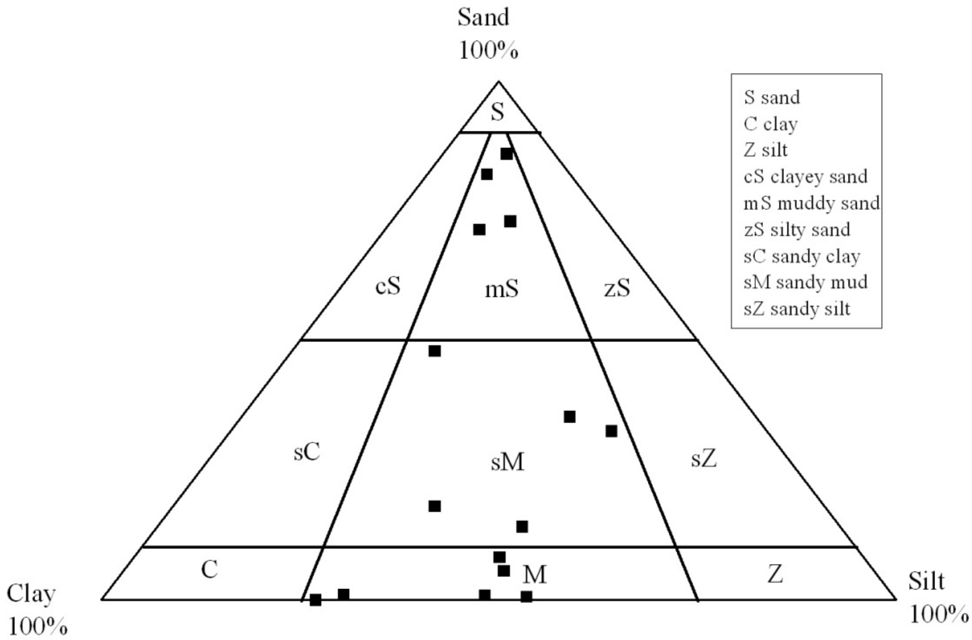


Figure 4.12: Ternary diagram showing the sand silt clay ratios of nearshore grab sediments (after Folk et al., 1980)

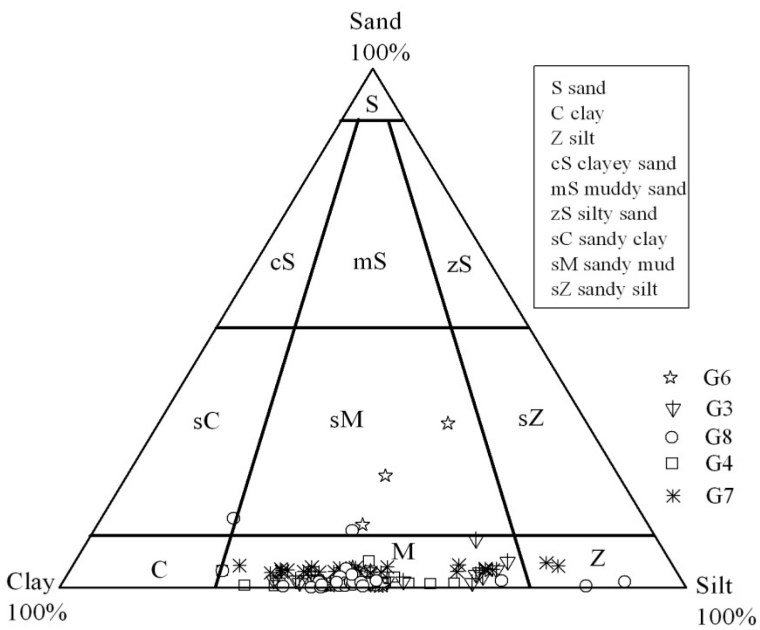


Figure 4.13: Ternary diagram showing the sand silt clay ratios of nearshore core sediments (after Folk et al., 1980)

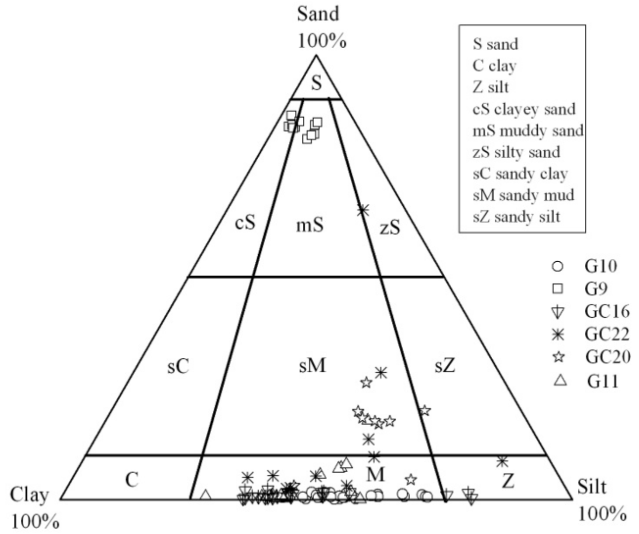


Figure 4.14: Ternary diagram showing the sand silt clay ratios of nearshore core sediments (after Folk et al., 1980)

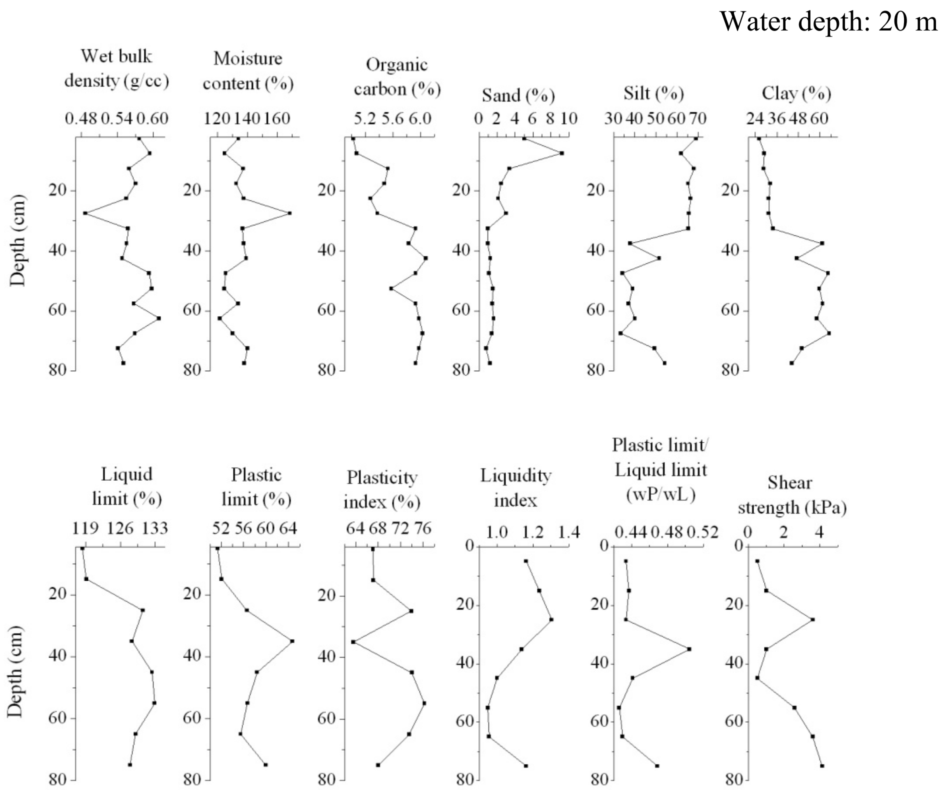


Figure 4.15: Down core variation of textural and geotechnical properties of a nearshore core (No. G3)

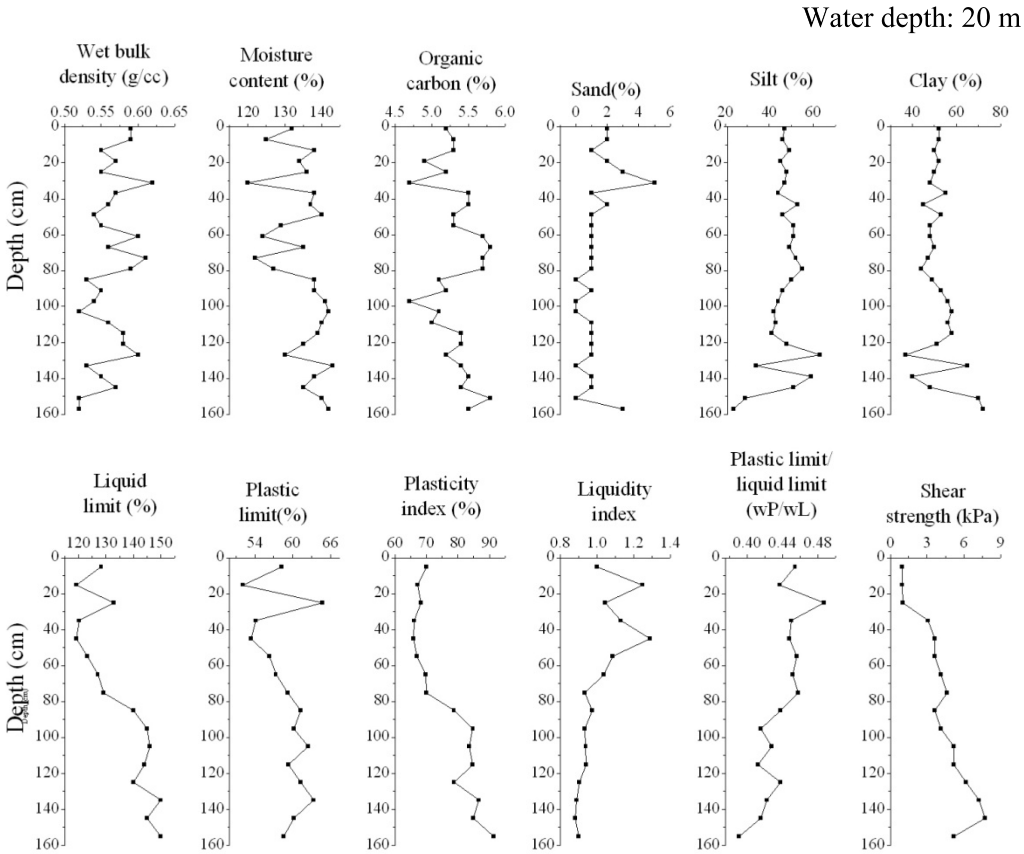


Figure 4.16: Down core variation of textural and geotechnical properties of a nearshore core (No. G4)

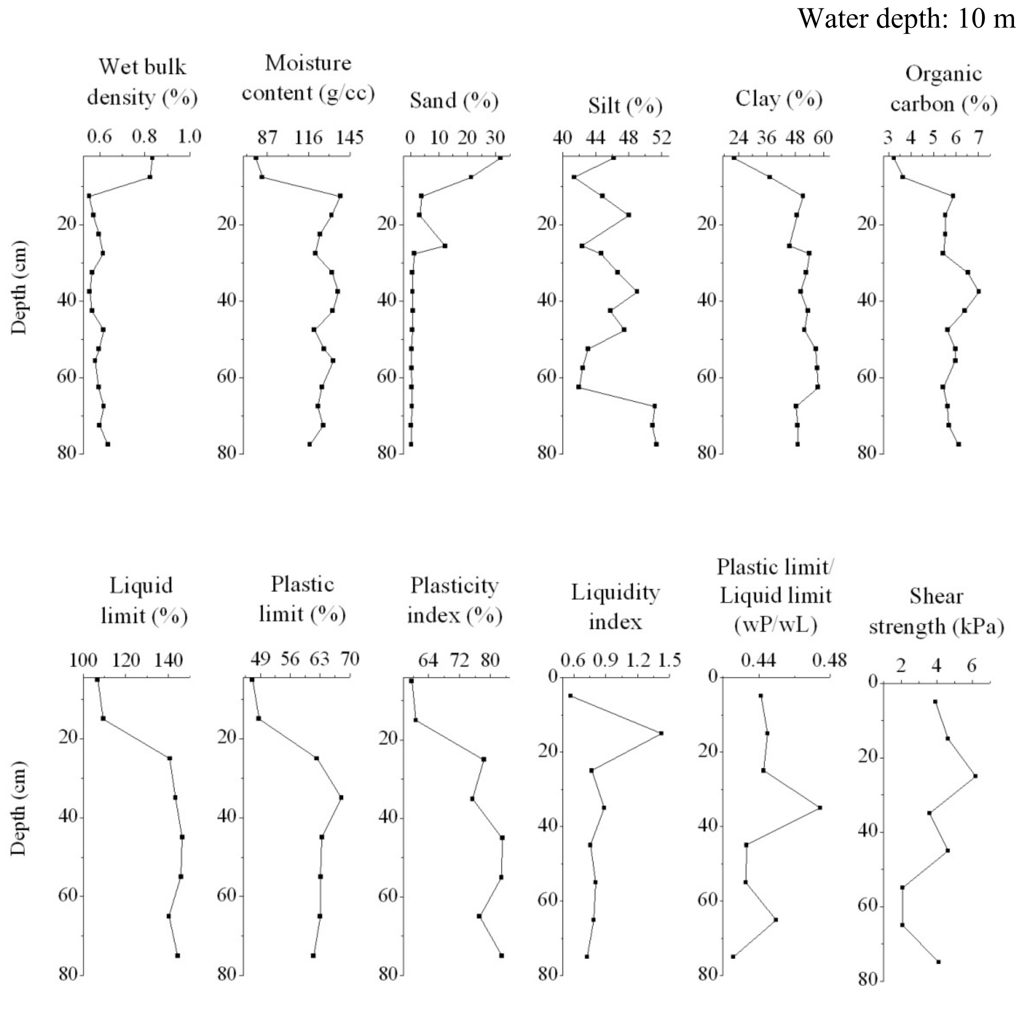


Figure 4.17: Down core variation of textural and geotechnical properties of a nearshore core (No. G6)

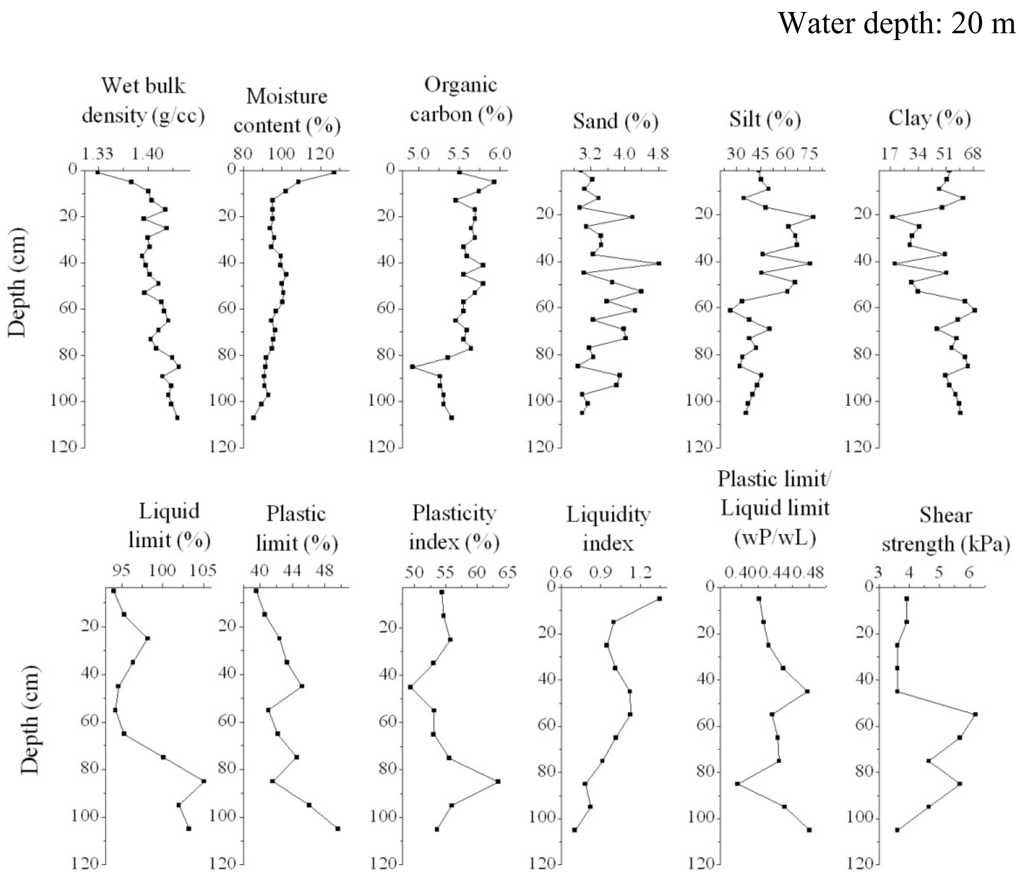


Figure 4.18: Down core variation of textural and geotechnical properties of a nearshore core (No. G7)

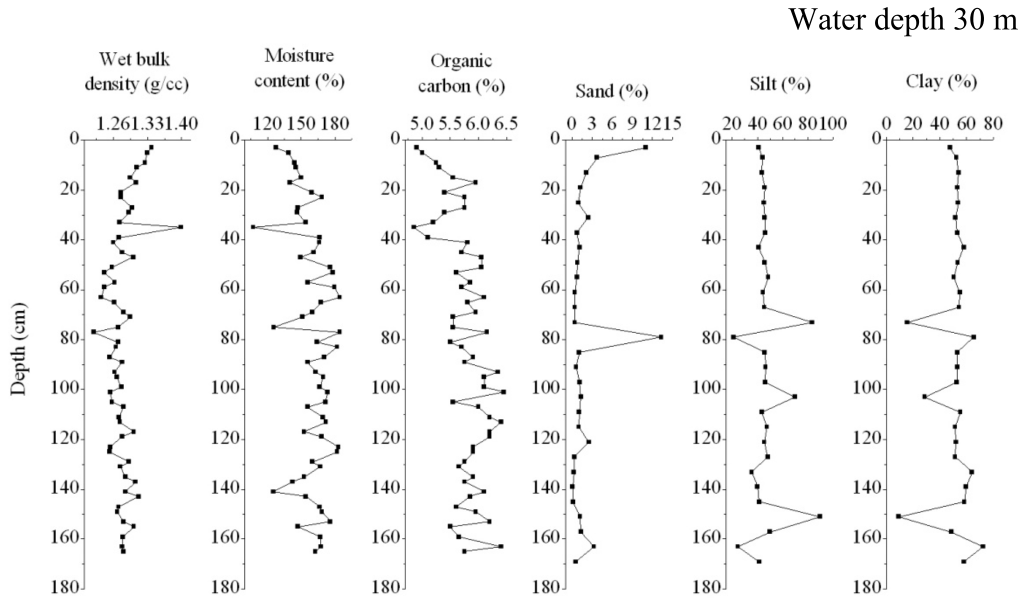


Figure 4.19: Down core variation of physical and textural properties of a nearshore core (No. G8)

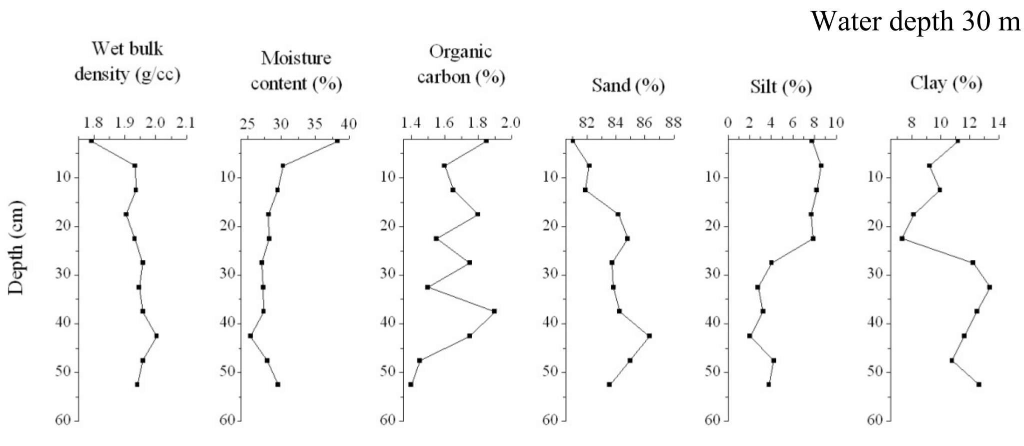


Figure 4.20: Down core variation of physical and textural properties of a nearshore core (No. G9)

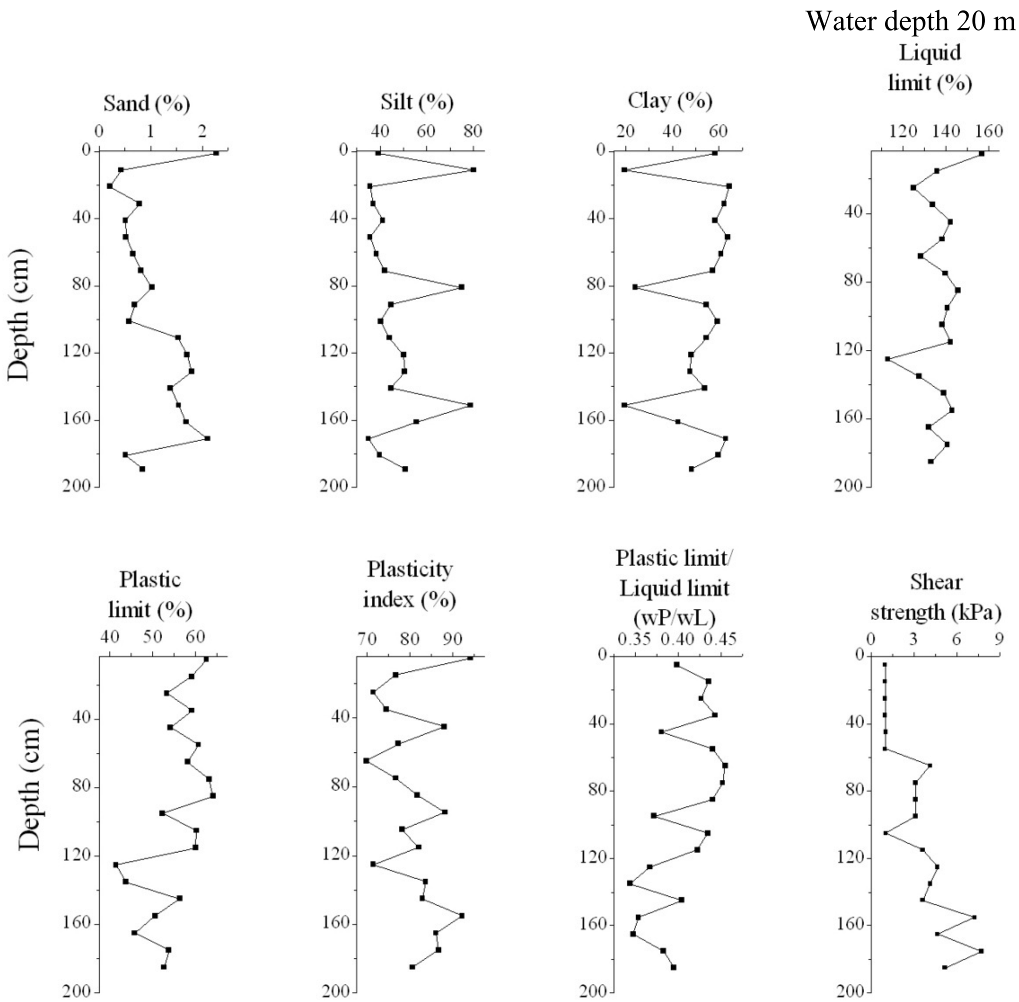


Figure 4.21: Down core variation of textural and geotechnical properties of a nearshore core (No. GC 16)

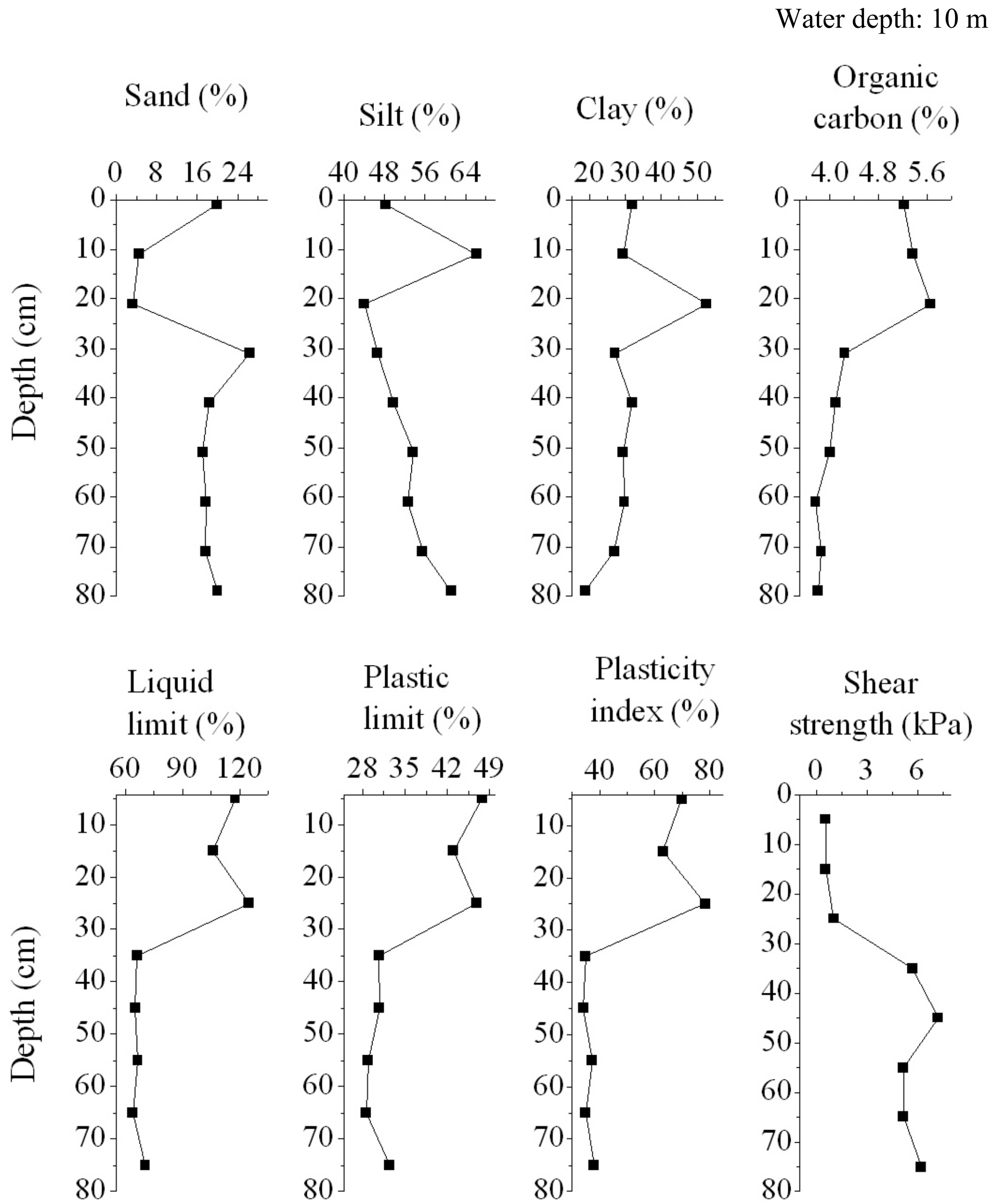


Figure 4.22: Down core variation of textural and geotechnical properties of a nearshore core (No. GC20)

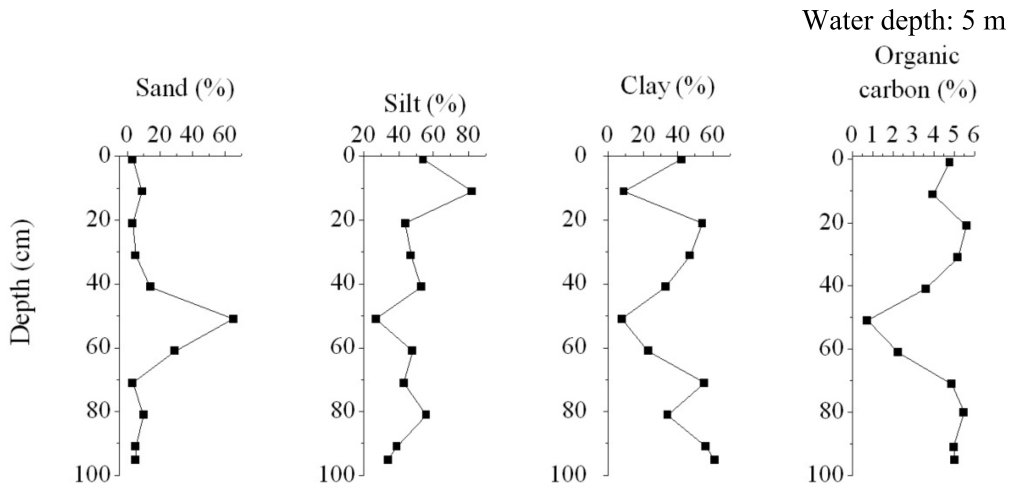


Figure 4.23: Down core variation of organic carbon and textural parameters of a nearshore core (No.GC22)

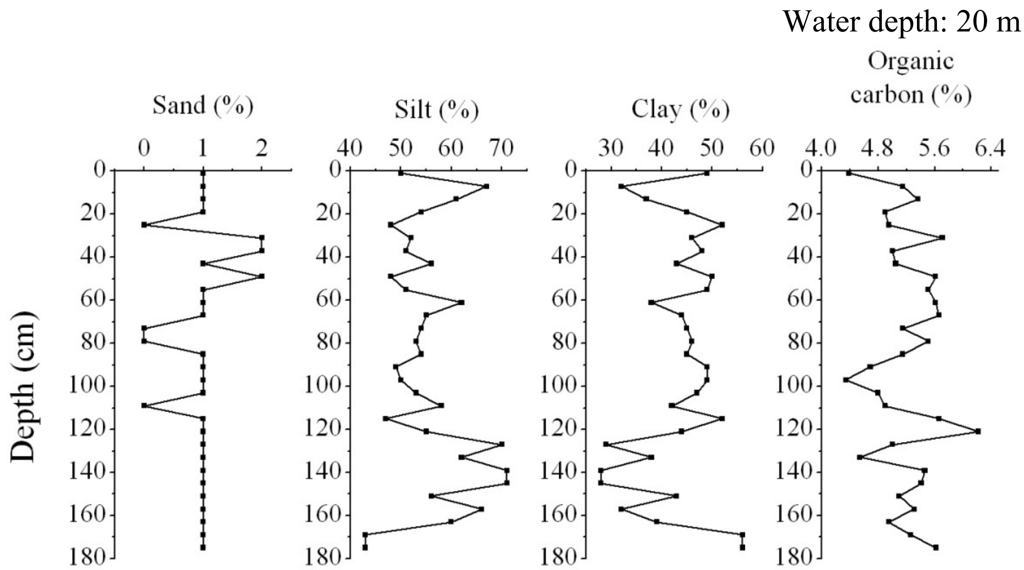


Figure 4.24: Down core variation of organic carbon and textural parameters of a nearshore core (No. G10)

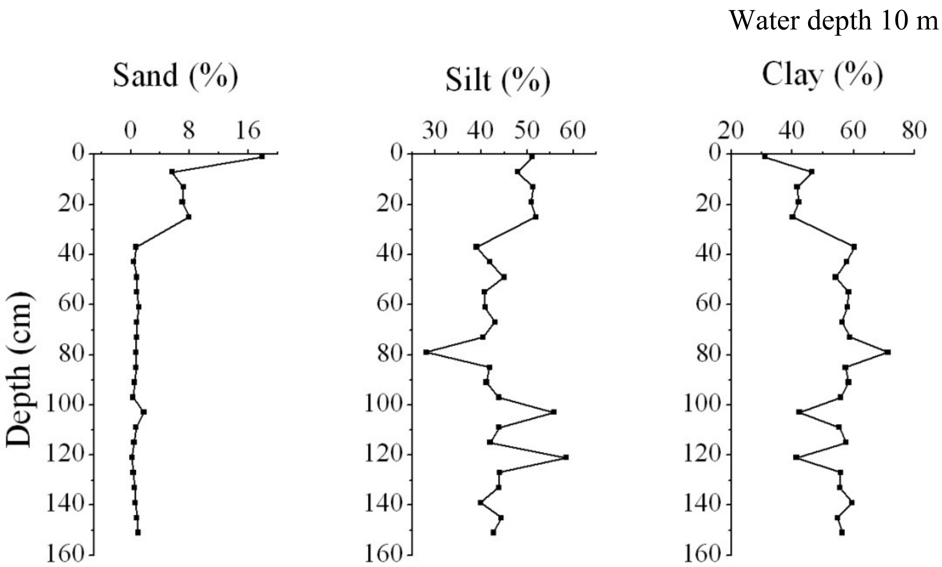


Figure 4.25: Down core variation of textural parameters of a nearshore core (No. G11)

4.3. Organic carbon

In estuaries and nearshore areas, the organic carbon is contributed by terrestrial processes. The global export of organic carbon (OC) is linked to the total flux of terrestrial sediment to the oceans. Continental margins are characterized by high levels of OC storage during present conditions as well as in the geologic past. An estimated 80–85% of carbon burial is currently occurring on continental margins, mainly off river-dominated coasts (Berner, 1982; Hedges and Keil, 1995). Rivers provide the major conduits for the transfer of terrestrial OC to marine sediments (McKee, 2003). In addition to this, a substantial amount of organic carbon rich sewage sludges are disposed through municipal drainages near the coastal areas.

Rivers are major conduits for the transport of terrestrial organic matter to the coastal marine environment (Meybeck, 1982; Spitzy and Ittekkot, 1991). However, only a small fraction of terrestrial organic matter reaches the open ocean based on biomarker and isotopic evidence (Hedges et al., 1997; Benner, 2004). Therefore, the degradation, transformation, and burial of terrestrial organic matter through physicochemical, microbial, and photochemical processes in estuaries are important (e.g., Benner and Opsahl, 2001; Goñi et al., 2005; Dagg et al., 2008). Indeed, the biogeochemical cycling of organic matter in estuaries at the land ocean interface has received increasing attention (e.g., Meybeck et al., 1988; Guo and Santschi, 1997; Raymond and Bauer, 2001a; Dagg et al., 2007; Spencer et al., 2007; Sleighter and Hatcher, 2008; Wang et al., 2010).

Organic carbon plays a significant role in the genesis of hydrocarbon. Sedimentation followed by diagenetic decomposition of organic carbon can alter the Eh-pH conditions of the aquatic environments. Further, it provides the main energy source for the heterotrophic organisms and also acts as sink as well as source for various metals under different geochemical set up.

4.3.1. Organic carbon in lagoonal environment

The percentage of organic carbon ranges from 0.54 to 9.27 in the core sediments. Organic carbon increases with clay percentage. Core PE-13 sediments have the highest organic carbon percentage. In this core the value ranges from 8.93 to 9.27 % with an average value of 9.1 % (Fig. 4.7). Hydrodynamic features such as tidal activity may be the main factor determining the fate of organic matter at this site. Organic matter and fine grained sediments may be accumulated and preserved at this site due to the low tidal action. Fine grained sediments act as a sink of organic matter. Core PE-11B sediments are showing a decreasing trend in the percentage of organic carbon from top to bottom. The value ranges from 0.54 to 1.53 % with an average value of 1.02 % (Fig. 4.6). Core PE-14 sediments have values ranging from 1.24 to 2.63 % (Fig. 4.8) with an average value of 1.86 % which is slightly higher than that of core PE-11B sediments. Core PE-20 sediments are having values ranging from 1.09 to 1.99 with an average value of 1.54 % (Fig. 4.10). Core PE-33A sediments range from 2.22 to 4.48 with an average value of 3.43%. The highest value of 4.48% is at a depth of 80-85 cm where the sand percentage is 1 (Fig. 4.11). Thus the antipathetic relationship of organic carbon with grain size is evident from these observations. The grain size of the sediments depends on the hydrodynamic conditions prevailing in the region. Organic carbon percentage is showing a positive correlation with silt and clay percentage. In addition, the Vembanad lagoon is adjacent to a highly industrialized area where a high supply of organic carbon is due to the disposal of sewage sludges transported to the lagoon.

4.3.2. Organic carbon in nearshore environment

The grab samples are showing variable organic carbon percentage and strongly depend on the grain size of the sediments. The organic carbon (OC) percentage ranges from 0.64 to 6.65 in the grab samples. The grab sample S-19 is having an OC percentage of 0.64 with a sand percentage of 86. The muddy sand samples are having organic carbon percent ranging from 0.64 to 1.97 where as the sandy mud and mud samples are having OC percent ranging from 1.88 to 6.65.

All cores recovered from the nearshore environment except core G-9 have organic carbon ranging from 4 to 7%. The organic carbon in core G-9 which is a sandy core is very less and the value ranges from 1 to 2% (Fig. 4.20). Core GC-20 and GC-22 were recovered further south away from the Cochin and Munambam outlets of Vembanad lagoon and are having slightly lower values of organic carbon percentage than other core sediments (Fig. 4.22 & Fig.

4.23). The value ranges from 3.76 to 5.66 % and 0.73 to 5.61 % with average values 4.44 and 4.22 % for core GC20 and GC22 sediments respectively.

Rivers supply land derived macro-nutrients (both anthropogenic and natural) to the coastal areas which in turn increases the primary productivity. Rivers also are major conduits for the transfer of terrestrially derived organic carbon to the coastal areas. Thus the higher value of organic carbon percentage in the core sediments is due to their proximity to the Cochin and Munambam outlet.

4.4. Geotechnical studies of sediment samples

4.4.1. Geotechnical studies of lagoonal environment

Water content

This is a measurement of how much moisture a soil is holding in its void spaces. It has a great impact on the consistency of the soil, its density, and its compaction. Water content is one of the most crucial factors influencing soil and rock strength. The water content of the core sediments ranges from 20 % to 143 % (Fig. 4.3 to Fig 4.11). Water content decreases with increase in sand percentage. In core PE-13 sediments the water content ranges from 113 % to 143 % and is having the highest water content among all the core sediments. Core PE-14 sediments have the lowest water content ranging from 22.17-36.35 % with an average of 25.83 %. In core PE-33A the sediments have water content ranging from 31 to 51 %. Increase of organic carbon and amount of clay sized material increases the water content in the sediments as shown in the figures.

Wet bulk density

The wet bulk density of core sediments ranges from 1.21 to 1.98 g/cc (Fig. 4.3 to Fig 4.11). Wet bulk density increases with increase in sand percentage and decreases with increase in water content. The wet bulk density of core PE-13 ranges from 1.31 to 1.37 g/cc and is having the least average wet bulk density. The wet bulk density of PE-14 ranges from 1.73 to 1.98 g/cc and is having the highest average wet bulk density of 1.85 g/cc.

Atterberg limits

It is widely accepted that liquid and plastic limit tests are very important geotechnical tests which can give valuable information about the engineering behaviour of fine grained soils (Spagnoli et al. 2012). Water content of cohesive soils are usually correlated to empirically defined boundaries between states of consistency i.e. non-plastic, plastic, and viscous liquid, are known as Atterberg Limit (Dias and Alves, 2009). The liquid and plastic limit reflects the ability of a sediment particle to attract water to its surface and increase with the particle tendency to hold water. It has been shown that these two limits and the plasticity index increase with both clay and organic content (Odell et al., 1960). These properties are most commonly affected by the physico-chemistry of clay-water system (Rosenquist, 1962) and the concentration of organic carbon (Sodorblom 1966; Rashid and Brown, 1975). Differences in plasticity are due to the nature of the sediment, grain size, textural and compositional characteristics (Chassefiere and Monaca, 1987).

(a) Liquid limit

The liquid limit is defined as the water content at which a standard V-groove cut in the soil just closes when it is allowed to fall in a standard way using Casagrande apparatus (Claudio et al., 2009). The liquid limit is also an indicator of the quantity of water held as double-layer water in clay minerals (Sridharan et al. 1986, 1988); a change in the liquid limit is related to a change in the double-layer thickness of clay minerals. The liquid limit values of most of the core sediments could not be identified because of the high sand percentage in the sediments. The liquid limit values ranges from 66 to 139 % in the core sediments (Fig. 4.7 & Fig. 4.11). It is higher than the corresponding water content in the sediments which means that the sediments are in their plastic state.

(b) Plastic limit

Interest in clays has increased in recent years due to their physical–chemical and plastic properties. Since Atterberg (1911) much work has been done in sediments in an attempt to evaluate the influence of the various factors involved in the plasticity of clay samples, such as their mineralogical composition, size distribution of particles, organic carbon etc. Plasticity is the outstanding property of clay–water systems. It is the property a substance has when deformed continuously under a finite force. When the force is removed or reduced, the shape is maintained. Mineralogical composition, particle size distribution, organic substances and additives can affect the plasticity of clays. Several measuring techniques and devices were proposed to determine the optimal water content in a clay body required to allow this body to be plastically deformed by shaping (Andrade et al., 2011). The plasticity of clays is related to the morphology of the plate-like clay mineral particles that slide over the others when water is added, which acts as a lubricant. As the water content of clay is increased, plasticity increases up to a maximum, depending on the nature of the clay (Handle, 2007).

The plastic limit is showing a similar trend with that of the liquid limit of the core sediments as shown in the figure 4.7 & 4.11. The plastic limit values range from 31 to 78% in the core sediments recovered from the lagoon. The plastic limits of the core PE 13 samples are lower than the corresponding water content of the samples. Thus core PE13 samples are in the plastic state. The plastic limit values are lower than the water content up to 60 cm depth in core PE33A. Below 60cm depth the plastic limit values are higher than the water content. Therefore up to 60 cm depth the sediments are in its plastic state. Beyond 60 cm depth the sediments are in the semi-solid state or they are cohesive soils.

(c) Plasticity index

The plasticity index (IP) is the difference between the liquid and the plastic limits, and defines the range of water contents within which the soil is in the plastic state, in which it can be molded without cracking.

The plasticity index of the sediments of core PE-33A ranges from 32 to 74 % with an average value of 59 % (Fig 4.11) and the plasticity index of core PE-13 ranges from 57 to 69 % with an average value of 63 % (Fig. 4.7).

The liquid limit values were plotted against plasticity index values, i.e., as plasticity chart in figure 4.25, to understand the level of plasticity. Core PE-13 sediments are lying below the A-line in the plasticity chart and are extremely high plasticity mud according to Cassagrande's classification of soil. Core PE-33A sediments are lying below and above the A-line in the plasticity chart. The sediments up to a depth of 40 cm are extremely high plasticity mud and below that depth is extremely high plasticity clay. The activity chart is shown in figure 4.26 (clay % versus plasticity index). The figure shows Skemptoris (1953) classification of clays as 'active', 'normal', and 'inactive clays'. Core PE13 sediments have an activity ranging from 0.75 to 1.17 with an average value of 0.90. Thus core PE13 sediments can be described as soils with normal activity. Core PE33A sediments exhibits activity values ranging from 1.21 to 4 with an average value of 2.07. Thus core PE33A sediments are showing high activity.

(d) Liquidity index

The relative consistency of a cohesive soil in the natural state can be defined by a ratio (Eq. 1) called the liquidity index (I_L):

$$I_L = (w_n - w_P)/I_P \qquad (1)$$

where w_n is the natural moisture content in the sediment w_P is the plastic limit and I_P is the plasticity index.

In this equation, if $I_L = 1.0$, the soil is at its liquid limit and if $I_L=0$, the soil is at its plastic limit. The liquidity index is believed by many authors to more usefully reflect the properties of plastic soil than the generally used plastic and liquid limits (wP and wL). The Atterberg tests are carried out on remoulded soil in the laboratory, but the same soil, in its in-situ state, may exhibit a different consistency at the same moisture content due to sensitivity effects. For this reason, it does not necessarily imply that a soil found to have a liquid limit of 50% will be in liquid state if its in-situ water content is also 50%. If w_n is greater than the test value of w_L then I_L is >1.0 and it is obvious that if the soil was remoulded it would become a slurry. Such a soil would probably be unconsolidated sediment with shear strength (Su) value in the order of 15–50 kPa. Most cohesive soil deposits have I_L values within the range 1.0–0.0. The lower the value of w_n, the greater the amount of compression that must have taken place and the I_L will be nearer to zero. If w_n is less than the test value of the plastic limit, then I_L ~0.0 and the soil cannot be remoulded (as it is outside the plastic range). In this case the soil is most likely to be compressed sediment.

Soil in this state will have a Su value varying from 50 to 250 kPa (Smith 1990, Yilmaz 2000).

Rominger and Rutledge (1952) and Means and Parcher (1963) consider that the liquidity index provides a reliable indication of the degree of consolidation of clayey soils. Means and Parcher (1963) reported that $I_L = 0$ for overconsolidated clays and the value of the liquidity index is negative for highly overconsolidated clays.

The liquidity index of core PE-33A ranges from -0.35 to 0.53 with an average value of -0.04 and the liquidity index in core PE13 ranges from 0.74 to 1.14 having an average value of 0.92. PE-33A sediment are over consolidated clays and PE 13 sediments are normally consolidated clay according to Means and Parcher classification based on the degree of consolidation of clays.

Vane shear

The shear strength of a soil depends on many factors which are, stress history, soil composition, water content, degree of saturation, soil structure, void ratio, drainage conditions, isotropic media in the soil, rate of loading.

The shear strength values ranges from 0.98 to 14.96 KPa in the estuary cores (Fig 4.3. Fig. 4.5 to Fig 4.8 & Fig 4.11). The cores are shallow having length ranging from 42 to 156 cm. Therefore the sediment samples are less consolidated and have low shear strength values. The longest core is PE-33A having a length of 156 cm. In this core the shear strength value increases with depth. The value ranges from 3.09 to 14.45 KPa. Based on the liquidity index the sediments of core PE-33A are over consolidated clays and thus having higher shear strength values with depth.

4.4.2. Geotechnical studies of nearshore environment

Water content

The water content of the sediments of core G3 ranges from 121 to 168 %, G6 ranges from 79 to 138 %, G7 ranges from 85 to 127 %, G8 ranges from 108 to 184 %, and G9 ranges from 25 to 38 % (Fig. 4.15 to Fig 4.20). Thus Core G9 sediments with the highest wet bulk density have the lowest water content. The water content is also inversely related to the sand percentage. Core G9 sediments having higher sand percentage is having the lowest water content. All other core sediments are having higher silt and clay percentages thus have higher water content.

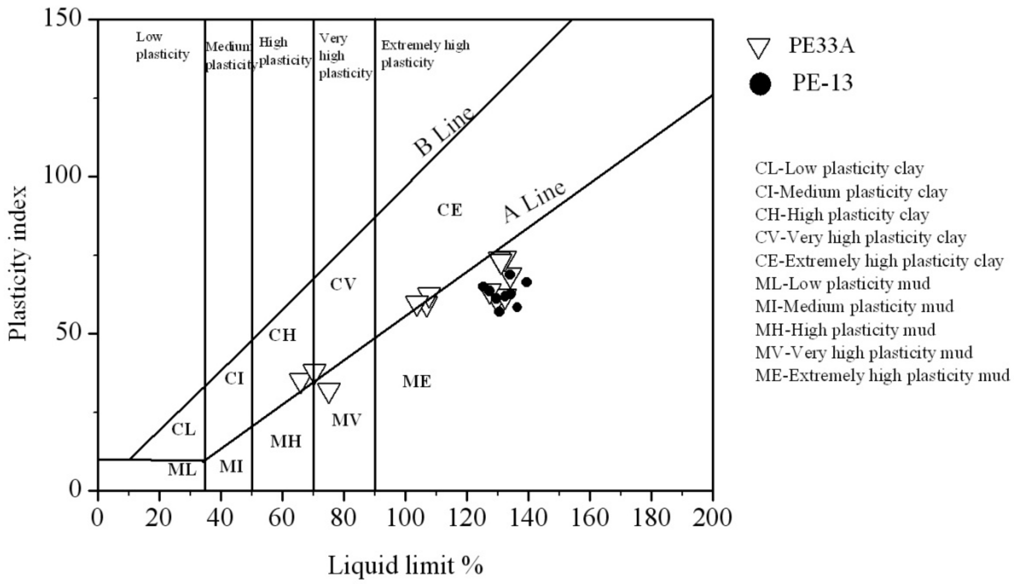


Figure 4.25: Distribution of samples of estuary cores on plasticity chart (after Casagrande, 1948)

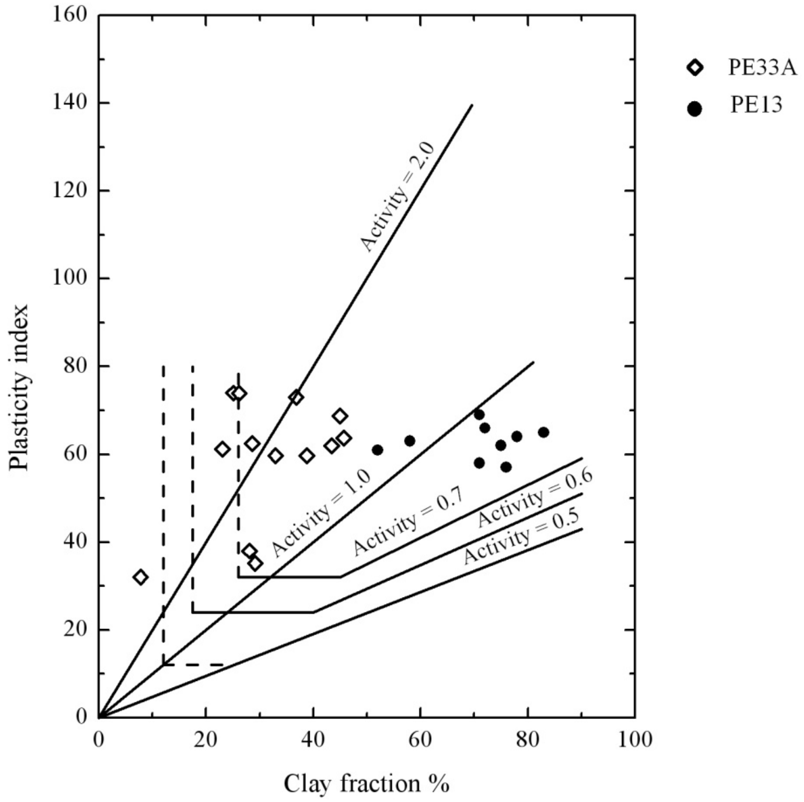


Figure 4.26: Activity chart showing the behaviour of sediment samples of estuary cores

Wet bulk density

The wet bulk density of core G3 sediments ranges from 0.49 to 0.61 g/cc. The wet bulk density of the sediments is less when compared to other core samples. The wet bulk density of the sediments of core G6 ranges from 0.55 to 0.83 g/cc, G7 ranges from 1.33 to 1.44 g/cc, G8 ranges from 1.22 to 1.40 g/cc, G9 ranges from 1.79 to 2.0 g/cc (Fig.4.15 to Fig. 4.20). G9 is a sandy core have the highest wet bulk density. Thus wet bulk density is inversely related to the water content and directly related to the sand percentage.

Atterberg limits:

(a) Liquid limit

The liquid limit of the grab samples ranges from 35 to 144 %. The grab sample S-1 shows the lowest liquid limit value which has a sand percentage of 82. Sample S-33 exhibits the highest liquid limit value of 144 % and the sand percentage of the sample is 0. The liquid limit values of the grab samples are showing strong negative correlation with sand percentage, positive correlation with silt and strong positive correlation with clay percentage (Table 4.1).

The liquid limit of core G-3 ranges from 118 to 133 %. In core G-3 up to 40 cm depth the water content of the sediment are higher than that of the corresponding liquid limits. Therefore up to 40 cm depth the sediments are in the liquid state and below 40 cm depth the water content values are lesser than the corresponding liquid limits. Thus sediments are in their plastic state below 40 cm depth. In the sediments of core G-4 the liquid limit values range from 119 to 150 %, in core G-6 the liquid limit ranges from 107 to 146 %, in core G-7 the values range from 94 to 105 %, in core GC-20 the value ranges from 64 to 125 % and these values are higher than the corresponding water contents (Fig. 4.14 to Fig. 4.18, Fig. 4.21 & Fig 4.22). In these core sediments except for the surface samples the liquid limit values are higher than the corresponding water content. Therefore the surface sediments are in the liquid state due to the direct contact with water. With depth the sediments change from liquid state to plastic state.

(b) Plastic limit

The plastic limit of the grab samples range from 26 to 73 % with an average value of 49 %. The highest plastic limit value is exhibited by S-33 sample. The plastic limit shows strong positive correlation with liquid limit, plasticity index, clay and organic carbon and negative correlation with sand as observed from the table 4.1.

The plastic limit of the core sediments has a similar trend with that of the corresponding liquid limit of the sediments. The plastic limit of core G-3 sediments ranges from 51 to 65 %, G4 ranges from 52 to 65 %, G-6 ranges from 47 to 68 %, core G-7 ranges from 40 to 50 %, core GC-20 ranges from 29 to 48 %, GC-16 ranges from 41 to 64 % (Fig. 4.14 to Fig 4.18, Fig 4.21 & Fig. 4.22). So in all the cores the plastic limit values are lower than the corresponding water content.

(c) Plasticity index

Plasticity index of grab samples were plotted against liquid limit values which is the plasticity chart as shown in the figure 4.27. The samples are lying above and below the A-line. As shown in the figure the samples are medium to extremely high plasticity clay and extremely high plasticity mud. The activity chart is shown in figure 4.28. The activity of the samples ranges from 0.98 to 1.76 with an average value of 1.34. According to Skemptoris (1953) classification the sediments are normal to active clays.

The plasticity chart of the nearshore core sediments is shown in the figure 4.28. All the core sediments are lying on the A-line or above or below near to the A-line. The plasticity index of the core sediments are having values greater than 40% except in core GC-20. So, all the core sediments are showing extremely high plasticity except for core GC-20 sediments. According to the Cassagrande's classification of clay the sediments are extremely plasticity clay or extremely plasticity mud. In core GC20 the plasticity index ranges from 34 to 78%. Thus the sediments are high to extremely high plasticity clay. The

Skemptoris classification of the core sediments is shown in the activity chart (Fig. 4.30). The sediments are active clays as shown in the figure.

(d) Liquidity index

All the sediments are having liquidity index (I_L) value in the range of 0.50 – 1.00 which means that the samples are cohesive soils or they are normally consolidated clays (after Means and Parcher 1963).

Table 4.1: Correlation matrix of textural and geotechnical properties of nearshore grab samples (n = 15)

	LL (%)	PL (%)	PI (%)	Activity	(WP/ WL)	SS (kPa)	OC (%)	S (%)	Z (%)	C (%)
LL (%)	1.00									
PL (%)	**0.95**	1.00								
PI (%)	**0.97**	**0.84**	1.00							
Activity	0.03	-0.15	0.18	1.00						
WP/WL	0.16	0.29	0.38	0.08	1.00					
SS (kPa)	0.16	0.29	0.38	0.08	**1.00**	1.00				
OC (%)	**0.85**	**0.65**	**0.78**	0.09	0.13	0.13	1.00			
S (%)	**-0.92**	**-0.82**	**-0.87**	-0.04	0.14	0.14	**-0.80**	1.00		
Z (%)	**0.58**	0.21	0.39	**0.55**	-0.28	-0.28	**0.50**	**-0.83**	1.00	
C (%)	**0.93**	**0.90**	**0.83**	-0.36	0.00	0.00	**0.85**	**-0.90**	0.49	1.00

LL-Liquid limit (%), PL-Plastic limit (%), PI-Plasticity Index (%), wP/wL-Plastic limit/Liquid limit, SS-Shear strength (kPa), OC-Organic carbon (%), S-Sand (%), Z-Silt (%), C-Clay(%)

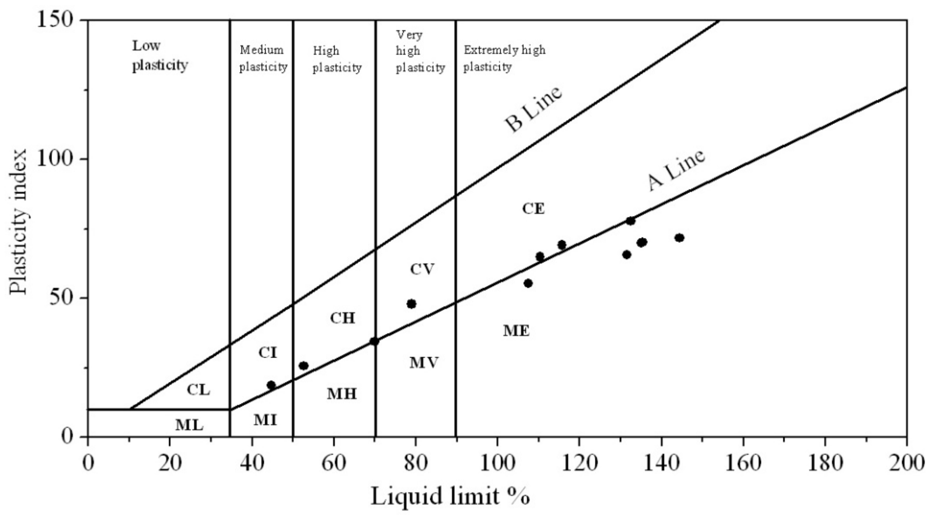


Figure 4.27: Distribution of nearshore grab samples on plasticity chart (after Casagrande, 1948)

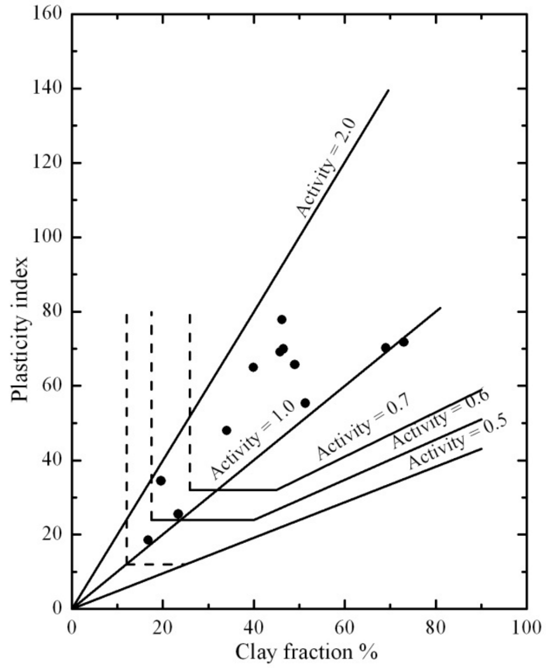


Figure 4.28: Activity chart showing the behaviour of nearshore grab samples

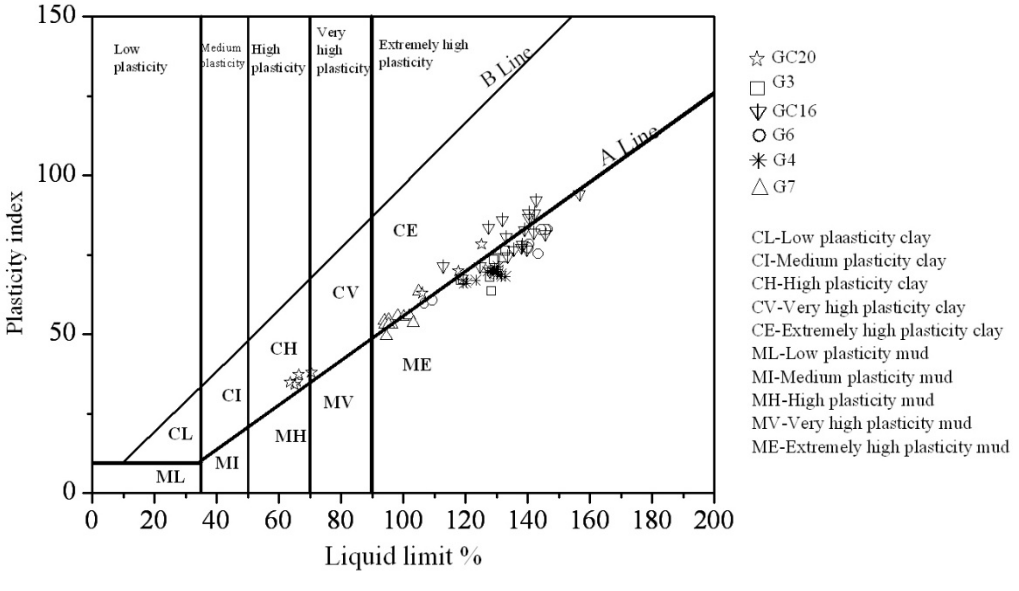


Figure 4.29: Distribution of samples of nearshore cores on plasticity chart (after Casagrande, 1948)

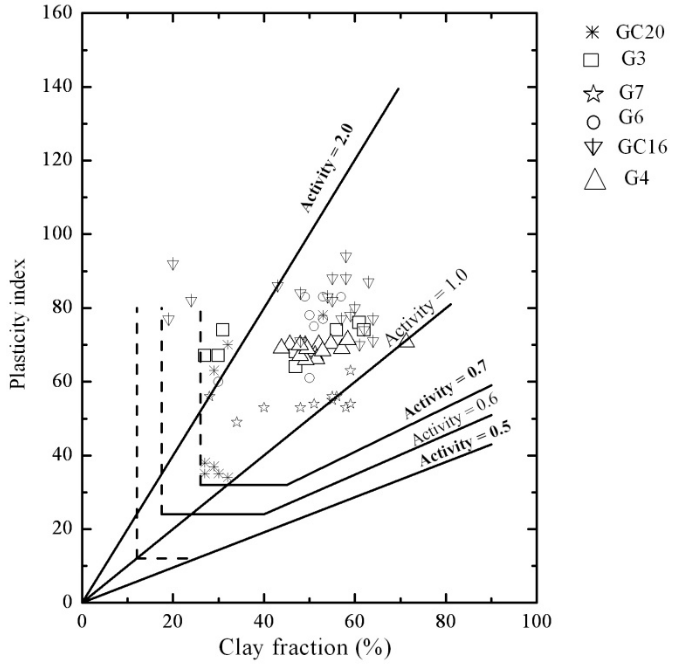
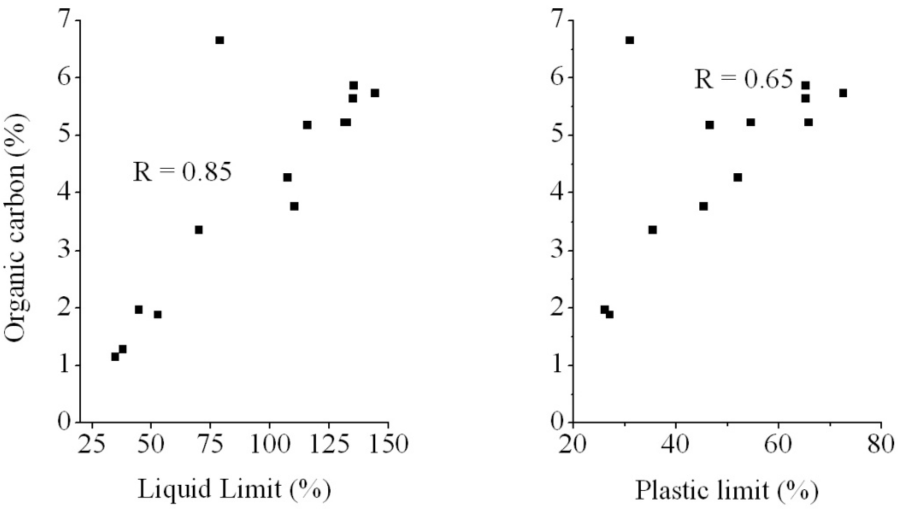


Figure 4.30: Activity chart showing the behaviour of samples of nearshore cores



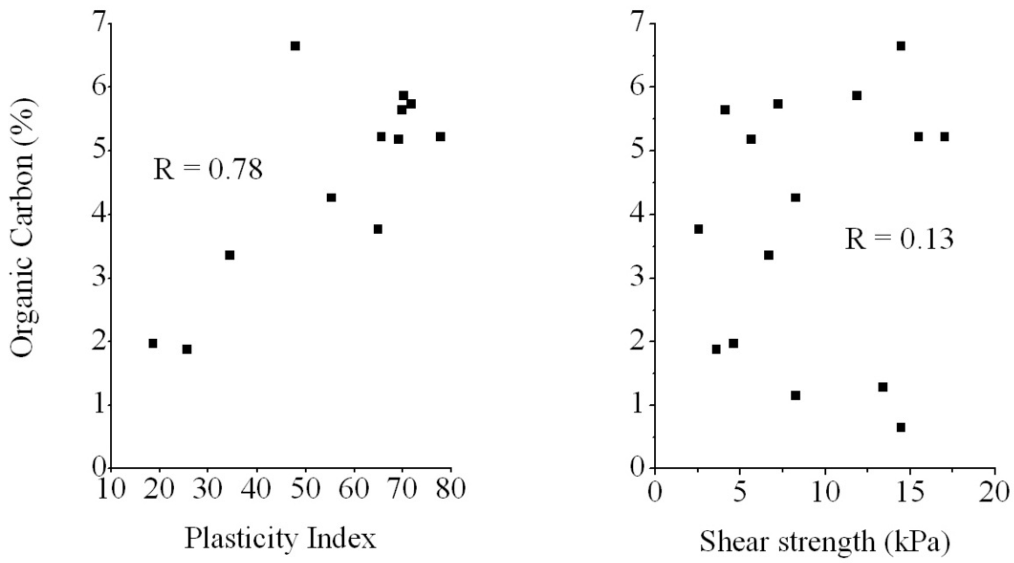


Figure 4.31: Interrelationship between organic carbon vs liquid limit, plastic limit, plasticity index and shear strength of nearshore grab samples

Vane shear

The shear strength of the nearshore grab samples ranges from 2.58 to 17.03 kPa with an average value of 9.19 kPa. The surface samples are highly saturated with water and has lesser degree of consolidation. This can be attributed to the low shear strength values for the surface sediments. The interrelationship between shear strength and organic carbon is shown in the figure 4.31.

The shear strength values of core G3 sediments ranges from 0.52 to 4.13 kPa, core G6 ranges from 2.07 to 6.2 kPa, core G7 ranges from 3.61 to 6.2 kPa, core GC20 ranges from 0.52 to 7.23 kPa, core GC16 ranges from 0.98 to 7.74 kPa (Fig 4.15 to Fig. 4.18, Fig 4.21 & Fig 4.22). The surface sediments are showing low shear strength values and the value increases with depth due to the consolidation.

4.5. Clay mineralogy

Clay respond to their chemical and thermal environment and their properties and species change accordingly (Velde, 1992), a sound knowledge of nature and distribution of clay minerals is very essential to understand the process of weathering, provenance studies, environmental analysis and paleoclimatic studies (Elliot et. al. 1996; Srivastava et. al. 1998, etc.). Illite, chlorite, kaolinite, and smectite have been analyzed to understand the Paleo-environmental conditions of the Vembanad lagoon and its surroundings.

4.5.1. Clay mineralogy of lagoonal environment

Clay mineral abundances in the lagoonal environment show from Smectite > Kaolinite > Illite > Chlorite. Figure 4.35 represents the cumulative clay mineral percentages. From the figure it is observed that PE-5 core has smectite in higher abundance. Illite and chlorite percentage are relatively very lower than smectite and kaolinite. Similarly, smectite is abundant in core PE-17 also, except at the interval 12-14 cm. At this depth, kaolinite and illite are abundant and smectite is 20 %. In PE-33A core, the relative concentration of illite is higher than other clay minerals. The cumulative concentration of illite ranges up to 60% whereas smectite and kaolinite are almost in equal proportion. Chlorite is found nil in the core except at 25-30cm, 100-105cm depth in very small proportion with respect to the other clay minerals in the core.

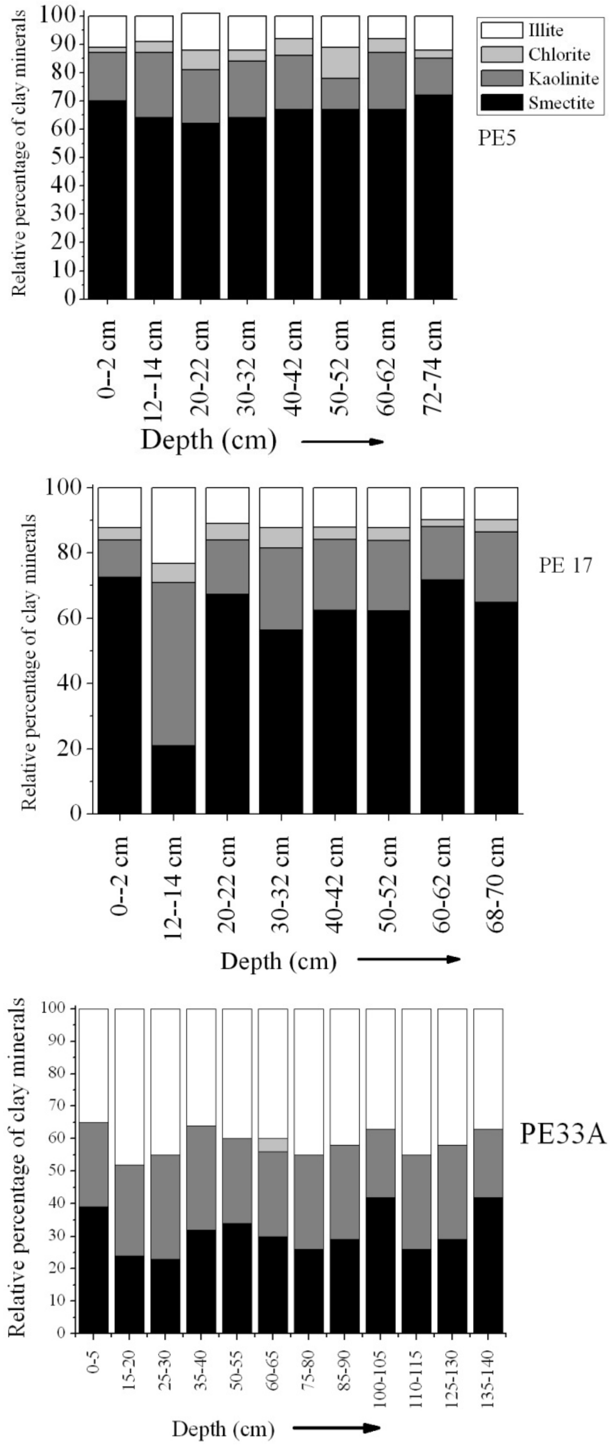


Figure 4.32: Relative percentage of clay minerals inVembanad lagoon core sediments

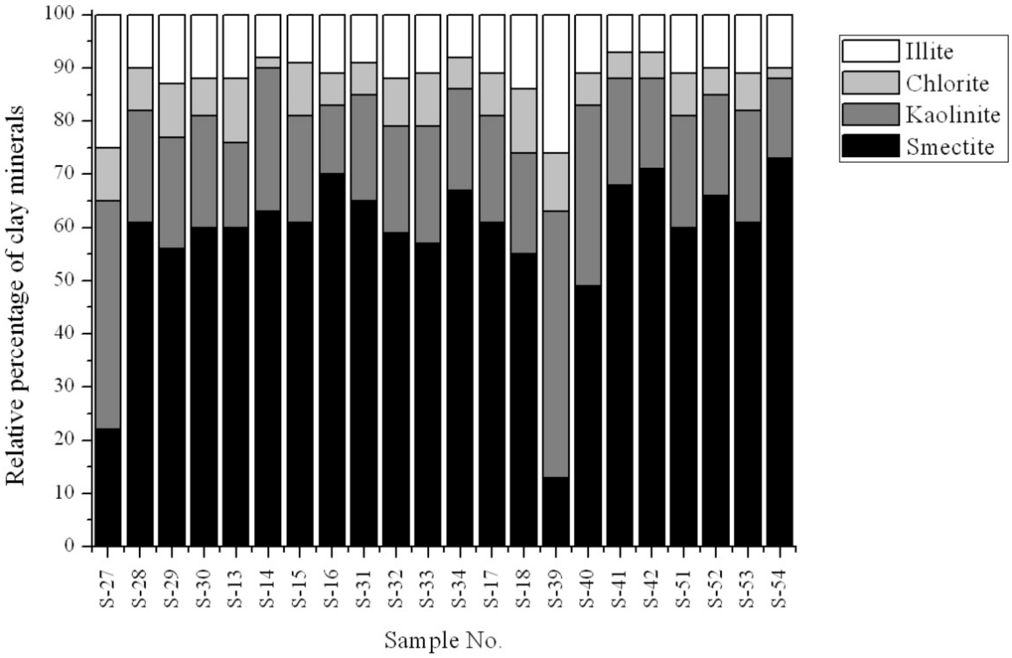


Figure 4.33: Relative percentage of clay minerals in nearshore grab samples

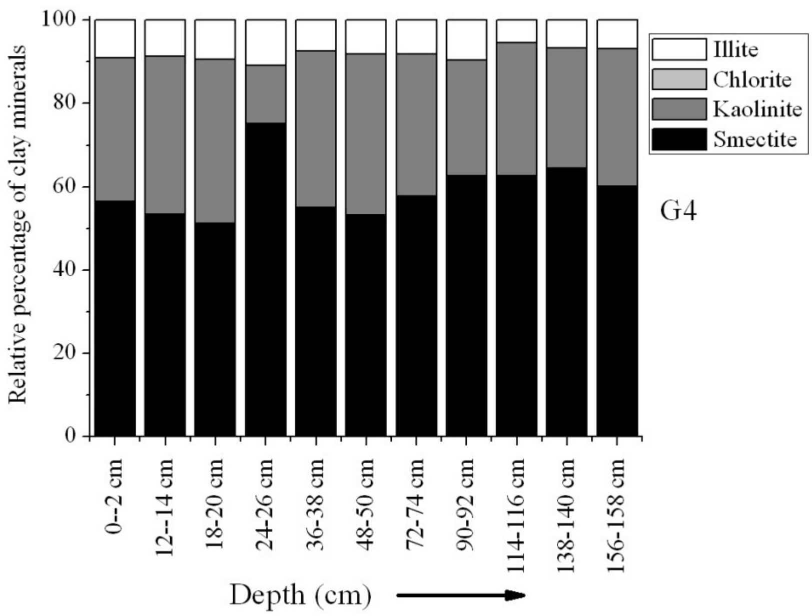


Figure 4.34a: Relative percentage of clay minerals in nearshore core sediments

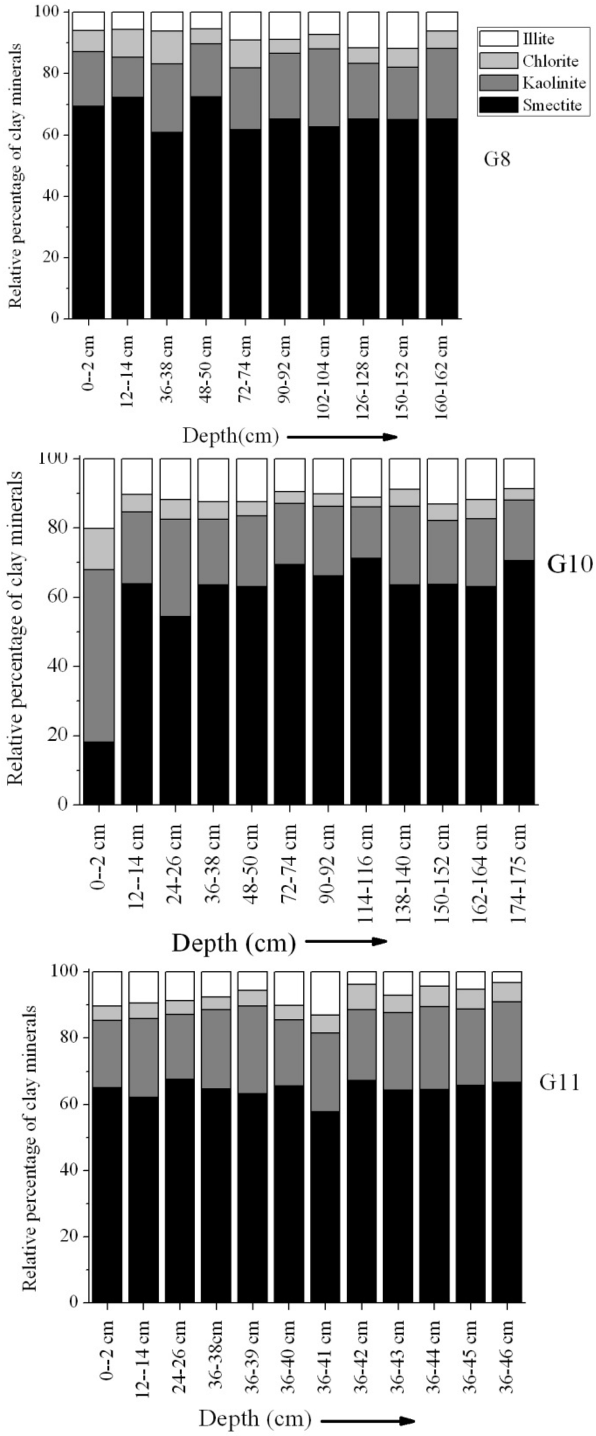


Figure 4.34b: Relative percentage of clay minerals in nearshore core sediments

Comparatively high proportions of kaolinite and illite near the river mouth stations in the Central Vembanad estuary might be due to the physical sorting of clay minerals depending on size. Whitehouse et al. (1960) explained that some of the changes, which the river-borne clay minerals undergo, based on differential settling velocities of clay minerals in water of increasing salinity are in the following order: illite, chlorite, kaolinite and smectite. Illite and kaolinite flocculate at lower salinities than smectite and get deposited near the river mouth stations whereas smectite will be deposited further down the estuary. This kind of physical sorting of clay minerals is supported by investigators like Laughman and Craig (1962), Gibbs (1977a). Thus core PE-33A sediments recovered near the mouth region of Chitrapuzha River is having higher illite percentage. Core PE-5 sediments recovered near the outlet of the Vembanad estuary at Munambam are having higher smectite percentage which can be attributed to the higher saline environment. Core PE-17 sediments recovered from the central part of the lagoon are also having higher smectite percentage because of the saline environment which can be due to the tidal influence through the Cochin outlet of Vembanad lagoon.

The decrease in the content of kaolinite and increase in the content of smectite towards the high saline zones of the Vembanad estuary could be explained in the light of the above attributes. Further, Grim et al. (1949); Griffin and Ingram (1955) and Grim (1968) have stated that kaolinite is unstable in alkaline waters and therefore it would alter either to illite or chlorite in estuarine and marine environments. This supports the role of physical sorting and size segregation processes in the observed clay mineralogical diversities of the study area. Hence, it is evident that the above clay minerals are source controlled and the lateral variations in the lagoonal regions are due to physical sorting. Among the estuary cores PE-5 and PE-17 are having high smectite percentage followed by kaolinite, illite and chlorite. Core PE-33A is having high illite percentage followed by smectite and kaolinite.

4.5.2. Clay mineralogy of nearshore environment

Relative percentages of clay minerals in nearshore grab samples show relatively high concentration of kaolinite and illite in S-27 and S-39 whereas all other samples have higher percentage of smectite (Fig. 4.33). Kaolinite has the second highest concentration and illite has the third highest percentage of concentration in the samples. Little varying percentage of chlorite was observed in all the samples. Smectite and kaolinite are the dominating clay minerals in G4 core with small concentrations of illite. Chlorite was found nil in this core. Similarly, G8 core also show abundant smectite and varying concentration of kaolinite, illite and chlorite. Abundant kaolinite was observed in the surface sediments of G10 core but smectite was again dominant in the core below 2 cm with similar variations like G8 core. G11 core also show an abundance of smectite after which kaolinite percentage is relatively higher than remaining clay minerals. Chlorite has least concentration in this core (Fig 4.34a & Fig. 4.34b).

Most fine-grained fluvially derived marine sediments are deposited in the nearshore and continental margins and only a small amount is transported to the deep sea (Chester, 1990). The fine-grained sediments are composed mainly of clay minerals and amorphous material. The composition and relative abundance of the clay minerals are controlled by their source rocks and weathering conditions. Their distribution on the continental shelf and slope is controlled by depositional processes, especially the current circulation patterns, and the settling of clay minerals in response to the salinity and energy conditions. Knowledge of these processes can also help to predict the transport pathways of pollutants (toxic metals and organic materials) which are preferentially concentrated in these fine-grained sediments.

In the present study nearshore cores are showing high smectite percentage followed by kaolinite, illite and chlorite. Rao and Rao (1995) has reported smectite and kaolinite-rich assemblage with minor illite, chlorite and gibbsite derived from the Gneissic Province occurs both on the shelf and slope between Goa and Cochin which support our study.

4.6 Summary

Lagoons are present on many coasts of the world in a variety of environment settings and are most common in microtidal environments (Hayes, 1975). Estuarine or lagoon and nearshores receive sediments from a number of sources including the watershed, the continental shelf, the atmosphere, and erosion of estuarine margins, bottom, and the metabolism within the lagoon due to biological and chemical activities (Abdullah et al., 2012). The dominance of one sediment source depends on its magnitude relative to all other source, processes, and the dynamics of erosion, transportation, and deposition. The transportation and deposition of sediments in estuarine environments are complex due to dynamic processes that circulate and mix estuarine waters, the flocculation processes which increases the settling velocity of the sediment particles and properties of the particles themselves (Abdullah et al, 2012). There have been many workers involved worldwide to reconstruct paleoenvironment from lagoonal and nearshore sedimentary sequences based on geotechnical and textural characteristics (Hayes, 1980; Roy et al., 1980; Calcagno and Ashley, 1984; Woodroffe et al., 1989; Alexander et al., 1991; Knebel et al., 1991; Nichols et al., 1991; Manoj et al., 1999; Chang et al., 2001; Anthony et al., 2002; Chen et al., 1998, Priju et al., 2004). Prior studies on surficial sediments of Vembanad lagoon between Cochin and Allepey was done by Murty and Veerayya (1972). Further, Mallik and Suchindan (1984) studied the Sedimentological aspects of central and southern part of the Lagoon. In this work, the textural and geotechnical analysis was divided in two regions as lagoonal environment and nearshore environment to understand the discriminating consequences.

4.6.1 Textural Studies

In this section, the general distribution pattern of sediment texture across the Vembanad lagoon has been discussed based on the variation of sand-silt-clay ratios of the sediment cores. This gives the idea of change in the textural pattern of the sediments which can be concluded as the change in the environment of deposition. The sedimentary lithofacies present in lagoonal beds are controlled by the hydrodynamic processes as well as the nature of the availability of the sediments. Vertical aeration of the sediments is mediated by the rate and nature of the sediments supply. Sediment supply could be either in-situ or transported from sea or land. Quiet water deposition of silt and clay occurs in lagoons in a manner analogous to pro-delta environments, where sediments are carried by rivers. Deposition from suspension of fluvial sediment is often enhanced by flocculation in the more saline lagoonal waters (Hobday, 1976). The study on the net sediment budget of the Laguna Madre, Texas by Morton et al (2000) suggests that tidal currents, wind, storm washover, upland runoff, interior shore erosion, and authigenic mineral production control the sedimentation. He also reports the aerial restriction of specific transportation processes. The studies of the sedimentological aspects of lagoonal and nearshore sediments of west coast of India have been focused on grain size variations of the sediments (Veerayya et al., 1973; Veerayya and Varadacheri, 1975; Pandarinath, 1991; Nasnolkar et al., 1996; Nair & Ramachandaran, 2002.

The textural characteristics were studies of Goa coast by Veerayya et al. (1973) to understand the depositional environments

As indicated in the textural variations of the lagoonal sediments, the system represents a high energy environment. This is characterized by sand-muddy sand-sandy mud facies. Core PE-33A represents a change in the energy conditions from low to high when moved upwards in the core. The grab samples also show a similar high energy environment by sandy mud textures. However, it shows mud textures in S-2, S-3, S-31, and S-32 grab samples. Nearshore samples broadly show a low energy environment of deposition represented by muddy textures. However, G-6 core has sandy mud texture in the surface sediments. Core G-9 shows a change in textural pattern downcore from muddy sand to clayey sand. Similarly, Core GC-20 also shows a downcore textural variation as sandy mud-mud-sandy mud.

Knowledge of the various characteristics of sediment permits the assessment of its state of evolution, it maturity and also a better understanding of its history. Chamley (1990) states that mature sediments with narrow granulometric range may be used as evidence for active and prolonged hydrodynamic processes such as littoral or desert dunes, beaches, and other shallow-marine exposed environments.

4.6.2 Organic Carbon

Hydrodynamic sorting of sediment according to size or density reworks sediments and associated organic matter across continental shelves (e.g. Goñi et al., 1997; Wakeham et al., 2009). The shelf zones, particularly the coastal systems, behave as carbon traps (Romankeviche, 1984; Wollast, 1991) where part of the organic matter in the overlying waters accumulates in the underlying sediments. The source of this organic matter may be dominated by autochthonous material from net primary production and allochthonous material supplied by rivers, or both together. Natural eutrophication may sometimes occur due to nutrient loadings and from discharges of industrial plants and sewage treatment works and from other human influenced ‘diffuse sources’ such as run-off from agricultural catchments that may increase the water eutrophication effects. The enrichment of water by nutrients causes an accelerated growth of algae to provide the causes and symptoms of eutrophication (de Jonge et al., 2002). A number of biological, physical and chemical factors such as primary productivity, dissolved oxygen in bottom waters, texture of sediments and rate of sedimentation govern the preservation of organic matter in marine environments (Pedersen and Calvert, 1990). Dissolved humic substances, which are abundant in the river water flocculate as they meet the saline estuarine waters. Since many rivers drain into the Vembanad lagoon/estuary, the major component of the total organic carbon (TOC) may be the humic substances (Resmi 2004).

The nearshore processes such as wind, waves and currents play an important role in removing the fine sediments near the shore area which contains high organic matter. In general, it can be said that in an estuary the organic matter is formed either by photosynthesis or transported as particulate or dissolved matter into the estuary by the rivers. Nearshore zone also receives organic matter dominantly from terrestrial source (Pandarinath and Narayana, 1998).

4.6.3 Geotechnical studies

The geotechnical properties of the sediments – high moisture content, high organic carbon, high plasticity index, and low bulk density-are generally associated with a low critical shear stress for erosion (Smerdon and Beasley, 1961; Vanoni, 1975; Thorn and Parsons, 1980) and increased erosion rate (Mehta et al., 1982). An inverse relation of wet bulk density and moisture content exist in most of the core samples. Moisture also shows variability with high organic carbon % downward in different cores indicating the varying erosional rate temporally. Organic carbon % shows increasing trend with moisture content and clay percentage. Clay has the water absorbing nature and acts as a sink of organic carbon. Similarly this can also be taken as an indication of hydrogel formation amongst organic matter (Louda et al., 2004). Organic carbon also shows an inverse relation with wet bulk density in the cores supporting the high Atterberg limits (cf. Keller, 1982). Sand also shows variability in its trend with silt and clay and hence, variable correlation with liquid and plastic limit. Lower concentration of silt and clays and the higher concentration of sand in the estuary cores indicate the lower cohesion of the sediments showing that the sediments are more susceptible to resuspension. Liquid limit, plastic limit and plasticity index of the core sediments are showing positive trend with silt and clay percentages. The surface sediments of the cores are having higher moisture content than the corresponding liquid limit. Therefore the surface sediments are in the liquid state. As depth increases the moisture content of the sediments shows lower values than the corresponding liquid limits. Thus the sediments are in the plastic state with increasing depth. Based on the Cassagrande's (1948) plasticity chart the sediments in the nearshore environment are lying above and below the A-line and are high plasticity to extremely high plasticity clay and mud. The activity chart of the cores (clay % versus plasticity index) showing Skemptoris (1953) classification of clays as 'active', 'normal', and 'inactive clays', represents that most of the samples are active in nature. The surface sediments of the cores are showing very low shear strength values and exhibit an increasing trend with depth.

According to Means and Parcher (1963) classification of clayey soil based on the degree of consolidation it can be concluded that most of the core sediments at depth are normally consolidated clays.

4.6.4 Clay Mineralogy

The nature and composition of clay minerals provide valuable informations regarding the type and intensity of weathering on land and also the degree of hydrolysis, at source rock regions (Chamley, 1989, Thamban et al., 2002). However, size sorting during transportation (Gibbs, 1977), organic interaction of clays (Degens and Ittekot, 1984, Thamban et al, 2002), flocculation at lower salinities (Grim, 1968), redistribution and reworking of clay minerals should be understood before the diagenetic studies for the provenance. The major clay mineral groups present in the borehole sediments are: illite, chlorite, kaolinite, and smectite. The relative concentration of clay minerals follows the following trend in almost all the cores from both the environments

Smectite>Kaolinite>Illite>Chlorite

Smectite is derived from chemical weathering of parent alumino silicate and ferromagnesian silicate under warm and humid conditions, and also from chemical weathering of basaltic rocks (Shinu, 2008). Smectite is considered as settling preferentially in areas of decreased current and associated grain sorting. The ability of smectite minerals to settle preferentially allows identification of current activity (Chamley, 1989). Hence, abundant smectite in the sediments indicate the low energy environment in the lagoonal system. Further, Whitehouse et al. (1960) explained that some of the changes, which the river-borne clay minerals undergo, based on differential settling velocities of clay minerals in water of increasing salinity are in the following order: illite, chlorite, kaolinite and smectite. Illite and kaolinite flocculate at lower salinities than smectite and get deposited near the river mouth stations whereas smectite will be deposited further down the estuaries. This can be attributed to the high illite percentage in the core near to the river mouth. Thus the clay minerals present in the Vembanad lagoon are transported by the rivers from the weathering of Precambrian gneisses and laterites and are differentially flocculated based on the salinity.

Weathering of Deccan trap basalts under a semi-arid climate is the source of Smectite and is carried out mainly by river systems (Subramanian, 1980; Naidu et al., 1985; Rao and Rao, 1995; Kessarkar et al., 2003). Kaolinite is readily found in soils of inter tropical land masses characterized by a warm, humid climate, and therefore displays a strong climatic dependence controlled by the intensity of continental hydrolysis (Chamley, 1989). Here, Precambrian gneissic rocks and laterites of Western Ghats are the main source for the kaolinite in the sediments. The detrital chlorite mainly results from chemical weathering of plutonic and metamorphic rocks. Chlorite is reported to be a product of physical weathering under an arid and cold climate (Biscaye, 1965; Weaver, 1989). However, the influence of the Indus river system to the study is not clear; the source for the illite could also be the weathering of Deccan trap basalts in-arid climate conditions. Hence, the source regions for the clay mineral deposition in the central Kerala coast are Deccan trap basalts for illite, chlorite and smectite; gneissic rocks and laterites of the adjacent terrestrial areas for Kaolinite.

****

Chapter 5

TRACE ELEMENT GEOCHEMISTRY

Contents

5.1 Introduction

Recent data on trace element behaviour in both the natural environment and that modified by human activities are reviewed by several authors (Babich, 1978; Bain, 1976). While geological and biological alterations of the Earth's surface have been very slow, changes introduced and stimulated by man have accumulated extremely quickly in recent years. Environmental pollution, especially by chemicals, is one of the most effective factors in the destruction of biosphere component. Among all chemical components, trace elements are believed to be a specific ecological, biological, and geological significance (Kabata-Pendias 2001). It is now well established that human activities have greatly altered the geochemical cycles of metals, resulting in widespread environmental contamination (Nrigau and Pacyna, 1998). Input from the river consist a dominant transporting agent of metals from the continents to the sea (Fang et al., 2009; Ip et al., 2007; Miller et al., 2003; Radakovitch et al., 2008). These inputs are augmented by anthropogenic (e.g. pollution) disturbances in direct or indirect way (e.g. alterations of landscape resulting in enhanced natural weathering).

Sediments have a major role in the overall fluxes of trace elements in coastal systems and can get recycled several times through the sediment-water interface before being permanently stored in sediments or released to the overlying waters (Sakellari et. al., 2011). Sediments may not only act as sinks but also as sources of contaminants in aquatic system. Concentration of trace elements in sediments is known to be a useful tool in biogeochemical exploration and environmental research (Anderson, 1973; Allen, 1980). Trace metals are introduced in the sediments by various pathways.

Copper, lead, and zinc are trace metals, which usually have an anthropogenic origin (Alvarez-Iglesias et al., 2003). However, the relative influence of natural and anthropogenic sources on the geochemistry of lagoonal sediments is not very clear. Therefore, for a better assessment of the pollution levels in the marine coastal environment, it is important to distinguish between natural and metal enrichments by human beings. (Angelidis & Aloupi, 2000). Down core variation in the concentration of trace elements, analyzed by ICP-MS at NGRI, Hyderabad for the marine cores G8, G10, and G11 recovered from the coast of central Kerala, are shown in figure 5.1. This chapter deals with trace element analysis, Total Organic carbon %, Enrichment Factor, Pollution Load Index (PLI), and Geo-accumulation Index (I_{geo}) to distinguish the natural and anthropogenic contamination, in the coastal region of central Kerala, India and also to understand the source for the contamination.

5.2 Trace Element concentrations in nearshore cores

Cobalt (Co)

In the Earth's crust, Cobalt (Co) has a high concentration in ultramafic rocks (100 to 220 ppm) when compared to its content in acid rocks (1 to 15 ppm). The Co abundance in sedimentary rocks ranges from 0.1 to 20 ppm and seems to be associated with clay minerals or organic matter. The crust of the Earth as a whole contains, on average, 25 ppm Cobalt. Co is associated as hidden in various Fe minerals. In geochemical cycles, Co closely resembles Fe and Mn. However, its fate in weathering processes and its distribution in sediments and in soil profile seems to be strongly determined by Mn oxide phase formation. During weathering, Co is relatively mobile in oxidizing acid environments, but due to a high sorption by Fe and Mn oxides, as well as by clay minerals, this metal does not migrate in a soluble phase.

Cobalt concentration in the core sediments ranges from 10 ppm to 30 ppm with an average of 12 ppm, 13 ppm, and 16 ppm of G8, G10, and G11 cores respectively. In G8 core, there has been a continuous increase in cobalt concentration by one ppm from 10 ppm, 11 ppm, 12 ppm, and 13 ppm at 0-2 cm, 6-8 cm, 18-20 cm, and 30-32 cm respectively. Concentration remains 13 ppm up to 42-44 cm interval of depth. Cobalt again decreases by 1 ppm (i.e. 12 ppm) at 48-50 cm and 102-104 cm interval of depth. Further it decreases to 11 ppm at 138-140 cm and increases to 13 ppm at 168-170 cm interval of depth.

Similarly, in core G11 sediments Co concentration increases from 11 ppm to 12 ppm at a depth of 6-8 cm. It further increases to 13 ppm at the depth interval of 18-20 cm, and the concentration remains constant up to a depth of 48-50 cm. There has been an increase to 14 ppm at the depth interval of 102-104 cm and then again the concentration remains 13 ppm at 120-122, 148-150, and 172-174 cm depth intervals.

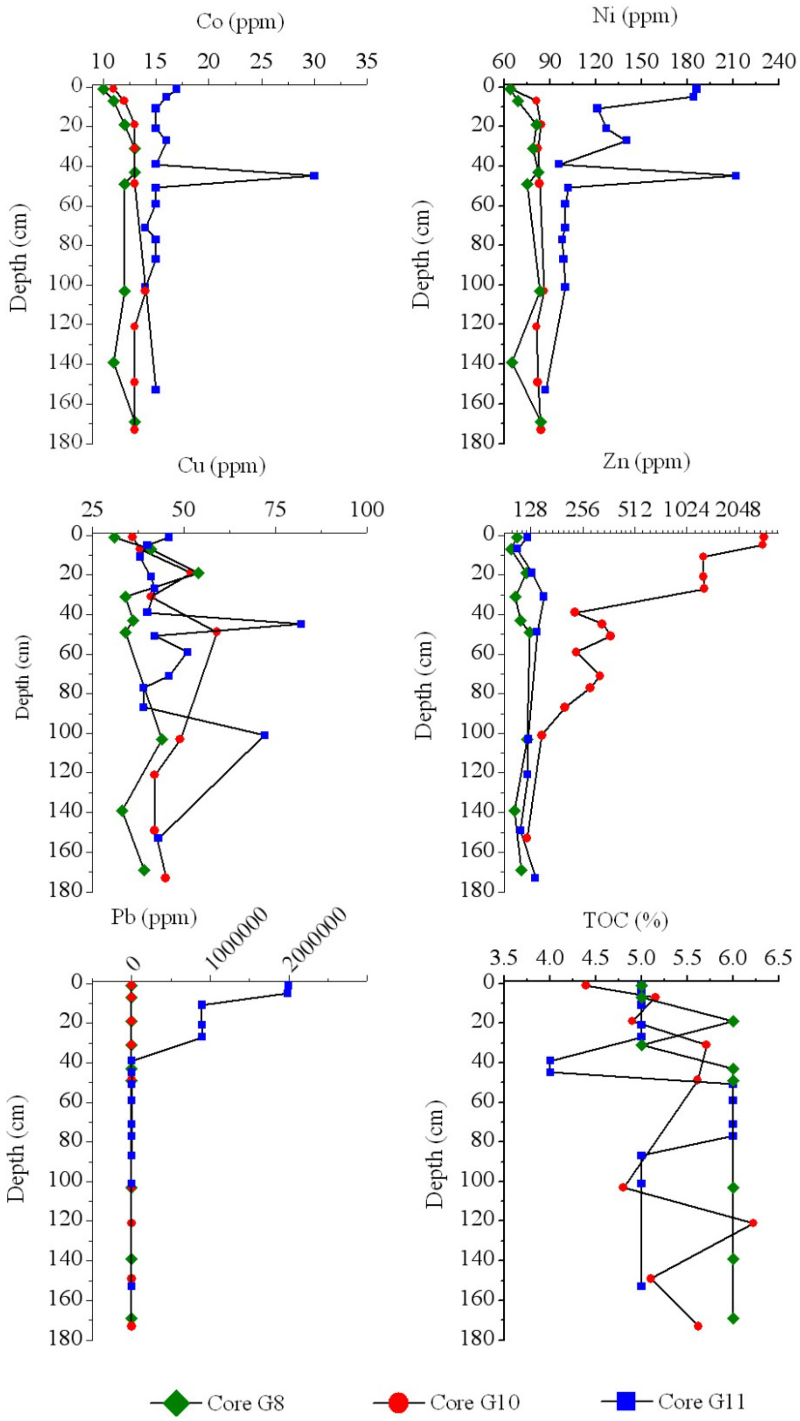


Figure 5.1: Down core variation of trace elements and total organic carbon in nearshore core sediments.

In G11 core, the concentration of Co is relatively higher than G8 and G10 cores. The concentrations decrease from 17 ppm to 16 ppm at 0-2 cm and 4-6 cm respectively. Further, there is a down core decrease from 16 ppm to 15 ppm at 10-12 cm, 20-22 cm, 38-40 cm, 50-52 cm, 58-60 cm, 76-78 cm, 86-88 cm, and 152-154 cm intervals of depth. The concentration of cobalt remains 16 ppm at 26-28 ppm and 14 ppm at the depth intervals of 70-72 cm and 100-102 cm. An abrupt change in concentration from 15 ppm to 30 ppm has been observed at the depth interval of 44-46 cm.

Nickel (Ni)

There is a general similarity between the distribution of Ni, Co, and Fe in the Earth's crust. Thus, Ni contents are highest in ultramafic rocks (1400 to 2000 ppm), and its concentrations decrease with increasing acidity of rocks down to 5 to 15 ppm in granites. Sedimentary rocks contain Ni in the range of 5 to 90 ppm, with the highest range being for argillaceous rocks and the lowest for sandstones. Geochemically, Ni is siderophilic and will join metallic Fe wherever such a phase occurs. Also the great affinity of Ni for Sulfur accounts for its frequent association with segregates of Sulfur bodies. In terrestrial rocks, Ni occurs primarily in sulfides (millerite, NiS), arsenides (niccolite, NiAs), and antimonides (breithauptile, NiSb), and most of it is in ferromagnesians, replacing Fe. Ni is also associated with carbonates, phosphates, and silicates. Ni is easily mobilized during weathering and then is coprecipitated mainly with Fe and Mn oxides. However, unlike Mn $2^{+}$and Fe 2^{+}, Ni 2^{+} is relatively stable in aqueous solutions and is capable of migration over a long distance. During weathering of Ni-rich rocks (mainly in tropical climates), the formation of garnierite, (Ni, Mg) $SiO_3 \cdot nH2O$, which is a poorly defined mixture of clay minerals, is observed.

The average concentrations of nickel for G8, G10, and G11 cores are 76 ppm, 81 ppm, and 125 ppm respectively. In G8 core, Nickel increases from 64 ppm to 69 ppm and from 69 ppm to 81 ppm at the depth intervals of 6-8 cm and 18-20 cm respectively. The concentration observed at the intervals of depth of 30-32 cm, 42-44 cm, 48-50 cm, 102-104 cm, 138-140 cm, and 168-170 cm are 79 ppm, 82 ppm, 75 ppm, 83 ppm, 65 ppm, and 84 ppm respectively.

The concentrations of nickel in G10 core are 65 ppm, 81 ppm, 84 ppm, 82 ppm, 83 ppm, 86 ppm, 81 ppm, 82 ppm, and 84 ppm at the depth intervals of 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 48-50 cm, 102-104 cm, 120-122 cm, 148-150 cm, and 172-174 cm respectively.

G11 core shows a relatively higher concentration as 186 ppm, 184 ppm, 121 ppm, 127 ppm, 140 ppm, and 96 ppm at the depth intervals of 0-2 cm, 4-6 cm, 10-12 cm, 20-22 cm, 26-28 cm, and 38-40 cm respectively. There has been a sharp increase of 212 ppm in concentration at the depth interval of 44-46 cm. Further at the depth intervals of 50-52 cm, 58-60 cm, 70-72 cm, 76-78 cm, 86-88 cm, 100-102 cm, and 152-154 cm, the concentrations are 102 ppm, 100 ppm, 100 ppm, 98 ppm, 99 ppm, 100 ppm, and 87 ppm respectively.

Copper (Cu)

Copper (Cu) in the Earth's crust is most abundant in mafic and intermediate rocks and is absent in carbonate rocks. Cu forms several minerals of which the common primary minerals are simple and complex sulfides. These minerals are quite easily soluble in weathering processes and release Cu ions, especially in acid environments. Therefore, Cu is considered among the more mobile of the heavy metals in hypergenic processes. However, Cu is a very versatile trace cation and in soils or depositional material exhibits a great ability to chemically interact with mineral and organic components of soil. The Cu ions can also readily precipitate with various anions such as sulfide, carbonate, and hydroxide. Thus, Cu is a rather immobile element in soils and shows relatively little variation in total content in soil profile.

Although soluble, therefore mobile and available, forms of Cu in soils are of great importance in agronomic practice, total Cu content of soils gives basic information for geochemical studies. The Cu geometric mean of over 32,000 agricultural soil samples of Poland is only 6.5 ppm, which is due to the predomination of light sandy acid soils. However, the variations in the spatial distribution are remarkable, indicating both relations to parent materials and to the impact of industrial and/or agricultural pollution. The concentration of Cu in samples of surface soil from major agricultural production areas of the U.S. varies from 0.6 to 495 ppm (geometric mean, 18 ppm). The regularity in large scale Cu occurrence in soils indicates that two main factors, parent material and soil formation processes, govern the initial Cu status in soils. Also, the clay fraction contributes significantly to the Cu content of soils. As the RDI value indicates, this is a main soil parameter influencing Cu soil status. Other soil properties, such as Fe and Mn oxides, and base saturation, explain about 15 to 25% of all impact factors. The common characteristic of Cu distribution in soil profile is its accumulation in the top horizons. This phenomenon is an effect of various factors, but above all, Cu concentration in surface soils reflects the bioaccumulation of the metal and also recent anthropogenic sources of the element.

Copper ranges on an average of 39 ppm, 45 ppm, and 47 ppm in G8, G10, and G11 cores respectively. It's concentration in G8 core changes by 31 ppm, 41 ppm, 54 ppm, 34 ppm, 36 ppm, 34 ppm, 44 ppm, 33 ppm, and 39 ppm at the depth intervals of 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 42-44 cm, 48-50 cm, 102-104 cm, 138-140 cm, and 168-170 cm respectively.

Similarly, in G10 core, the concentrations change by 36 ppm, 38 ppm, 52 ppm, 41 ppm, 59 ppm, 49 ppm, 42 ppm, 42 ppm, and 45 ppm respectively at the depth intervals 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 48-50 cm, 102-104 cm, 120-122 cm, 148-150 cm, and 172-174 cm respectively.

In G11 core, the variation in the concentrations at the depth intervals 0-2 cm, 4-6 cm, 10-12 cm, 20-22 cm, 26-28 cm, 38-40 cm, 44-46 cm, 50-52 cm, 58-60 cm, 70-72 cm, 76-78 cm, 86-88 cm, 100-102 cm, 152-154 cm are 46 ppm, 40 ppm, 38 ppm, 41 ppm, 42 ppm, 40 ppm, 82 ppm, 42 ppm, 51 ppm, 46 ppm, 39 ppm, 39 ppm, 72 ppm, and 43 ppm respectively. A rapid increase in the concentration of the copper is observed at the depth interval of 44-46 cm.

Zinc (Zn)

Zn seems to be distributed rather uniformly in the magmatic rocks, and only slight increase in mafic rocks (80 to 120 ppm) and slight decrease in acid rocks (40 to 60 ppm). The Zn concentration in argillaceous sediments and shales is enhanced, ranging from 80 to 120 ppm; while in sandstones and carboniferous rocks, concentrations of this metal range from 10 to 30 ppm. The Zn occurs chiefly as single sulfides (ZnS), but is also known to substitute for $Mg2^{+}$ in silicates. The solubilization of Zn minerals during weathering produces mobile Zn^{2+}, especially in acid, oxidizing environments. Zn is, however, also easily absorbed by mineral and organic components and thus, in most soil types, its accumulation in the surface horizons is observed. Mean total Zn contents in surface soils of different countries and of the U.S. range from 17 to 125 ppm. Thus, these values may be considered as background Zn contents. The highest means were reported for some alluvial soils, solonchaks, and rendzimas, while the lowest values were for light mineral and light organic soils. Grand mean Zn for worldwide soils may be calculated as 64 ppm. Soil organic matter is known to be capable of bonding Zn in stable forms; therefore, the Zn accumulation in organic soil horizons and in some peats is observed. However, stability constants of Zn-organic matter in soils are relatively low, but a high proportion of Zn is bound to organic matter in mineral soils.

Zinc concentration is relatively higher in the three cores with an average value of 112ppm, 127ppm, and841 ppm for G8, G10, and G11 respectively. Zinc shows a consistent higher concentration in the three cores with highest concentration in G11 core. The concentration of Zinc in G8 core ranges from 98ppm to 126 ppm. It ranges from 107 ppm to 152 ppm in G10 core and from 122 to 2861 ppm in G11 core. The concentration in G11 core changes abruptly from 232 ppm to 1295 ppm at the depths of interval 38-40 cm to 26-28 cm respectively.

The concentrations of Zinc in G8 core at the 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 42-44 cm, 48-50 cm, 102-104 cm, 138-140 cm, and 168-170 cm intervals of depth are 106 ppm, 98 ppm, 121 ppm, 104 ppm, 112 ppm, 126 ppm, 122 ppm, 122 ppm, 103 ppm, and 113 ppm respectively. Similarly the concentrations in G10 core are 123 ppm, 107 ppm, 130 ppm, 152 ppm, 139 ppm, 124 ppm, 123 ppm, 112 ppm, and 136 ppm at the depth intervals of 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 48-50 cm, 102-104 cm, 120-122 cm, 148-150 cm, and 172-174 cm respectively. The concentrations of zinc in G11 core are very high in the shallow depths and relatively lower in the downcore. The concentrations are 2861 ppm, 2814 ppm, 1276 ppm, 1279 ppm, 1295 ppm, 232 ppm, 332 ppm, 372 ppm, 235 ppm, 322 ppm, 284 ppm, 202 ppm, 148 ppm, and 122 ppm at the depth intervals of 0-2 cm, 4-6 cm, 10-12 cm, 20-22 cm, 26-28 cm, 38-40 cm, 44-46 cm, 50-52 cm, 58-60 cm, 70-72 cm, 76-78 cm, 86-88 cm, 100-102 cm, and 152-154 cm respectively.

Lead (Pb)

The terrestrial abundance of Pb indicates a tendency for Pb to concentrate in the acid series of magmatic rocks and argillaceous sediments in which the common Pb concentrations range from 10 to 40 ppm, while in ultramafic rocks and calcareous sediments its range is from 0.1 to 10 ppm. The average abundance of Pb in the Earth's crust is estimated at about 15 ppm. In the terrestrial environment, two kinds of Pb are known: primary and secondary. Primary Pb is of geogenic origin and was incorporated into minerals at the time of their formation, and secondary Pb is of a radiogenic origin from the decay of U and Th. The ratio of Pb of various origins is used for dating the host materials. Pb has highly chalcophilic properties and thus its primary form in the natural state is galena (PbS). Pb occurs mainly as Pb2, although its oxidation state 4^{+}, is also known, and it forms several other minerals which are quite insoluble in natural waters. During weathering, Pb sulfides slowly oxidize and have the ability to form carbonates and also to be incorporated in clay minerals, in Fe and Mn oxides, and in organic matter. The geochemical characteristics of $Pb2^{+}$ somewhat resemble the divalent alkaline-earth group of metals; thus, Pb has the ability to replace K, Ba, Sr, and even Ca, both in minerals and in sorption sites. The natural Pb content of soil is inherited from parent rocks. However, due to widespread Pb pollution, most soils are likely to be enriched in this metal, especially in the top horizon.

Lead concentration is moderately high among the trace metals in G8 and G10 cores whereas it is higher in G11 core with an average concentrations of 40 ppm, 42 ppm, and 459 ppm. Enrichment in lead concentration is observed at the surface levels of all the three cores.

In G8 core, the concentration varies downcore as 54 ppm, 24 ppm, 30 ppm, 18 ppm, 55 ppm, 52 ppm, 57 ppm, 44 ppm, and 27 ppm at the depth intervals 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 42-44 cm, 48-50 cm, 102-104 cm, 138-140 cm, and 168-170 cm respectively.

In G10 core, the concentration varies downcore as 47 ppm, 28 ppm, 26 ppm, 48 ppm, 53 ppm, 50 ppm, 38 ppm, 47 ppm, and 40 ppm respectively at the depth intervals of 0-2 cm, 6-8 cm, 18-20 cm, 30-32 cm, 48-50 cm, 102-104 cm, 120-122 cm, 148-150 cm, and 172-174 cms respectively. There has been a gradual increase at the middle depth in this core which has further varied in the same fashion like the surficial concentrations.

In G11 core, the lead concentrations are very high in the upper region as 994 ppm, 988 ppm, 895 ppm, 894 ppm, and 899 ppm respectively at the depth intervals of 0-2 cm, 4-6 cm, 10-12 cm, 20-22 cm, and 26-28 cm respectively. After this interval, the concentration decreases to lower values as 186 ppm, 216 ppm, 386 ppm, 180 ppm, 306 ppm, 265 ppm, and 153 ppm at the depth intervals 38-40 cm, 44-46 cm, 50-52 cm, 58-60 cm, 70-72 cm, 76-78 cm, and 86-88 cm respectively. At the depth intervals of 100-102 cm and 152-154 cm the concentration further decreases to 27 ppm and 42 ppm respectively.

5.3 Organic carbon

Organic matter is virtually completely decomposed and recycled in soils (Reshmi, 2004). Carbon-rich sediments may contain both inorganic and organic carbon. Inorganic carbon may be present as both biogenic and abiogenic carbonate. Sedimentary organic carbon is mainly in the form of particulate organic carbon that has settled out from the overlying water column. Organic matter in surface sediments plays a major role in influencing community structure and metabolism of benthos. The main carbon species in suspended and bottom sediments include carbohydrates, fats, and proteins from decomposing biota, humic acids from decomposing humus, and carbonates from detritus.

Both resistant autochthonous and allocthonous organic residue sum up in sediment derived mainly from continental weathering (Priju, 2004). The biogenetic and/or geogenetic, origin of marine sediments is reflected not only in their mineralogy and lithology, but also in character of the soluble and insoluble organic matter which they contain (Simoneit, 1978). The study of organic matter is especially interesting as it throws light on the role of organic matter in adsorption and desorption of trace elements (Rashid, 1974). Earlier, Murty and Veerayya (1972a) have recorded an average of 1.5% organic carbon in the bulk sediments of the Vembanad estuary. Sajan and Damodaran (1981) have reported an average of 2.4% for the bulk sediments of Ashtamudi lake. A comparison of the organic carbon content in the bulk sediments of the Central Vembanad estuary with that of the world nearshore sediments indicates 1.5 fold increase in the former than the latter.

Total average organic carbon content in G8 core, G10 core, and G11 core is indicated by 5.77 %, 5.20 %, and 5.09 % respectively as shown in figure 5.1. The organic carbon % in G8 core does not vary much and remains precise to the average organic carbon % except at 26 – 42 cm depth. The surficial layers show depleted organic carbon %. In G10 core, the organic carbon % is minimum at the depth interval 96-98 cm and highest at 120-122 cm. In G11 core, the organic carbon % is enhanced in the interval from 48 cm to 110 cm. Remaining upper and lower part of the core are depleted in organic carbon %.

5.4 Enrichment Factor (EF) of trace elements

The enrichment factor (EF) reflects the geochemical background of the contaminant elements in estuarine or marine sediments (Turekian and

Wedepohl, 1961; Förstner and Muller, 1973; Macias et al., 2006). Based on the EF value, the emergence of an element can be distinguished, i.e. whether it is from natural or anthropogenic origin (Liu et al., 2003). The mean Enrichment factor is shown in table 5.1 and plotted in figure 5.2. Elements having EF values <1 reveal a natural origin, whereas values >1 indicate enrichment due to anthropogenic inputs (Mil-Homens et al., 2006).

The mean enrichment factor of cobalt, nickel, copper, zinc, and lead in G8 core is 0.77, 1.45, 30.53, 1.53, and 2.61 respectively. A strong anthropogenic influence is showed by all the elements excluding cobalt. A very high enrichment is indicated by copper. Similarly, 0.16, 0.24, 25.93, 0.61, and 8.28 are the mean enrichment factors for cobalt, nickel, copper, zinc, and lead in G10 core. Copper and lead show a very high enrichment. G11 core indicates an extreme mean enrichment factor of lead and copper as 54.67 and 32.73 respectively. The mean enrichment factor of cobalt, nickel, and zinc in G11 core are 0.05, 0.31, and 0.65 respectively.

5.5 Pollution Load Index (PLI)

Pollution Load Index (PLI) represents the number of times by which the metal content in the sediment exceeds the background concentration, and gives a summative indication of the overall level of heavy metal toxicity in a particular sample. PLI serves as a very effective tool in identifying the pollution level in a particular area (Satyanarayana et al., 1994; Angulo, 1996; Panda et al., 1995; Rubio et al., 2000). The Tomilson's Pollution Load Index for the samples was calculated using the heavy metal data and metal concentration for the world shale average (Wedepohl, 1971; Groumet et al.,1984) as background value are given in table 5.2. The PLI value >1 is polluted whereas <1 indicates no pollution (Seshan, 2010).

Table 5.1: Down core variation and mean of Enrichment factor (EF) of trace elements in nearshore sediments

Sl No.	Core No.	Depth (cm)	Co	Ni	Cu	Zn	Pb
1	G8	0--2	0.67	1.21	26.14	1.45	3.50
		6--8	0.71	1.32	28.72	1.35	1.53
		18-20	0.75	1.55	30.08	1.65	1.92
		30-32	0.80	1.50	32.64	1.42	1.17
		42-44	0.82	1.57	33.32	1.53	3.61
		48-50	0.81	1.43	31.65	1.73	3.39
		102-104	0.78	1.58	31.77	1.68	3.71
		138-140	0.74	1.25	27.65	1.41	2.86
		168-170	0.80	1.60	32.83	1.54	1.77
		Mean	0.77	1.45	30.53	1.53	2.61
2	G10	0--2	0.13	0.21	24.32	0.49	8.01
		6--8	0.23	0.24	26.01	0.53	6.99
		18-20	0.25	0.24	26.04	0.71	8.42
		30-32	0.14	0.25	26.38	0.56	9.90
		48-50	0.12	0.24	26.54	0.80	9.04
		102-104	0.13	0.28	26.75	0.68	8.06
		120-122	0.17	0.26	25.66	0.58	7.98
		148-150	0.14	0.24	25.63	0.58	7.28
		172-174	0.16	0.25	26.07	0.61	8.85
		Mean	0.16	0.24	25.93	0.61	8.28
3	G11	0--2	0.01	0.32	34.32	0.62	185.99
		4--6	0.01	0.31	33.70	0.54	182.91
		10--12	0.01	0.29	30.94	0.51	82.91
		20-22	0.01	0.29	31.73	0.57	83.16
		26-28	0.01	0.30	33.74	0.57	84.16
		38-40	0.04	0.28	31.16	0.55	15.06
		44-46	0.05	0.57	47.46	1.13	21.60
		50-52	0.02	0.29	31.36	0.57	24.18
		58-60	0.04	0.28	32.20	0.70	15.30
		70-72	0.03	0.28	31.38	0.63	20.96
		76-78	0.03	0.28	31.57	0.54	18.48
		86-88	0.05	0.28	30.94	0.54	13.12
		100-102	0.27	0.27	29.61	0.98	9.62
		152-154	0.17	0.28	28.14	0.59	7.93
		Mean	0.05	0.31	32.73	0.65	54.67

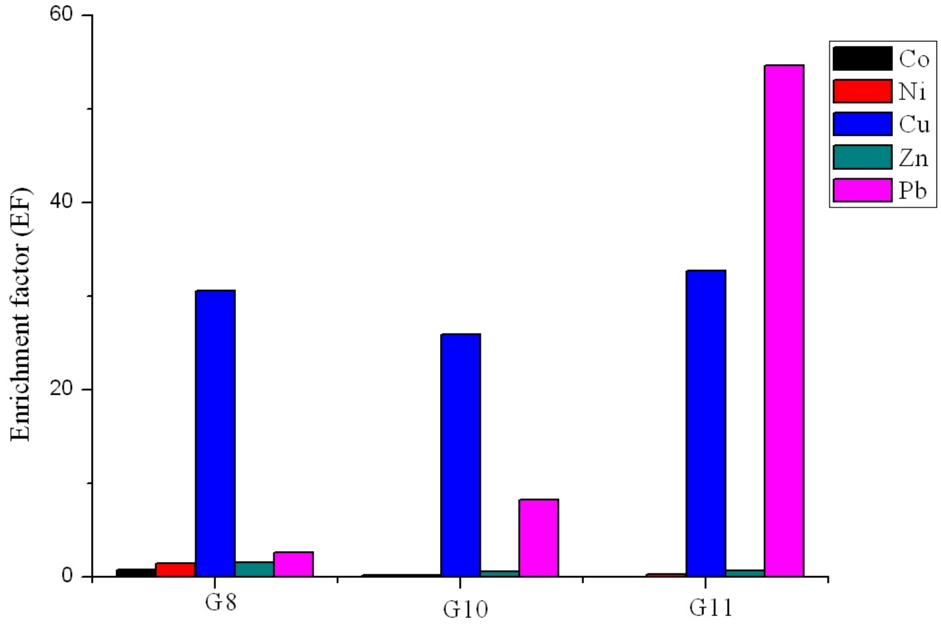


Figure 5.2: Mean enrichment factor of trace elements in nearshore cores

Table 5.2: Down core variation of Pollution load index (PLI) in nearshore sediments

Depth (cm)	PLI (G8)	Depth (cm)	PLI (G10)	Depth (cm)	PLI (G11)
0--2	0.97	0--2	1.02	0--2	4.98
6--8	0.89	6--8	0.98	4--6	4.81
18-20	1.07	18-20	1.08	10--12	3.60
30-32	0.87	30-32	1.20	20-22	3.71
42-44	1.13	48-50	1.29	26-28	3.84
48-50	1.10	102-104	1.24	38-40	1.80
102-104	1.19	120-122	1.11	44-46	3.10
138-140	0.96	148-150	1.13	50-52	2.35
168-170	1.00	172-174	1.16	58-60	1.89
				70-72	2.19
				76-78	2.01
				86-88	1.69
				100-102	1.25
				152-154	1.17

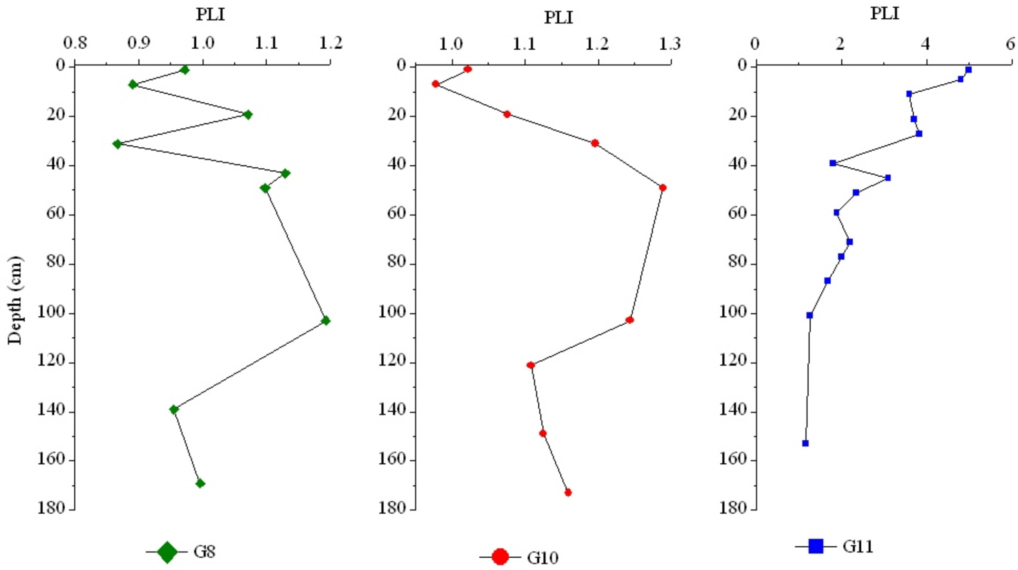


Figure 5.3: Down core variation of Pollution load index (PLI) of trace elements in nearshore sediments

G8 core reflects an increasing downcore trend where the surficial sediments have PLI value of 0.97, 0.89 at 0-2 cm and 6-8 cm depth. At the depth intervals of 30-32 cm and 138-140 cm, PLI is 0.87 and 0.96. PLI represents a high degree of pollution (PLI>1) at 42-44cm, 48-50cm, 102-104 cm, and 168-170 cm. G10 core shows PLI > 1 throughout the core. However, at 6-8 cm depth interval PLI is 0.98. An extreme pollution is shown by G11 core where the degree of pollution has been continuously increasing from bottom to the surficial sediments. The Surficial sediments show a very high PLI value of 4.98 and 4.81 at 0-2 cm and 4-6 cm. PLI ranges 3.6, 3.71, 3.84, and 3.10 at the depth intervals of 10-12 cm, 20-22cm, 26-28 cm, 44-46 cm respectively. Relatively low PLI value of 1.80 is shown at the depth interval of 38-40 cm. Low but still polluted sediments with PLI values 1.69, 1.25, and 1.17 are the bottom sediments at the depth intervals of 86-88, 100-102 cm, and 152-154 cm respectively. PLI values of 2.35, 2.19, and 2.01 are at the depth intervals of 50-52 cm, 70-72cm, and 76-78 cm respectively.

5.6 Geo-accumulation Index (I_{geo})

The geo-accumulation index is a quantitative measure of the extent of contamination in aquatic sediments (Förstner et al., 1990). It is calculated as $I_{geo} = \log_2 [C_n/1.5* B_n]$ Where, C_n is the examined element in the sediment and B_n is the geochemical background of a given element. The classification of sediments depends on the I_{geo} value (Muller, 1979). Thus: Igeo >5 = extremely contaminated, 4-5 = strongly to extremely contaminated, 3-4 = strongly contaminated, 2-3 = moderately to strongly contaminated, 1-2 = moderately contaminated, 0-1 = uncontaminated to moderately contaminated and <0 = uncontaminated. This classification is a methodological approach based on the geochemical data. It helps in mapping the study area and discriminating different sub-areas according to their degree of contamination. In addition it is feasible to obtain a proper comparison between various marine areas in terms of their heavy metal quality (Magesh et al., 2011). The Igeo values of the present study are summarized in the table 5.3 and plotted in the figure 5.4.

I_{geo} of G8 core shows an uncontaminated to moderately contaminated behaviour with an average I_{geo} value of 0.12, 0.15, 0.40, 0.22, and 0.24 for cobalt, copper, lead, nickel, and zinc respectively. Similarly, G10 core also shows an uncontaminated to moderately contaminated trend for cobalt, copper, lead, nickel, and zinc with average values of 0.13, 0.18, 0.42, 0.24, and 0.27 respectively. Lead (4.61) and zinc (1.78) have strongly to extremely contaminated and moderately contaminated average I_{geo} values respectively in G11 core. Cobalt, copper, and nickel show uncontaminated to moderately contaminated average values of 0.16, 0.19, and 0.37 respectively.

Table 5.3: Down core variation and mean values of geo-accumulation index in nearshore sediments

Sl No.	Core No.	Depth (cm)	Co	Cu	Pb	Ni	Zn
1	G8	0--2	0.10	0.12	0.54	0.19	0.22
		6--8	0.11	0.16	0.24	0.20	0.21
		18-20	0.12	0.22	0.30	0.24	0.26
		30-32	0.13	0.14	0.18	0.23	0.22
		42-44	0.13	0.15	0.56	0.24	0.24
		48-50	0.12	0.14	0.52	0.22	0.27
		102-104	0.12	0.18	0.57	0.24	0.26
		138-140	0.11	0.13	0.44	0.19	0.22
		168-170	0.13	0.16	0.27	0.25	0.24
		Mean	0.12	0.15	0.40	0.22	0.24
2	G10	0--2	0.11	0.14	0.47	0.19	0.26
		6--8	0.13	0.15	0.28	0.24	0.23
		18-20	0.13	0.21	0.26	0.25	0.27
		30-32	0.13	0.17	0.48	0.24	0.32
		48-50	0.13	0.24	0.54	0.24	0.29
		102-104	0.15	0.20	0.50	0.25	0.26
		120-122	0.14	0.17	0.38	0.24	0.26
		148-150	0.13	0.17	0.47	0.24	0.24
		172-174	0.13	0.18	0.40	0.25	0.29
		Mean	0.13	0.18	0.42	0.24	0.27
3	G11	0--2	0.17	0.18	9.97	0.55	6.04
		4--6	0.16	0.16	9.91	0.54	5.94
		10--12	0.15	0.15	8.98	0.36	2.69
		20-22	0.15	0.17	8.97	0.37	2.70
		26-28	0.16	0.17	9.02	0.41	2.73
		38-40	0.15	0.16	1.86	0.28	0.49
		44-46	0.30	0.33	2.17	0.63	0.70
		50-52	0.15	0.17	3.87	0.30	0.79
		58-60	0.15	0.21	1.81	0.29	0.50
		70-72	0.15	0.18	3.07	0.29	0.68
		76-78	0.15	0.16	2.66	0.29	0.60
		86-88	0.15	0.16	1.54	0.29	0.43
		100-102	0.14	0.29	0.27	0.29	0.31
		152-154	0.15	0.17	0.42	0.26	0.26
		Mean	0.16	0.19	4.61	0.37	1.78

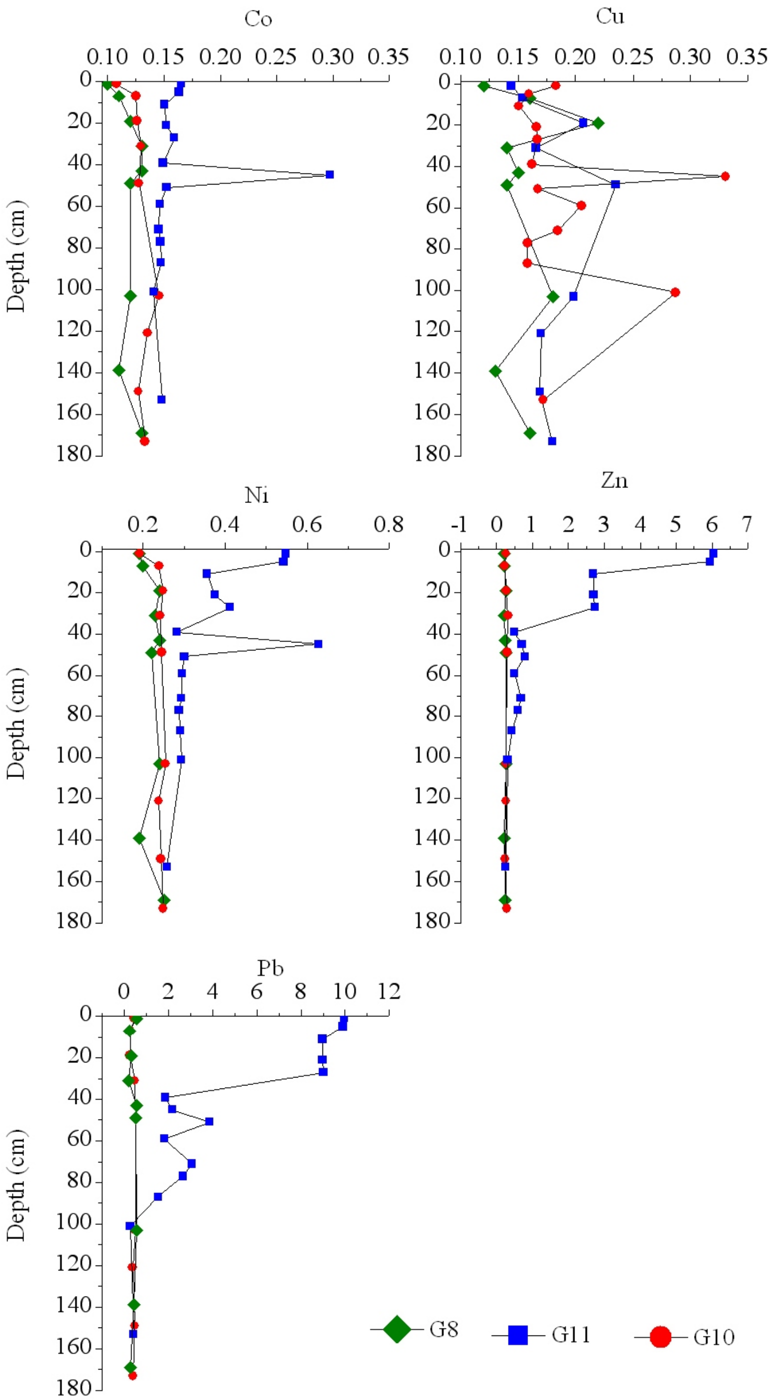


Figure 5.4: Down core variation of geo-accumulation index (Igeo) of trace elements in nearshore sediments

5.7 Summary

Trace elements are present in natural waters and their sources are associated with either natural processes or man's activities. The basic natural processes contributing trace elements to waters are chemical weathering of rocks and soil leaching (Garcia-Montelongo et al., 1994; Jordao et al., 2002). Both processes also may be largely controlled by biological and mineralogical factors (Kabata-Pendias, 2001). Trace metals cycling and their fate in coastal wetland, a typical dynamic ecosystem, have fascinated many researchers. Study of geochemical characters of major and trace components in sediment cores from an intertidal region could reveal the sedimentary environment and pollution history of the area (Fernandes et al., 2011). Copper, lead, and zinc are trace metals, which usually have an anthropogenic origin (Alvarez-Iglesias et al., 2003).

Cobalt concentration in the cores ranges from 10 ppm to 30 ppm with an average of 12 ppm, 13 ppm, and 16 ppm in G8, G10, and G11 cores respectively. The average concentrations of nickel for G8, G10, and G11 cores are 76 ppm, 81 ppm, and 125 ppm respectively. Copper ranges on an average of 39 ppm, 45 ppm, and 47 ppm in G8, G10, and G11 cores respectively. Zinc concentration is relatively higher in the three cores with an average value of 112 ppm, 127 ppm, and 841 ppm for G8, G10, and G11 respectively. Lead concentration is moderately high among the trace metals in G8 and G10 cores whereas it is higher in G11 core with an average concentrations of 40 ppm, 42 ppm, and 459 ppm. Enrichment in lead concentration is observed at the surface levels of all the three cores.

Both resistant autochthonous and allocthonous organic residue sum up in sediment derived mainly from continental weathering (Priju et al., 2004). The study of organic matter is especially interesting as it throws light on the role of organic matter in adsorption and desorption of trace elements (Rashid, 1974). Total average organic carbon content in G8 core, G10 core, and G11 core is indicated by 5.77 %, 5.20 %, and 5.09 % respectively as shown in figure 6.3.

The enrichment factor (EF) reflects the geochemical background of the contaminant elements in estuarine or marine sediments (Turekian and Wedepohl, 1961; Förstner and Muller, 1973; Macias et al., 2006). The mean enrichment factor of cobalt, nickel, copper, zinc, and lead in G8 core is 0.77, 1.45, 30.53, 1.53, and 2.61 respectively.

Pollution Load Index (PLI) represents the number of times by which the metal content in the sediment exceeds the background concentration, and gives a summative indication of the overall level of heavy metal toxicity in a particular sample. All three cores reflect an increasing down core trend of PLI. From the PLI values it can be concluded that the region is moderately to highly polluted. The Pollution index (PLI > 1) is increasing spatially from G8, G10, and G11 core respectively.

The geo-accumulation index is a quantitative measure of the extent of contamination in aquatic sediments (Förstner et al., 1990). I_{geo} of G8 core shows an uncontaminated to moderately contaminated behaviour with an average I_{geo} value of 0.12, 0.15, 0.40, 0.22, and 0.24 for cobalt, copper, lead, nickel, and zinc respectively. Similarly, G10 core also shows an uncontaminated to moderately contaminated trend for cobalt, copper, lead, nickel, and zinc with average values of 0.13, 0.18, 0.42, 0.24, and 0.27 respectively. Lead (4.61) and zinc (1.78) have strongly to extremely contaminated and moderately contaminated average I_{geo} values respectively in G11 core. Cobalt, copper, and nickel show uncontaminated to moderately contaminated average values as 0.16, 0.19, and 0.37 respectively.

Trace metal variations in the three cores indicates that the source rock could range from mafic to acidic rocks which are present (Charnockite, Granitic Gneiss) in the surrounding regions. Hence, local source rock could be the source for the trace elements in the sediments. Anthropogenic factor has also enhanced to a very high degree of the trace metal concentration in the sediments. Sudden increase (at the depth interval of 44-46 cm onwards) in the trace metals like lead, zinc, and nickel in G11 core reveals the anthropogenic contribution in polluting the area. The possible reason for this enhancement could be due to ship transportation in the region. In addition the three cores were recovered from locations near to the Munambam fishing harbor and the navigation of the fishing boats also contribute to the increased trace metal concentration. Concentration of these trace elements is high in G8 and G10 core too supports the anthropogenic influence in space. Based on the temporal distribution of trace metals, it is inferred that accumulation of Pb, Zn and Ni in the recent past has increased in the coastal sediments. This is supported by the lesser concentration in the bottom sediment of the cores. Higher enrichment factor (Fig. 4.5) is recorded for Pb and Cu in all the cores. The Pollution index (PLI > 1) is increasing spatially from G8, G10, and G11 core respectively. The enrichment factor and PLI, which are the indices of sediment toxicity, are found higher in the study area when compared to background values. I_{geo} also supports the high degree of contamination in the region. Hence, it can be inferred that not only the industrial effluents but also domestic sewage, navigation, fishing and rapid urbanization are becoming a major threat for the coastal region in central Kerala.

****

Chapter 6

SEDIMENTATION HISTORY OF CENTRAL KERALA COAST: INFERENCES FROM TEXTURAL, GEOTECHNICAL, AND PALYNOLOGICAL ASPECTS

Contents

6.1 Introduction

The coastal zone is an important ecosystem with high productivity, dense population, exploitation of renewable and non-renewable resources, discharge of effluents, and municipal sewage, development of industries & spurts in recreational activities (Narayana & Priju, 2006). The south western coastal sedimentary basins of India are mainly located offshore (Jayalakshmi et al., 2004). The coastal stretch of Kerala is ~560 km (Narayanan & Anirudhan, 2003) and the width of coastal land varies between 5-30 km, where coastal sand dunes and beach ridges interrupt the monotony of morphology (Narayana & Priju, 2006). Beaches, beach-cliffs, stacks, shore-platforms, spits, and bars, beach ridges, estuaries and lagoons, mudflats and tidal flats, and deltas are the coastal geomorphic features of Kerala (Nair, 1999). Cliffs of crystalline rocks and Tertiary exposure predominate between Vizhinjam to Paravur and wide sandy beaches, strand lines, lagoons, and mudflats, from Paravur to Kochi, are dominating geomorphic features (Narayanan & Anirudhan, 2003).

Studies of sediment texture, clay minerals, geotechnical, and pollen records can be used as reliable parameters for understanding environmental deposition and sedimentation history of coastal environments. Texture helps in determining the litho characters and describing the facies of a body of sediments which can provide the information about the depositional environments (Dalrymple, 1992). The properties of different clast sizes and the texture of terrigeneous clastic sediments can provide information about the history of transport, of the weathering and the environment of deposition.

Clay mineral studies have a wide range of application in geosciences. As clay respond to their chemical and thermal environment and their properties and species change accordingly (Velde, 1992), a sound knowledge of nature and distribution of clay minerals is very essential to understand the process of weathering. The distribution of various clay minerals in sediments is used as a tool for provenance studies, environmental analysis and paleoclimatic studies (Elliot et al., 1997; Srivastava et al., 1998, etc.).

Geotechnical properties of sediments provide an insight into their behaviour when natural state of deposition is changed to manmade or artificial conditions (Khadge, 2002). The plastic limit is the water content at which the soil can just be rolled into a 3mm diameter thread without crumbling whereas water content of cohesive soils are usually correlated to empirically defined boundaries between states of consistency i.e. non-plastic, plastic, and viscous liquid, are known as Atterberg limits. Liquid limit is the water content at which a standard V-groove cut in the soil just closes when it is set to fall in a standard way using Casagrande apparatus.

Pollen record study deals with the analysis of pollen grains produced by higher plants (angiosperms and gymnosperms) which contains the male genetic material responsible for reproduction. Pollen grains ranges in size from 10µM to 150µM and are protected by a chemically resistant outer layer called as exine (Bradley, 1999). Pollens are considered to be aeolian sediment which on falling on the site of accumulation of sediments becomes the part of stratigraphic record (Traverse, 1994). Hence, pollen analysis can provide the vegetation history and informations about past climatic conditions. Sediments and peat records of the coastal environments present an ideal archive to study the past fluctuations in climate and environmental systems. The temporal variations in sedimentation rates and pollen records serve as good proxies to understand the Late Quaternary palaeoclimate.

Abundant organic rich deposits called as peats have been identified at many onshore locations adjacent to the present shoreline along the west coast of India (Agarwal & Gazdar, 1974; Kale & Rajagarn, 1983; Rajendran et al., 1989; Bruckner, 1989; Tissot, 1990; Caratini et al., 1990; Caratini & Rajagopalan, 1992; Narayana et al., 2001b). Mangrove peat stratigraphy worldwide reveals that peat is not recorded during early Holocene since a rapid sea level rise precludes growths of mangroves, chokes them with high sediment input, redistributes organic sediment, mixes with all allocthonous material, all due to higher wave activity and erosive tendency of the rising sea (Ellision & Stoddart, 1991). Mangroves established expansive colonies when sea level stabilized (Parkinson, 1989). Therefore, ideal condition for peat formation prevailed during mid-Holocene times when the first extensive mangrove swamps appeared in sheltered areas particularly between 7000 to 5500 years BP and later.

The geological properties explain the textural and compositional aspects of the sediment, whereas the geotechnical properties describe the deformational response of the sediments to loads. Sediment and peat samples from a borehole drilled in the Munambam palaeodelta (Narayana et al., 2001a) to a depth of 40 m near Cochin, southwest coast of India were recovered and analyzed for ^{14}C ages, sediment characteristics (textural and geotechnical), and pollen record analysis to understand the environmental changes during the Late Quaternary.

6.2 Geochronology- ^{14}C Ages

Radiometric dating has become most important technique to determine the age of the rocks, fossils, sediments, etc. Radiocarbon dating is widely used in paleoclimatological and paleoenvironmental studies worldwide, among numerous studies from different fields of research. In this study, radiocarbon ages spanning the total borehole length are discussed in relation to textural, geotechnical, ad palynological parameters.

The chronology of the borehole samples recovered from Paravur coast (Central Kerala) was subjected to find the ^{14}C ages of the peat samples. The statistical error on the age is one σ value. Peat deposits from this study represent the calibrated age of 4692 and 9550-9330 years BP at the depth of 6.13-6.23 and 24.40-24.50 meters respectively. At the depth interval of 31.40-31.50 meter, the calibrated age is 11689-11198 years BP. 11912- 11233 years BP is the age of the bottom depth of the borehole at 39.75-39.85 meters. Contrasting older calibrated ages of 12116 and 10330 years BP was found at the depths of 15.31-15.35 and 22.65-22.95 m respectively. This older age indicates the reworking of the sediments either by wave action or activities of organism or both.

The radiocarbon dates reported by early workers (Agarwal et al., 1970; Powar et al., 1983; Rajendran et al., 1989; Shajan, 1998; Tissot, 1990; Narayana et al., 2001, 2007) and on the implication of on-shore occurrence of peat deposits in the coastal land of south-west India are given in table 6.8. Mangroves colonies established after the stabilization of sea level (Parkinson, 1989) and peats formed ideally during mid-Holocene during 7000-5500 yrs BP and later (Woodroffe et al., 1989; Ellison and Stoddart, 1991). Tissot (1990) on the basis of continental & marine microfossil ratio and their frequencies, suggested that the sea level remained constant since 6000yrs. BP. Peat deposits at various depths in the sediments and their radiocarbon ages strongly indicate that the southwest coast was covered with abundant mangrove vegetation during >40,000 yrs. BP (Narayana, 2007).

6.3 Sediment Texture

Grain size characteristic is used to identify and describe depositional processes and sedimentary environment (Visher, 1969; Tucker & Vacher, 1980). The grain-size distributions of sediments are studied in two ways: (i) the study of grain size in relation to mechanism of sediments transportation and deposition, and (ii) the study of grain size as an environmental indicator. In all these studies, surface sediments were employed to differentiate the modern sedimentary environments but in coastal environments, sediment transport and depositional processes are more complicated and are governed by both fluvial and marine processes.

Spatial and temporal variations in sediment characteristics and accumulation rates can be used to study the coastal processes. Differential erosion, transportation, and deposition reflect different depositional environments and may possess distinctive particle size distributions. By determining the particle size distributions, it is possible to hypothesize about the environmental reconstruction and so to utilize this technique as a tool for reconstruction of environments (Lario et al., 2002). In order to get significant relationships between sediment properties and geological processes or settings, quantitative analysis of the size distributions of the sediments is needed.

Jayalakshmi et al., (2001) reports the ~ 70 m Quaternary sediment thickness in the Kerala coast where oldest part is unfossiliferrous consisting of grey clay with admixture of lithoclasts of laterite derived from adjacent terrestrial regions. Prakash and Prithviraj (1988) reports predominance of clayey silts and silty clays in the sediment distribution of the inner shelf zone of the central Kerala coast. He also reports that the clayey silts extend over the inner portion of the shelf adjoining the coastline and estuaries whereas silty clays dominate the central and the seaward portion of the inner shelf. Knowledge of various characteristics of sediment inputs permits the assessment of its state of evolution, its maturity and also a better understanding of its history (Priju, 2004). Veerayya and Murty (1974) have revealed the coarser sand fraction in central and southern parts of Vembanad lake and finer fraction in the remaining areas. The textural studies of surface sediments of central Vembanad lake by Seralathan et al. (1993) and Seralathan and Padmalal (1994) report the muddy sand, sandy mud, sand, mud, clayey sand, and silty sand facies. Sunil and Antony (1994) and Badarudeen et al. (1996) have reported an admixture of sand and silt (silty sand and sandy silt) in the sediments of the mangrove areas in the central and southern parts the Vembanad Lake.

Data on the textural parameters of sediment samples recovered from a borehole drilled for 40 meter depth below the surface, are given in Table 6.1.

The sand percentage ranges from 0 – 95 with an average value of 39% (Table 6.4). Up to 6 m depth the sand percentage ranges from 74 to 95. Below 6 m it generally shows a decreasing trend except at certain depths. Silt ranges from 5 – 58 % with an average value of 34%. Clay ranges from 0 – 54 % and has an average value of 28 % (Table 6.4). The litholog of the Paravur borehole (Fig. 6.3) shows downward dominance of peat with silty clay intercalations from 24 m to 40 m. Mud has high abundance at the depth range of 17 m to 24 m. Silty sand overlies a thin layer of sandy mud at depth range of 14.5 m to 16 m. Sandy mud is the textural character in the depth range of 10 m to 14.5 m. The litho character when moved upward from 10 m to the surficial sediments, alternating sand and mud intercalations are observed. However, muddy sand is dominating in their intercalations. Sand shows a dominating character just below the surficial muddy sand sediments at the depth range 1.95 m to 3 m. Figure 6.2 shows the opposite and similar variations of sand & silt, and silt & clay respectively with depth. According to the Folk's classification (Fig. 6.1) the sediments exhibit sandy, muddy sand, sandy mud, and muddy texture. The sediment texture shows an increasing trend for sand from bottom to the top of the borehole whereas silt and clays has a decreasing trend from bottom to top (Fig. 6.2). From 6 to 6.5m depth the sediments show sandy mud texture. Beyond 19 m depth the sediments exhibit sandy mud to muddy texture. Hence, in a nutshell, the borehole shows a coarsening upward sequence of sediments. The borehole reveals muddy sand, sand, sandy mud, silty sand, and mud facies when moved surficial to subsurficial.

Table 6.1: Physical and textural parameters of Paravur borehole

Borehole depth (m)	Wet bulk density (g/cc)	Moisture content (%)	Organic carbon (%)	Sand (%)	Silt (%)	Clay (%)	Textural nomenclature after Folk (1980)
0.95 - 1.0	1.84	28	0.30	89	6	5	Muddy sand
1.95-2.0	1.89	24	0.10	95	5	0	Sand
2.85-3.0	1.90	26	0.30	93	5	2	Sand
3.9-4.0	1.91	29	0.84	74	13	13	Muddy sand
5.20-5.30	1.80	42	0.75	83	6	11	Muddy sand
5.40-5.50	1.86	37	0.70	82	8	10	Muddy sand
6.05-6.13	1.46	79	5.56	28	31	41	Sandy mud
6.31-6.36	1.54	66	4.93	49	25	26	Sandy mud
6.46-6.50	1.39	98	5.40	21	38	41	Sandy mud
7.40-7.50	1.64	49	2.72	58	18	24	Muddy sand
8.70-8.80	1.77	43	1.90	61	23	16	Muddy sand
8.90-9.0	1.57	66	4.16	40	33	27	Sandy mud
10.15-10.17	1.83	30	2.14	53	25	22	Muddy sand
10.25-10.28	1.90	26	0.93	78	13	9	Muddy sand
10.46-10.50	1.83	38	0.97	79	17	4	Silty sand
11.55-11.60	1.56	67	3.20	37	43	20	Sandy mud
11.80-11.85	1.67	54	2.22	38	45	17	Sandy mud
11.98-12.0	1.78	43	1.65	46	44	10	sandy silt
13.70-13.80	1.55	67	3.08	46	37	17	Sandy mud
15.0-15.05	1.72	45	2.34	67	24	9	Muddy sand
15.24-15.31	1.46	77	5.30	19	42	39	Sandy mud
15.48-15.50	1.57	68	2.87	16	43	41	Sandy mud
16.85-17.0	1.92	25	3.15	78	18	4	silty sand
17.90-18.0	1.83	26	1.11	86	11	3	silty sand
18.82-18.92	1.53	68	5.88	27	40	33	Sandy mud
19.65-19.75	1.51	75	6.97	4	46	50	Mud
19.95-20.0	1.53	70	6.02	3	50	47	Mud
21.20-21.30	1.52	70	5.42	33	36	31	Sandy mud
21.40-21.50	1.61	55	4.55	43	32	25	Sandy mud
22.65-22.75	1.51	71	5.79	18	47	35	Sandy mud
22.95-23.0	1.53	65	5.12	2	53	45	Mud
24.10-24.20	1.54	64	5.50	11	46	43	Sandy mud
24.30-24.40	1.56	63	4.72	0	50	50	Mud
25.40-25.50	1.50	61	5.69	13	45	42	Sandy mud
26.40-26.50	1.58	62	5.80	0	50	50	Mud
27.40-27.50	1.58	60	4.95	0	53	47	Mud
28.40-28.50	1.57	54	4.91	0	46	54	Mud
29.40-29.50	1.55	66	4.58	33	37	30	Sandy mud
30.20-30.30	1.60	55	5.00	38	33	29	Sandy mud
31.30-31.40	1.52	70	5.75	16	46	38	Sandy mud
32.40-32.50	1.55	59	5.82	0	58	42	Mud
33.40-33.50	1.56	64	6.88	1	47	52	Mud
34.80-34.90	1.57	65	5.52	60	26	14	Muddy sand
37.40-37.50	1.62	51	4.35	40	36	24	Sandy mud
38.40-38.50	1.53	70	7.31	22	39	39	Sandy mud
39.90-40.0	1.61	54	5.97	2	52	46	Mud

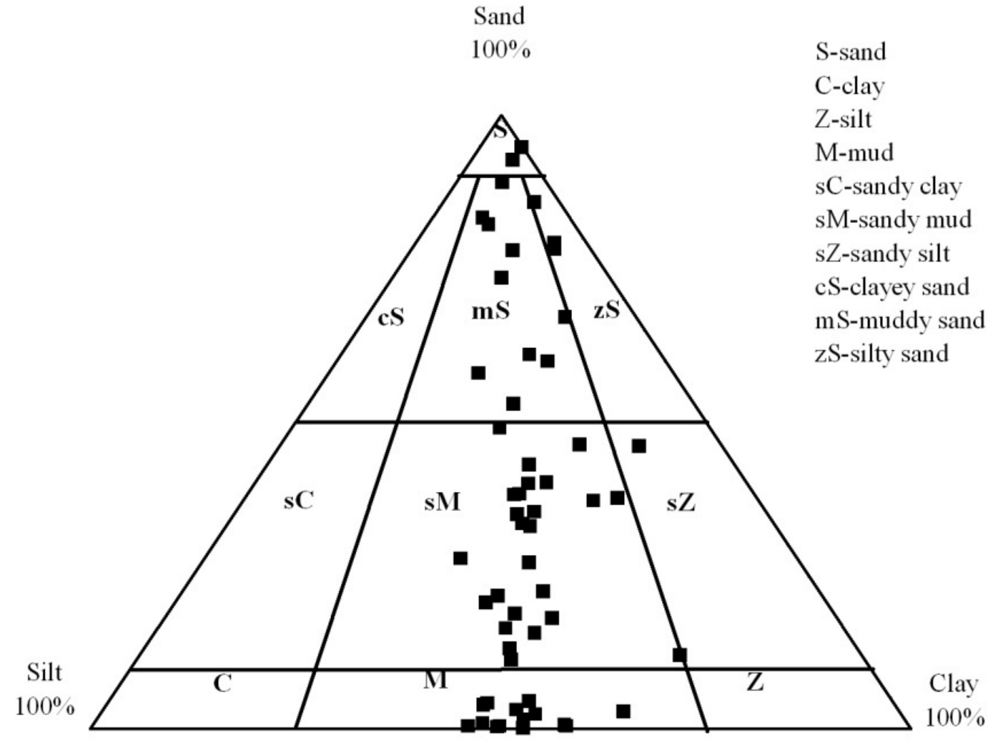


Figure 6.1: Ternary diagram showing sand silt clay ratios in Paravur borehole samples (after Folk et al., 1980)

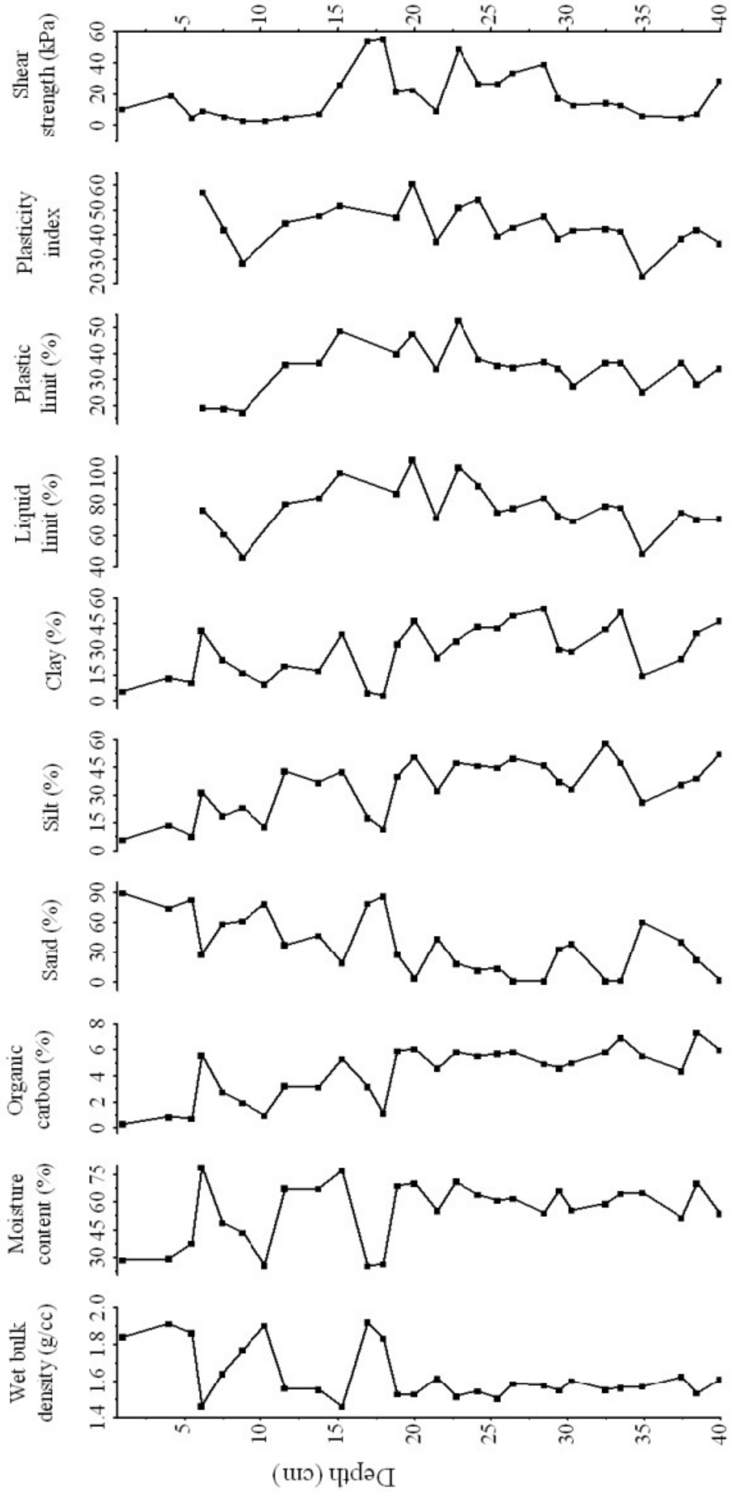


Figure 6.2: Down core variation of physical, textural and geotechnical parameters of Paravur borehole samples

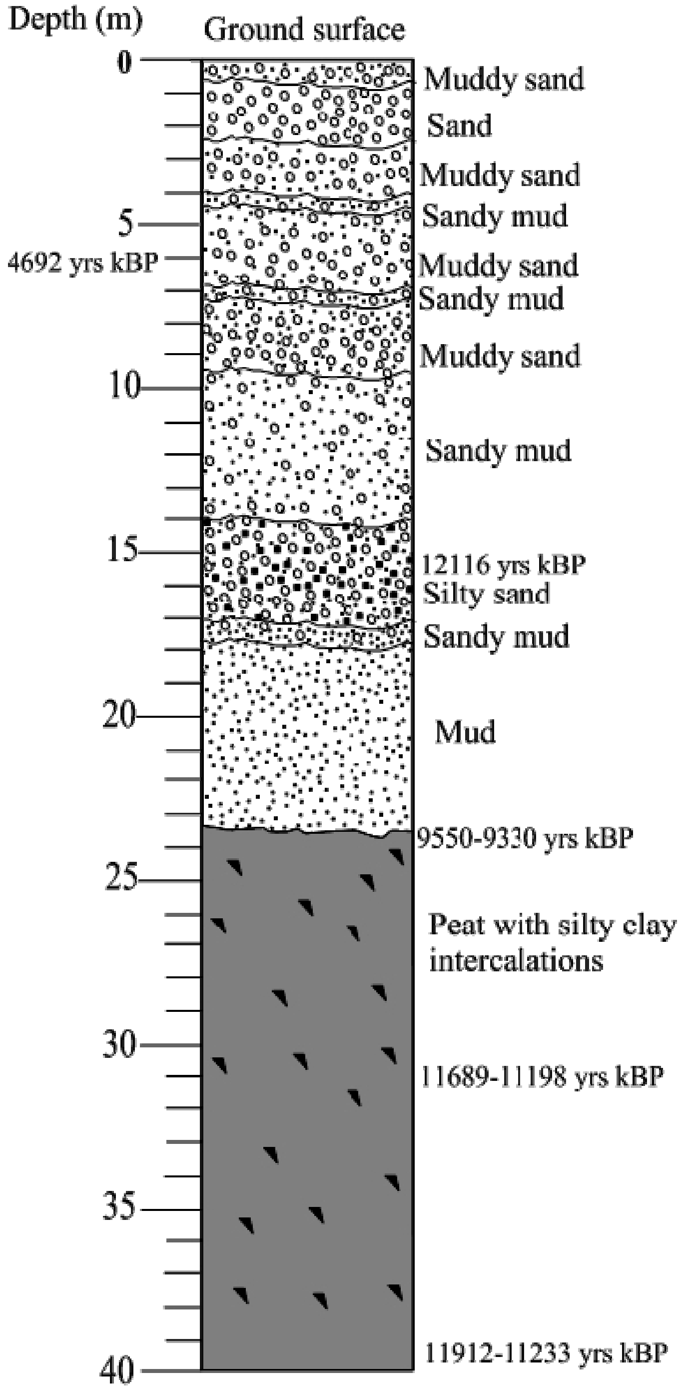


Figure 6.3: Lithofacies observed in vertical cross-section of Paravur borehole and radiocarbon dates of samples.

Table 6.2: Geotechnical properties of Paravur borehole

Sl No.	Borehole depth (m)	Liquid limit (%)	Plastic limit (%)	Plasticity index	Liquidity index (I_L)	Activity	Liquid limit wL/ Plastic limit wP	Shear strength (kPa)
1	0.85-1	--	--	--		--	--	10.33
2	4.10-4.20	--	--	--		--	--	18.58
3	5.38-5.50	--	--	--		--	--	4.65
4	6.13-6.23	76	19	57	1.05	1.40	0.25	8.78
5	7.50-7.60	60	19	42	0.72	1.76	0.31	5.16
6	8.70-8.80	46	17	28	0.92	1.78	0.38	2.58
7	10.28-10.33	--	--	--	--	--	--	2.58
8	11.60-11.67	80	36	45	0.71	2.22	0.44	4.65
9	13.70-13.90	84	36	48	0.65	2.77	0.43	7.23
10	15.14-15.21	100	48	52	0.55	1.33	0.48	25.81
11	16.85-17.0	--	--	--	--	--	--	53.68
12	17.9-18.0	--	--	--	--	--	--	54.71
13	18.72-18.92	87	40	47	0.61	1.43	0.46	21.16
14	19.75-19.95	108	47	61	0.45	1.81	0.44	22.19
15	21.30-21.50	71	34	37	0.57	1.49	0.48	8.78
16	22.75-22.95	103	53	51	0.36	1.47	0.51	49.03
17	24.10-24.30	92	37	54	0.49	1.26	0.41	26.84
18	25.40-25.50	74	35	39	0.65	0.93	0.47	26.84
19	26.40-26.50	77	34	43	0.63	0.86	0.45	33.55
20	28.40-28.50	84	37	47	0.36	0.88	0.44	38.71
21	29.30-29.40	72	34	38	0.83	1.28	0.47	17.03
22	30.30-30.50	69	27	42	0.67	1.46	0.4	12.39
23	32.40-32.50	79	36	42	0.53	1.01	0.46	13.94
24	33.40-33.50	78	36	41	0.68	0.80	0.47	12.75
25	34.80-34.90	48	25	23	1.73	1.61	0.52	5.68
26	37.40-37.50	75	36	38	0.39	1.58	0.49	4.65
27	38.40-38.50	70	28	42	1.00	1.06	0.4	6.86
28	39.85-40.00	70	34	36	0.54	0.78	0.49	28.39

6.4 Clay Mineralogy

Clay minerals are mainly the weathered products of rocks and soils on land. The compositions of sediments depend upon climatic conditions, provenance, and rock types of the region (Singer, 1984). The composition and distribution of clay minerals have been used as indicators of sediment dispersed in various environments (Biscaye, 1964; Grim, 1968, Rao et al., 1995). Clay mineral study is useful in better understanding of the origin, diagenesis, and depositional environment of the sediments. Numerous studies show that clay minerals indicate spatial and temporal changes of the geology and the weathering characteristics on the adjacent continental landmass (Biscaye, 1965; Kolla et al, 1976; Nair et al., 1982; Weaver, 1989). The continuous erosion of top layers of the adjacent landmasses causes accumulation of clay minerals which are transported by various natural agents and gets deposited in low lands. It is now well established from several studies that temporal variation of clay minerals can be used as a paleoclimatic proxy to understand past climatic condition provided they are of detrital origin (Thamban et al., 2002). Many researchers (Kolla et al., 1981; Nair et al., 1982; Rao, 1991; Chauhan, 1994) have reported clay mineral composition of sediments from the eastern Arabian Sea mainly based on surficial sediments to understand the provenance and transport pathways of fine grained terrigeneous sediments. Since the source and transport mechanism of clay minerals by now are fairly well identified in the area, it is possible to evaluate the utility of clay minerals as paleoclimatic proxies. The objective of clay mineral studies of the borehole was to understand their temporal variations to reconstruct the climatic events and sedimentation history during the Late Quaternary.

The clay % along with the physical and textural parameters of the Paravur borehole is shown in table 6.1 and plotted in figure 6.2. The total clay % in the sediments is low in the surficial part of the borehole. Similar low clay percentage was observed at the depths 10.25-10.28 m, 10.46-10.50 m, 16.85-17.0 m, and 17.90-18.0 m respectively. On an average, the total clay content in sediments is increasing downwards. Relative percentage of smectite, kaolinite, chlorite, and illite and their ratios were computed (tables 6.3 & 6.4) and plotted (fig 6.4 & 6.5). Maximum, minimum, and average smectite % was found as 71%, 10%, and 34% respectively.

Table 6.3: Relative abundance of clay minerals in an onshore borehole drilled at Paravur to a depth of 40m

Depth (m)	Relative Percentages				Ratios	
	Smectite	Kaolinite	Chlorite	Illite	Chlorite/Illite	Kaolinite/Chlorite
1.95-2.0	38	4	4	54	0.07	0.96
5.20-5.30	56	33	0	11	0.00	0.00
5.40-5.50	71	13	6	10	0.59	2.34
6.05-6.13	45	50	0	5	0.00	0.00
6.46-6.50	47	37	11	6	1.94	3.31
7.50-7.60	27	24	7	28	0.26	3.29
8.80-8.90	56	24	12	8	1.42	2.00
10.15-10.17	51	13	6	29	0.22	2.06
10.46-10.50	49	17	0	34	0.00	0.00
11.55-11.60	50	12	8	30	0.27	1.56
11.98-12.0	18	16	10	55	0.19	1.57
13.70-13.80	13	21	12	53	0.22	1.79
15.0-15.05	18	20	7	55	0.13	2.86
15.48-15.50	23	18	9	50	0.18	2.00
16.85-17.0	56	13	0	31	0.00	0.00
18.82-18.92	10	13	9	68	0.14	1.35
19.95-20.0	28	17	7	49	0.14	2.36
21.20-21.30	39	13	7	41	0.18	1.76
22.65-22.75	26	18	15	41	0.38	1.13
22.95-23.0	51	9	7	33	0.20	1.43
24.10-24.20	30	32	0	37	0.00	0.00
24.30-24.40	31	18	9	42	0.21	2.03
25.40-25.50	32	16	7	45	0.16	2.27
26.40-26.50	40	15	6	38	0.17	2.41
27.40-27.50	21	30	0	49	0.00	0.00
28.40-28.50	30	27	0	43	0.00	0.00
29.40-29.50	27	21	7	45	0.16	2.86
30.20-30.30	23	17	10	50	0.20	1.69
31.30-31.40	20	23	12	45	0.25	2.00
32.40-32.50	24	23	16	37	0.43	1.47
33.40-33.50	30	27	0	42	0.00	0.00
34.80-34.90	21	25	0	55	0.00	0.00
37.40-37.50	31	22	0	47	0.00	0.00
38.40-38.50	30	13	10	46	0.21	1.35
39.90-40.0	17	18	8	56	0.15	2.19

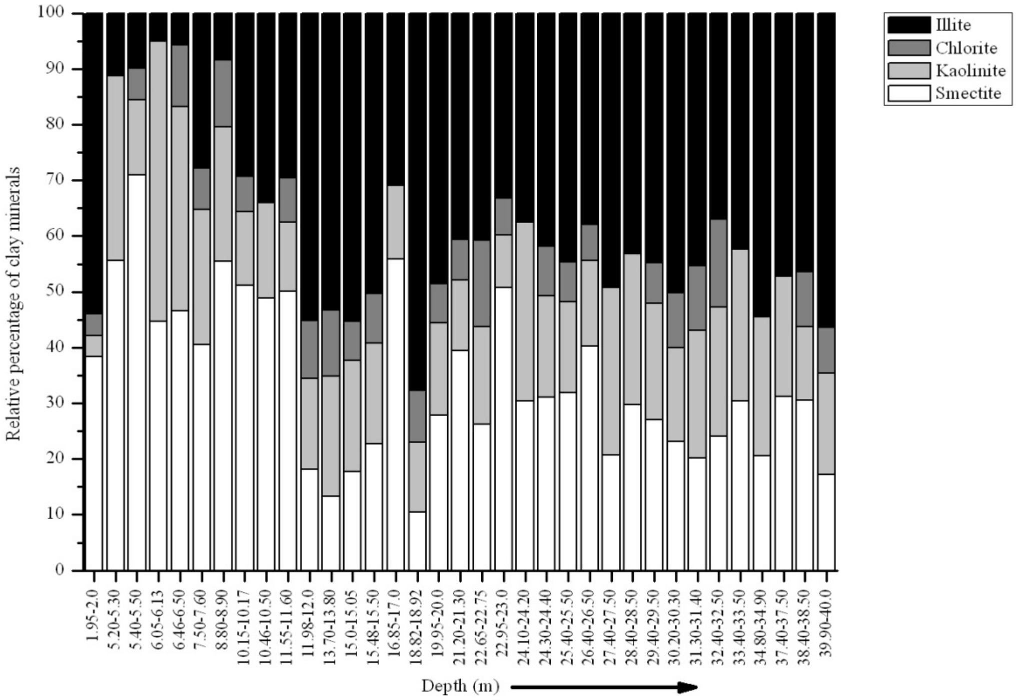


Figure 6.4: Relative percentage of clay minerals in Paravur borehole samples

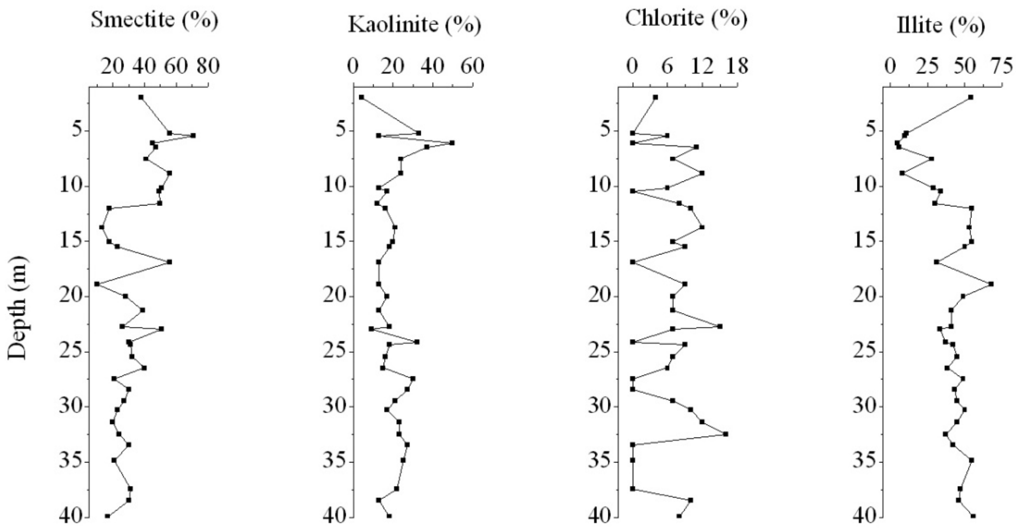


Figure 6.5: Down core variation of clay minerals in Paravur borehole drilled to 40m depth

Table 6.4: Showing maximum, minimum and average values of textural and geotechnical properties and clay minerals

PBH	γ (g/cc)	w (%)	C_{org} (%)	S (%)	Z (%)	C (%)	Su (kPa)	LL (%)	PL (%)	PI (%)	A	(wP/ wL)	Sm (%)	Ka (%)	Ch (%)	Ill (%)
Max	1.92	98	7.31	95	58	54	55	108	53	61	2.77	0.52	71	50	16	68
Min	1.39	24	0.10	0	5	0	3	46	17	23	0.78	0.25	10	4	0	5
Average	1.64	55	3.89	39	34	28	19	77	34	43	1.41	0.44	34	20	6	39

PBH: Paravur borehole, γ : wet bulk density, w: water content, C_{org} organic carbon, S:sand, Z:silt, C: clay, Su: undrained vane shear strength, LL: liquid limit, PL: plastic limit, PI: plasticity index, A activity, wP/wL: plastic limit/ liquid limit, Sm: smectite, Ka: Kaolinite, Ch: Chlorite, Ill: Illite

Table 6.5: Correlation Matrix of the physical and textural parameters of Paravur borehole samples (n = 46)

	Depth (m)	Wet bulk density (g/cc)	Moisture content (%)	Organic carbon (%)	Sand (%)	Silt (%)	Clay (%)
Depth (m)	1.00						
Wet bulk density (g/cc)	**-0.50**	1.00					
Moisture content (%)	0.37	**-0.97**	1.00				
Organic carbon (%)	**0.73**	**-0.87**	**0.81**	1.00			
Sand (%)	**-0.65**	**0.83**	**-0.76**	**-0.86**	1.00		
Silt (%)	**0.66**	**-0.78**	**0.72**	**0.80**	**-0.96**	1.00	
Clay (%)	**0.59**	**-0.81**	**0.74**	**0.85**	**-0.96**	**0.84**	1.00

Kaolinite has a maximum value of 50 % and minimum 4 % with an average of 20 %. Maximum chlorite % in the borehole is 16% with an average of 6 %. Chlorite was nil at depths 6.05-6.13 m, 10.46-10.50 m, 16.85-17.0 m, 27.40-27.50 m, 28.40-28.50 m, 33.40-33.50 m, 34.80-34.90 m, and 37.40-37.50 m. Illite was observed with a maximum value of 68%, minimum of 5 %, and an average of 39 %.

Correlation matrix of clay minerals exhibits a moderate negative correlation (-0.55) of smectite with depth and moderate positive correlation (0.53) with illite. Smectite has a strong negative relation (-0.84) with illite. Kaolinite does not show any good relation with other clay minerals. Chlorite shows moderate (0.51) and good positive correlation (0.69) with the ratios of chlorite to illite and kaolinite to chlorite. Illite shows a moderate negative correlation (-0.52) with the ratio of chlorite to illite. However, illite does not show any relation with chlorite. Ratios chlorite to illite and kaolinite to chlorite shows a moderate positive correlation of 0.51.

6.5 Physical Properties

The physical properties of sediments discussed in this section comprise of wet bulk density, moisture content and organic matter. Wet bulk density defined as the mass of the water-sediment mixture per unit volume is an important parameter to understand the characters of the coastal sediments. The wet bulk density of the borehole samples ranges between 1.39 - 1.92 g/cc and have an average value of 1.64 g/cc. The highest value is at a depth of 16.85 – 17.0 m and the lowest value is at a depth of 6.46 - 6.50 m. The pattern of wet bulk density variation is exactly like a mirror image i.e. opposite pattern of the moisture content variation which is evident from the figure 6.2.

In the borehole the moisture content varies from 24 – 98 % and has an average value of 55 %. The highest value is at a depth of 6.46-6.50 m and the lowest value is at a depth of 16.85 – 17.0 m. Thus the area of low wet bulk density corresponds to area of high moisture content and vice versa.

The percentage of organic carbon varies from 0.1 - 7.31 and has an average value of 3.89 % in the borehole. Up to 6m depth the percentage of organic carbon is less than 1 and has the least value (0.1 %) at a depth of 1.95 – 2.0m where the sand percentage is 95. The highest value is at a depth of 38.40 - 38.50m where the sand percentage is 22. From 6 to 6.5 m depth the organic carbon ranges from 4.9 to 5.4 %. Beyond 19 m depth the organic carbon value ranges from 4.4 to 7.3%.

6.6 Geotechnical Parameters

Generally, the physico-chemistry of clay water system and organic carbon content of sediments influence the Atterberg limits (Rosenquist, 1962; Sodorblom, 1966; Rashid & Brown, 1975; Benett et al., 1981; Narayan et al., 2008).

The liquid limit (wL) values ranges from 46 – 108 % with an average value of 77 %. It shows higher values than the corresponding moisture content in the sediments except for two samples indicating that the sediments are in their plastic state. The liquid limit values were plotted against plasticity index values (Fig 6.7) as plasticity chart to understand the plasticity level. The points are seen to fall on and around A-line in the plasticity chart. Liquid limit has positive correlation with the moisture content, silt, clay, plastic limit, plasticity index and shear strength and have negative correlation with wet bulk density and sand. Liquid limit is highest at a depth of 19.75-19.95 m where the sand ranges from 3 - 4% and silt and clay ranges from 46 – 50%. The Plastic limit (wP) value ranges from 17 – 53 % and has an average value of 34 %. It is lesser than the corresponding moisture content in the sediments. This confirms that the sediments are in their plastic state.

The moisture content at which the soil can just be rolled into a 3mm diameter thread without crumbling is defined as the plasticity index (Dias and Alves, 2009). The plasticity index of the borehole samples ranges from 23 – 61% with an average value of 43%. Hence, the samples are showing medium plasticity to extremely high plasticity (Fig 6.6) and most of the samples have very high plasticity.

The activity chart is shown in figure 6.8 (clay % versus plasticity index). The figure shows Skemptoris (1953) classification of clays as 'active', 'normal', and 'inactive clays'. The activity of the samples ranges from 0.78 to 2.77 with an average value of 1.41 which says that most of the samples are active.

Rominger and Rutledge (1952) and Means and Parcher (1963) consider that the liquidity index provides a reliable indication of the degree of consolidation of clayey soils. Means and Parcher (1963) reported that $I_L = 0$ for over consolidated clays and the value of the liquidity index is negative for highly over consolidated clays (fig. 6.6). The liquidity index of the borehole samples ranges from 0.36 – 1.73 and has an average value of 0.69. So most of the samples are having I_L value in the range of 0.50 – 1.00 which means that the samples are cohesive soils or they are normally consolidated clays (after Means and Parcher 1963). Prakash and Sridharan (2004) indicated that the ratio of wP/wL remained the same for a particular type of clay mineral. Prakash and Sridharan (2004) reported the values of the ratio of wP/wL for some typical clay minerals, as summarized below:

Clay mineral	wP/wL
Na-bentonite	0.07–0.26
Illite	0.31–0.51
Kaolinite	0.36–0.70
Attapulgite	0.54

The Paravur borehole samples have a wP/wL ratio ranges from 0.25 to 0.52 and has an average value of 0.44 which accounts the dominance of illite. The clay minerals identified in the borehole samples are illite, smectite, kaolinite and chlorite (Table 6.3). Their relative abundance in the samples is shown in the figure 6.4. The illite percentage ranges from 5 to 68% with an average value of 39%. The mineral has an increasing trend with depth. The smectite percentage ranges from 10 to 71 and has an average value of 34. The mineral shows a decreasing trend with depth. The kaolinite percentage ranges from 4 to 50 with an average value of 20%. The chlorite values ranges from 0 to 16% with an average value of 6%. Smectite has a negative correlation with illite as shown in the table 6.3.

The shear strength ranges from 3 to 55 kPa with an average of 19 kPa. The highest value of shear strength is at a depth of 17.9 – 18.0 m. From 16.85 to 18.0 m, the shear strength value ranges from 54 to 55 kPa where the sand percentage ranges from 78 – 86 %. The lowest value of shear strength is at a depth of 8.7 – 10.3 m. Thus a high value shear strength result suggests that sediments are of high shear strength. Shear strength shows a negative correlation with sand. Sand

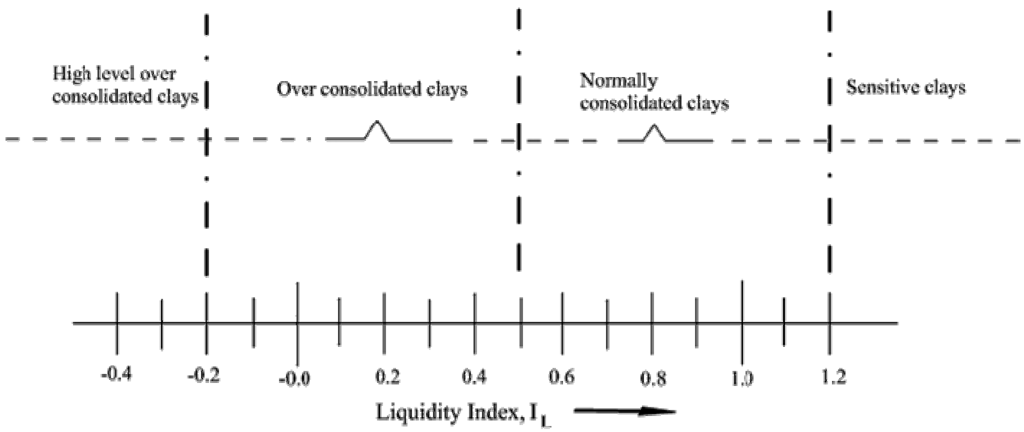


Figure 6.6: Relationship between liquidity index and degree of consolidation (after Means and Parcher, 1963)

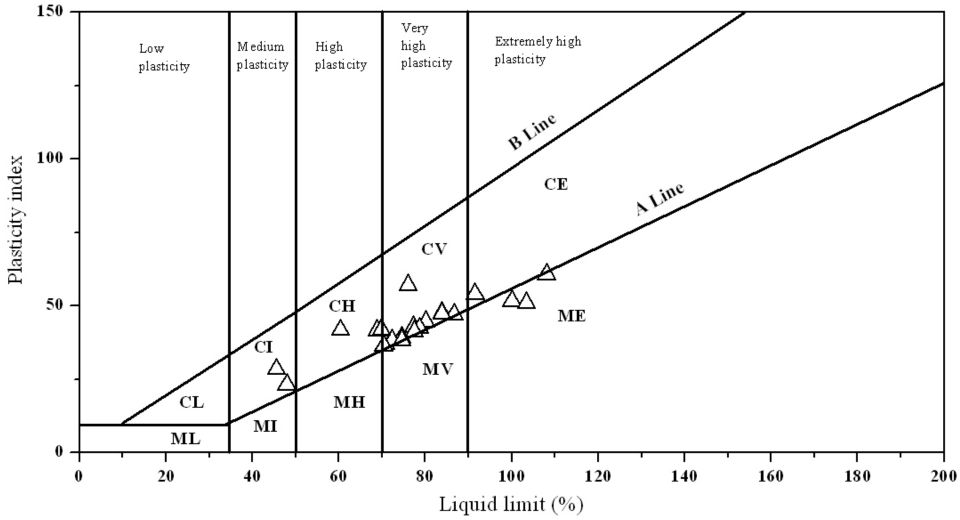


Figure 6.7: Distribution of sediment samples of Paravur borehole on plasticity chart (after Casagrande, 1948)

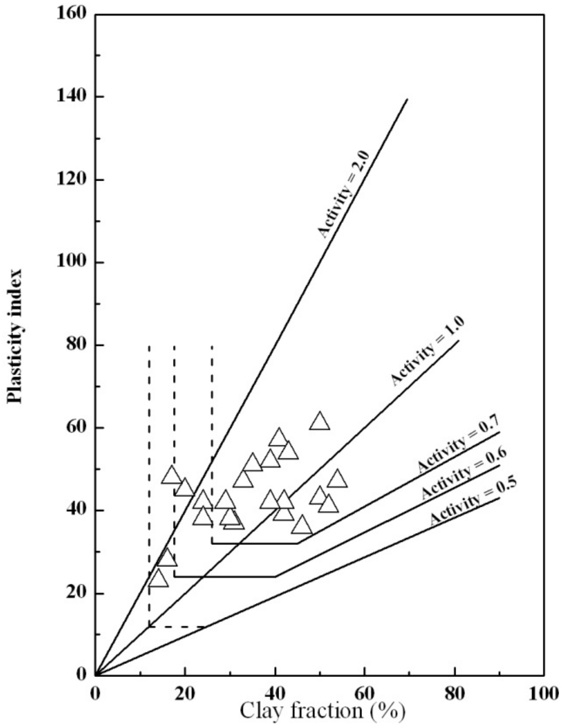


Figure 6.8: Activity chart showing the behaviour of sediment samples of Paravur borehole

Table 6.6: Correlation Matrix of the textural and geotechnical parameters of Paravur borehole having a depth of 40m (n = 22).

	BD (m)	γ (g/cc)	w (%)	C_{org} (%)	S (%)	Z (%)	C (%)	Su (kPa)	LL (%)	PL (%)	PI	A	(wP/ wL)
BD (m)	1.00												
γ (g/cc)	0.01	1.00											
w (%)	-0.16	**-0.89**	1.00										
C_{org} (%)	**0.58**	**-0.61**	**0.53**	1.00									
S (%)	-0.40	0.47	-0.29	**-0.72**	1.00								
Z (%)	0.44	-0.44	0.28	**0.60**	**-0.91**	1.00							
C (%)	0.31	-0.44	0.26	**0.72**	**-0.95**	**0.73**	1.00						
Su (kPa)	0.18	-0.37	0.21	0.42	**-0.68**	**0.63**	**0.64**	1.00					
LL (%)	-0.10	**-0.68**	**0.62**	0.42	**-0.60**	**0.62**	**0.52**	**0.61**	1.00				
PL(%)	0.19	**-0.53**	0.45	0.42	**-0.56**	**0.69**	0.39	**0.66**	**0.87**	1.00			
PI	-0.36	**-0.65**	**0.63**	0.31	-0.48	0.37	**0.51**	0.40	**0.86**	**0.50**	1.00		
A	**-0.58**	0.13	0.09	**-0.63**	**0.68**	-0.49	**-0.74**	-0.48	0.01	-0.07	0.09	1.00	
(wP/wL)	-0.34	**0.51**	-0.30	**-0.66**	**0.71**	**-0.59**	**-0.71**	**-0.81**	-0.48	**-0.50**	-0.33	**0.64**	1.00

γ : wet bulk density, w: water content, C_{org} organic carbon, S:sand, Z:silt, C: clay, Su: undrained vane shear strength, LL: liquid limit, PL: plastic limit, PI: plasticity index, A activity, wP/wL: plastic limit/ liquid limit.

percentage, in general, shows a decreasing trend with depth. The upper few meters are predominantly sandy but have lower shear strength because of under consolidation. With depth sediments are more compacted and thus have higher shear strength values even in silty and clayey regions.

6.7 Interrelationship of the Textural, Physical, and Geotechnical parameters

Physical and Textural Interrelationship

Parameters depth, wet bulk density, moisture content, organic carbon, sand, silt and clay have a good correlation except for moisture content, with depth (table 6.5). Organic carbon, silt and clay show a positive correlation with depth of burial whereas wet bulk density and sand shows a negative correlation. Hence, the percentage of organic carbon, silt and clay increases with the depth of burial whereas the wet bulk density and sand percentage decreases with the depth of burial. Wet bulk density shows negative correlation with depth, moisture content, organic carbon, silt, and clay where as a positive correlation with sand.

Moisture content has positive correlation with organic carbon, silt and clay and a negative correlation with sand. Organic carbon is having a positive correlation with depth, silt and clay and negative correlation with sand. Thus, the percentage of organic carbon decreases with increase in sand percentage or vice versa in the case of moisture content, silt and clay percentages.

Textural & Geotechnical Interrelationship

Correlation matrix of the textural and geotechnical parameters of the borehole samples is shown in Table 6.6. The borehole depth has an equal and moderate positive (0.58) and negative (-0.58) correlation with organic carbon and activity respectively. Wet bulk density shows a very good negative correlation (-0.89) with moisture content in the sediments. Similarly, wet bulk density has a good negative correlation (-0.61) with organic carbon, liquid limit (-0.68), and plasticity index. Moderate negative and positive relation with plastic limit (-0.53) and the ratio of plastic limit to liquid limit (0.51) is shown by wet bulk density. Moisture content has a good positive relation with liquid limit (0.62) and plasticity index (0.63) and moderate positive relation with organic carbon (0.53).

Organic carbon shows a negative correlation with sand and equal positive correlation with clay content of the sediments. However, a good positive relation of organic carbon is observed with silt (0.60) and good negative correlation with activity (-0.63) and the ratio of plastic limit to liquid limit (-0.66).

Sand shows a strong negative correlation with silt (-0.91) and clay (-0.95); good negative correlation with undrained vane shear strength (-0.68) and liquid limit (-0.60); and a moderate negative correlation with plastic limit (-0.56). However, sand % goes good and positive with activity (0.68) and the ratio of plastic limit to the liquid limit (0.71). Silt has good positive relation with organic carbon (0.73), undrained vane shear strength (0.63), liquid limit (0.62), and plastic limit (0.69). Silt also has moderate negative correlation with the ratio of plastic limit to the liquid limit. Clay shows a good negative behaviour with activity (-0.74) and the ratio of plastic limit to the liquid limit (-0.71). Clay has moderate positive relation with liquid limit (0.52) and plasticity index (0.51) and good positive relation with undrained vane shear strength (0.64).

Undrained vane shear strength shows a very good negative variation with the ratio of plastic limit to liquid limit and good positive relation with liquid limit (0.61) and plastic limit (0.66). Liquid limit shows a very good positive variation with plastic limit (0.87) and plasticity index (0.86). Plastic limit shows an equal and opposite relation with plasticity index (0.50) and the ratio of plastic limit to liquid limit (-0.50). Activity also shows a good positive variation with the ratio of plastic limit and liquid limit (0.64).

Interrelationship between organic carbon versus liquid limit (0.43), plastic limit (0.32), plasticity index (0.44), and shear strength (0.11) of the Paravur borehole samples is shown in (Fig 6.9) which does not show any good correlation between these parameters.

6.8 Pollen Records

Sediment and peat samples from the same borehole were recovered and pollen analysis was done to understand the environmental changes during the Late Quaternary. Peat deposits of Middle to Late Holocene age occur along the coastal tracts at varied subsurface levels. Occurrence of peat deposits at subsurface levels reveals that mangrove vegetation was predominantly present along the coastal tracts in the Late Quaternary period, subsequently inundated by the higher sea levels, and led to the formation of peat deposits.

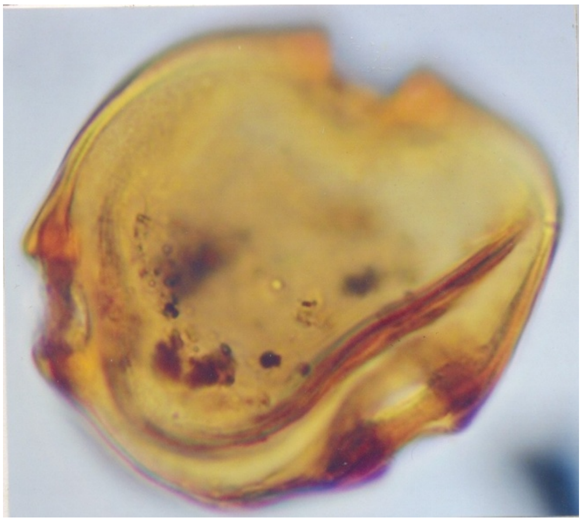
Pollen analysis reveals abundant occurrence of Rhizophoraccae pollen (mangrove). Occurrence of other plant taxa assignable to Combretaccae, Anacardiaccae, etc., in low to moderate values, indicates existence of peripheral mangroves (Plates 1 to 18).

Dinoflagellates are unicellular protists that produce a cyst (dinocyst) as part of their life cycle. The living cells of dinoflagellates do not yield remains that fossilize. However, during the course of their life cycle some species produce resistant organic-walled cysts. The cyst permits the survival of cells for periods of variable lengths (from the season to decades). It is composed of refractory organic matter that permits preservation in the sediment and fossilization. The fossil form is usually referred to as dinocyst. They are usually well preserved in sediments of continental margins including epicontinental seas, estuaries, continental shelves, slopes and rises. They are widely used as proxy indicators of marine conditions and provide valuable information on the natural variability of climate (Vernal and Rochon, 2011). About 20% of dinoflagellates produce dinocysts that fossilize as palynomorphs (Dale, 1976).

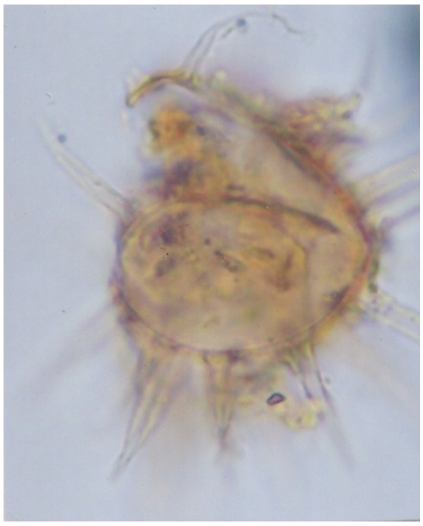
Occurrence of dinocyst and foraminifera in the peat samples reveals a period of marine transgression in the region. The major midland taxa are Oleaccae, Fabaccae (legumes) and Meliaccae whereas Poaccae, Lamiaccae, Urticaccae Asteraccae, Apiaccae and Chenopods, being ubiquitous, are present in low to moderate numbers. Plant derived organic matter, with fungal and pteridophytic remains, was also recorded in the collected sediments. The samples represent important sub-environment of the delta during seasonal hydro-periods that indicate variable influence of the brackish water – fresh water revealed by the deposition of various palynomorphs. Humid climate accounts for the development of mangrove vegetation, as the abundant rainfall increases the continental drainage.

Narayana (2007) reported the pollen analysis of peat samples (>40,000 yr BP) showing 40% pollen assemblages belonging to the family *Rhizoporaceae* and over 60% pollen

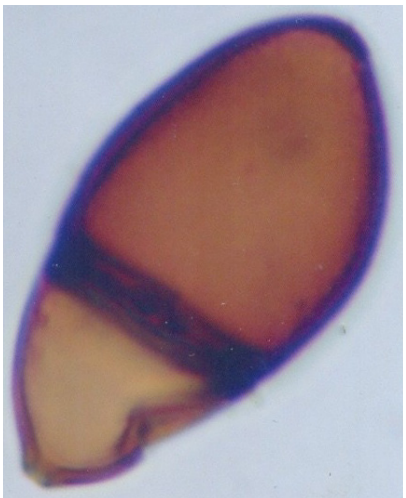
The following plates show the pollen records obtained at different levels in the Paravur borehole.



Tiliaceae x 2500



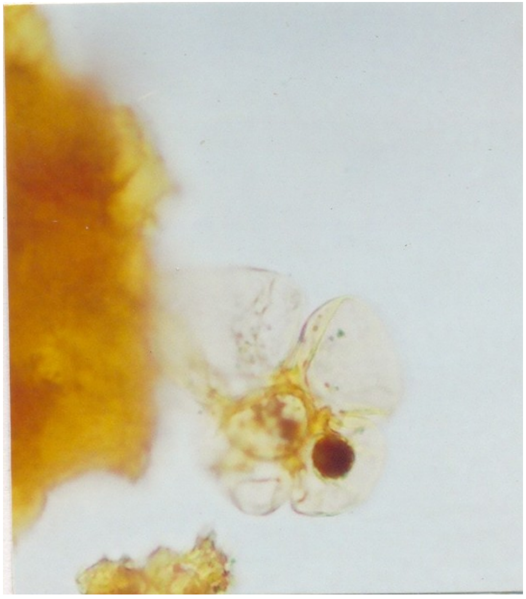
Dinocyst x 1500



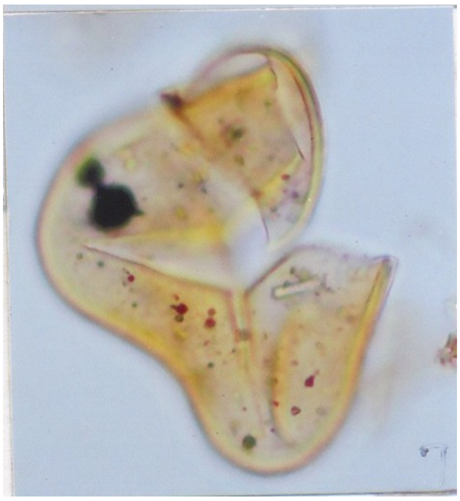
Fungal spore x 1500

Poaceae x 1500

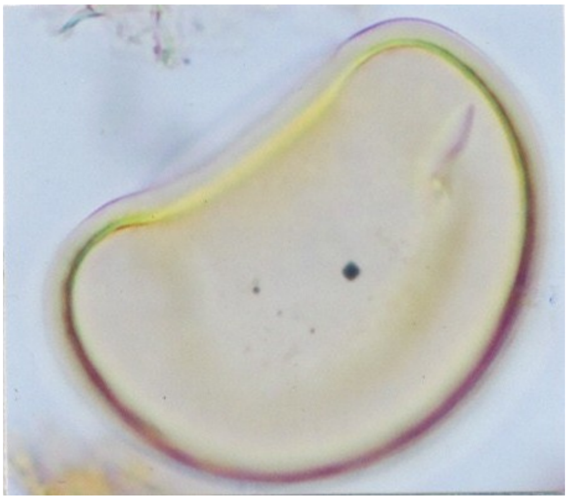
Fruiting body x 1500



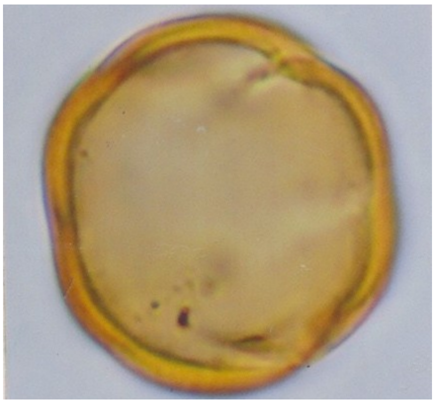
Foraminifera x 1000



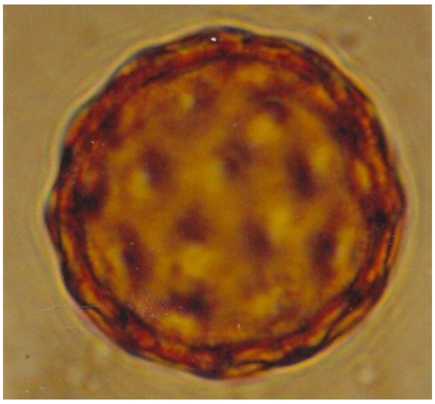
Trilete fern spore x 1000



Monolete fern spore x 1000



Terminalia **sp. (Combretaceae) x 1500**



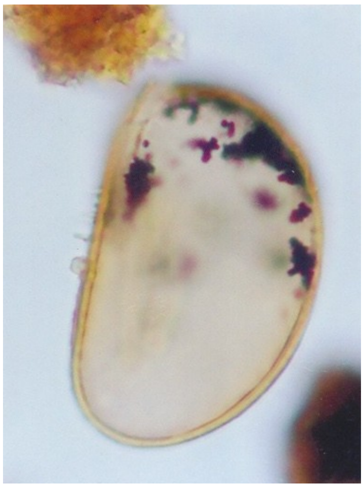
Chenopodiaceae x 1200

Rhizophora **sp. (Rhizophoraceae) x 1500**

Fungal spore x 1000

Fabaceae x 1000

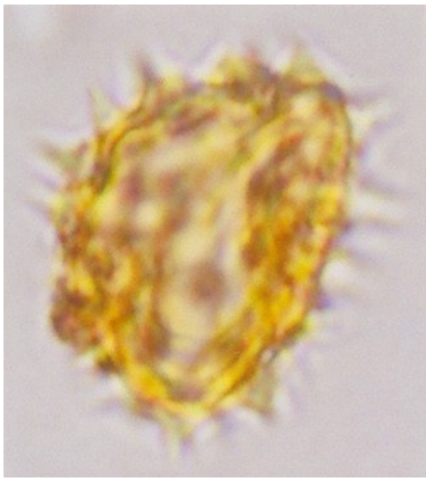
Anacardiaceae x 1500



Monolete fern spore x 1000

***Plantogo* sp. x 1500**

***Emblica* sp. (Euphorbiaceae) x 1500**



Asteraceae x 1200

corresponding to *Mallotus*. This revealed the paleo-mangrove-environment dominated by the presence of evergreen to semi-evergreen types of mangrove plants in the west coast of India during Late Quaternary

6.9 Discussion: Sedimentation History

^{14}C ages of peat deposits of the borehole sediments gives the mid to early Holocene ages. However, the upper portions which are confined up to 6 m from top may be of late Holocene age but can't be assured as we lack the age informations. The older ages at the depths 15.31-15.35 m (12116 yrs BP), and 22.65-22.95 m (10330 yrs BP) than at the depth 24.40-24.50 m (9550-9330 yrs BP) indicates the reworking of the sediments. This reworking of the sediments is possible tectonic influences in the coastal region of central Kerala. The radiocarbon dates from the surrounding regions are reported by many works. Rajendran et al., (1989) reports the age of the 2 m peats as 7230 yrs BP. Shajan (1998) reports 7450 yrs BP peats from Poovathumkadu (Cochin). Periyar river mouth peats at 24 m depth in the sediment column has been reported by Narayana (2007) as 8460 yrs BP. Kale and Rajaguru (1983) suggests that the sea level attained its maximum landward level during mid-Holocene, and this was followed by a lowering and stabilization at the present level.

Coastal environments can be used as an archive for understanding both erosional and depositional features preserved in the sediments. This offers an opportunity to understand the sedimentary processes and tectonics (Chen et al., 2003). Padmalal et al. (2011) says about the geomorphic and geologic evolution of coastal lands which has been influenced by changes in climate, sea level, and local and regional tectonics. The deposition of alluvium in preexisting channels in the coastal zones of Kerala concluded the deltaic sequence in the river confluence zone due to sea level rise during Late-Pleistocene and Early Holocene (Padmalal et al., 2011). Textural nomenclature after Folk (1980) classifies the total borehole length in to six litho facies as: Muddy Sand, Sand, Sandy Mud, Silty Sand, Sandy Silt, and Mud. Knowledge of the various characteristics of sediment permits the assessment of its state of evolution, its maturity and also a better understanding of its history. Chamley (1990) states that mature sediments with narrow granulometric range may be used as evidence for active and prolonged hydrodynamic processes such as littoral or desert dunes, beaches, and other shallow-marine exposed environments. Hence, sediments of the borehole shows creaseless hydrodynamic action, fast and strong, as encountered in the central to distal zone of the deltaic facies. However, the grain size shows an increasing trend in the upper portion of the borehole. This indicates the change of the sediment deposition towards the rough hydrodynamic actions by fluvial facies. Hence, textural studies of the borehole sediments show a conversion of depositional environment from distal zone to the central zone of deltaic facies. This may be inferred as the rise in sea level during early-Holocene. Padmalal et al. (2011) also reports evidences of many marine transgression and regression in the region. However, an alternate shift in textures has been observed below the depth of 18.82-18.92 m indicating the alternate low and high weathering in the source regions respectively with variation in grains from mud to sandy mud or muddy sand. An upward coarsening of the sediments can also indicate the increasing trend of weathering in the adjacent areas during sea level rise. Wang et al. (2010) also reports post glacial evolution of the subaqueous Yangtza delta of China linked to the fluctuation of sea level and the effect of East Asian monsoon. He also reports the high sedimentation rate and coarse grain size of estuarine deposition during 8.4-11.4 kyrs BP to a rapid sea level rise and the warm and wet climatic condition of the East Asian monsoon. The surficial and subsurficial zone is blanketed with muddy sand and sand indicating the proximal zone deposition of the sediments mainly by fluvial actions.

Clay minerals can be used as a potential tool to study geology and geomorphology of the source region since it indicates the intensity of weathering, especially the degree of hydrolysis, at source rock regions (Chamley, 1989, Thamban et al., 2002). However, size sorting during transportation (Gibbs, 1977), organic interaction of clays (Degens and Ittekot, 1984, Thamban and Rao, 2002), flocculation at lower salinities (Grim, 1968), redistribution and reworking of clay minerals should be understood before the diagenetic studies for the provenance. In this work, clays from the mid to early Holocene age in the borehole sediments were used to understand the adjacent terrestrial environment. The major clay mineral groups present in the borehole sediments are: illite, chlorite, kaolinite, and smectite. The temporal relative abundances (Fig 6.4) of clay minerals have varied significantly. Illite and smectite are the dominant clay minerals in the borehole showing opposite trends. Smectite shows higher abundance in upper sediments from surface to 11.60 m depth after which illite is dominating throughout. Kaolinite also has a similar trend like smectite whereas chlorite is very low throughout the borehole. Thamban et al (2003) gives the informations of modern distribution of the clay minerals along the western continental margin of India by pioneer workers as illite and chlorites are originated by the physical weathering of Precambrian terrains in the Himalayas (Kolla et al., 1981; Nair et al., 1982; Rao ad Rao, 1995; Kessarkar et al., 2003) and are getting deposited in the Arabian sea by Indus river. Weathering of Deccan trap basalts under a semi-arid climate is the source of smectite and are carried out mainly by Narmada and Tapti river systems (Subramanian, 1980; Naidu et al., 1985; Rao and Rao, 1995; Kessarkar et al., 2003). Precambrian gneissic rocks and laterites of Western Ghats are the main source for the kaolinite deposits in the west coast. Chlorite is reported to be a product of physical weathering under an arid and cold climate (Biscaye, 1965; Weaver, 1989). However, the influence of the Indus river system to the study is not clear; the source for the illite could also be the weathering of Deccan trap basalts in-arid climate conditions. Hence, the source regions for the clay mineral deposition in the central Kerala coast are Deccan trap basalts for illite, chlorite and smectite gneissic rocks and laterites of the adjacent terrestrial areas for kaolinite. Temporal variation in the relative percentage of the clay mineral groups have shown a distinct fluctuation at the depths of 1.95-2.0 m (wet to dry), 11.55-11.60 m (dry to wet), 16.85-17.0 m (dry to wet), and 18.82-18.92 m (wet to dry) respectively in climate.

The geotechnical properties of the sediments – high moisture content, high organic carbon, high plasticity index, and low bulk density-are generally associated with a low critical shear stress for erosion (Smerdon and Beasley, 1961; Vanoni, 1975; Thorn and Parsons, 1980) and increased erosion rate (Mehta et al., 1982). Figure 6.2 shows an inverse relation of wet bulk density and moisture with low wet bulk density. Moisture also shows a similar variation with high organic carbon content downward. Hence, the sediments indicate an increased erosion rate in time. Organic carbon shows increasing trend with moisture content due to water absorbing nature of the organic matter. Similarly this can also be taken as an indication of hydrogel formation amongst organic matter (Louda et al., 2004). Organic carbon also shows an inverse relation with wet bulk density supporting the high Atterberg limits (cf. Keller, 1982). Sand shows an inverse trend with silt and clay and hence, negative correlation with liquid and plastic limit (Fig.6.2). High concentration of silt and clays in the lower portion and higher concentration of sand in the upper portion indicates the higher cohesion of the sediments in the past and relatively lower cohesive nature of the sediments of the recent times which may have caused the recent sediments more susceptible to resuspension.

The pollen analysis of peat samples reveals abundant occurrence of Rhizophoraceae pollen (core mangrove). Occurrence of other plant taxa assignable to Combretaceae, Anacardiaceae, etc., in low to moderate values, indicates existence of peripheral mangroves. The major midland taxa are Oleaceae, Asteraceae, Apiaceae, and Chenopods, being, ubiquitous, are present in low to moderate number. Plant derived organic matter, with fungal and pteridophytic remains, was also recorded in the collected sediments. Dinocyst are usually well preserved in sediments of continental margins including epicontinental seas, estuaries, continental shelves, slopes and rises. They are widely used as proxy indicators of marine conditions and provide valuable information on the natural variability of climate (Vernal and Rochon, 2011). About 20% of dinoflagellates produce dinocysts that fossilize as palynomorphs (Dale, 1976). The samples represent important sub-environment of the delta during seasonal hydro-periods that indicate variable influence of brackish water - fresh water revealed by the deposition of various palynomorphs. Humid climate accounts for the development of mangrove vegetation, as the abundant rainfall increase in the continental drainage. Narayana (2007) reported the pollen analysis of peat samples (>40,000 yr BP) showing 40% pollen assemblages belonging to the family *Rhizoporaceae* and over 60% pollen corresponding to *Mallotus.* This revealed the paleo-mangrove-environment dominated by the presence of evergreen to semi-evergreen types of mangrove plants in the west coast of India during Late Quaternary. Hence, the pollen record of the sediments supports the paleo-mangrove-environment dominated by presence of semi-evergreen types of mangrove plants in mid and early Holocene times also.

****

Chapter 7

SUMMARY AND CONCLUSION

The study deals with the sedimentological and geotechnical aspects of central Kerala coast. The main focus was given to different depositional environments in the coastal regions such as lagoonal sediments, continental shelf sediments (up to 30 m water depth) and paleo-deltaic onshore sediments recovered from a bore hole to a depth of 40 m which is approximately 6 km away from the present coast line. Clay mineralogy, sedimentological and geotechnical studies such as texture, moisture content, wet bulk density, organic carbon, Atterberg limits and vane shear were done for the lagoonal sediments. Trace element analysis was done in addition to the clay mineralogy, sedimentological and geotechnical studies for selected core samples of the nearshore continental shelf sediments. ^{14}C dating and palynological studies were done in addition to the clay mineralogy, sedimentological and geotechnical studies to unravel the geochronology, depositional history, sea level variation during the Late Quaternary and sediment provenance in the paleo- delta sediments.

The general distribution pattern of sediment texture across the Vembanad lagoon has been discussed based on the variation of sand-silt-clay ratios of the sediment cores. This gives the idea of change in the textural pattern of the sediments which can be concluded as the change in the environment of deposition. The sedimentary lithofacies present in lagoonal beds are controlled by the hydrodynamic processes as well as the nature of the availability of the sediments. Vertical aeration of the sediments is mediated by the rate and nature of the sediments supply. Sediment supply could be either in-situ or transported from sea or land. Quiet water deposition of silt and clay occurs in lagoons in a manner analogous to pro-delta environments, where sediments are carried by rivers. As indicated in the textural variations of the lagoonal sediments, the system represents a high energy environment. This is characterized by sand - muddy sand - sandy mud facies. Core PE-33A represents a change in the energy conditions from low to high when moved upwards in the core. The grab samples also show a similar high energy environment by sandy mud textures. However, muddy texture representing low energy environment have been shown by S-2, S-3, S-31, and S-32 grab samples. Nearshore samples broadly show a low energy environment of deposition represented by muddy textures. However, G-6 core has sandy mud texture in the surface sediments. Core G-9 shows a change in textural pattern downcore from muddy sand to clayey sand. Similarly, Core GC-20 also shows a downcore textural variation as sandy mud – mud - sandy mud. Knowledge of the various characteristics of sediment permits the assessment of its state of evolution, its maturity and also a better understanding of its history.

Textural nomenclature after Folk (1980) classifies the total borehole length in to six litho facies as: Muddy Sand, Sand, Sandy Mud, Silty Sand, Sandy Silt, and Mud. Sediments of the borehole shows creaseless hydrodynamic action, fast and strong, as encountered in the central to distal zone of the deltaic facies. However, the grain size shows an increasing trend in the upper portion of the borehole. This indicates the change of the sediment deposition towards the rough hydrodynamic actions by fluvial facies. Hence, textural studies of the borehole sediments show a conversion of depositional environment from distal zone to the central zone of deltaic facies. This may be inferred as the rise in sea level during early-Holocene. The surficial and subsurficial zone is blanketed with muddy sand and sand indicating the proximal zone deposition of the sediments mainly by fluvial actions. High concentration of silt and clays in the lower portion and higher concentration of sand in the upper portion indicates the higher cohesion of the sediments in the past and relatively lower cohesive nature of the sediments of the recent times which may have caused the recent sediments more susceptible to resuspension.

The source of the organic matter is dominated by autochthonous material from net primary production and allochthonous material supplied by rivers, or both together. The nearshore processes such as wind, waves and currents play an important role in removing the fine sediments near the shore area which contains high organic matter. Primary productivity enhances due to nutrient loadings and from discharges of industrial plants and sewage treatment works and from other human influenced 'diffuse sources' such as run-off from agricultural catchments that may increase the water eutrophication effects. In general, it can be said that in an estuary the organic matter is formed either by photosynthesis or transported as particulate or dissolved matter into the estuary by the rivers. Dissolved humic substances, which are abundant in the river water flocculate together with the fine grained sediments as they meet the saline estuarine waters. Nearshore zone also receives organic matter dominantly from terrestrial source and to a lesser degree from primary productivity. The bore hole sediments were deposited in a fluvial environment and shows high organic carbon percentage with depth. This supports the changing depositional environment from a high energy central zone of deltaic facies to the low energy distal zone.

An inverse relation of wet bulk density and moisture content exist in most of the core samples. Moisture also shows variability with high organic carbon % downward in different cores indicating the varying erosional rate temporally. Organic carbon % shows increasing trend with moisture content and clay percentage. Clay has the water absorbing nature and acts as a sink of organic carbon. Similarly this can also be taken as an indication of hydrogel formation amongst organic matter as opined by Louda et al. (2004). Organic carbon also shows an inverse relation with wet bulk density in the cores supporting the high Atterberg limits as earlier stated by Keller, (1982). Sand also shows variability in its trend with silt and clay and hence, variable correlation with liquid and plastic limit. Lower concentration of silt and clays and the higher concentration of sand in the estuary cores indicate the lower cohesion of the sediments showing that the sediments are more susceptible to resuspension. Liquid limit, plastic limit and plasticity index of the core sediments are showing positive trend with silt and clay percentages. The surface sediments of the cores are having higher moisture content than the corresponding liquid limit. Therefore the surface sediments are in the liquid state. As depth increases the moisture content of the sediments shows lower values than the corresponding liquid limits. Thus the sediments are in the plastic state with increasing depth. Based on the Cassagrande's (1948) plasticity chart the sediments in the nearshore environment are lying above and below the A-line and are high plasticity to extremely high plasticity clay and mud. The activity chart of the cores (clay % versus plasticity index) showing Skemptoris (1953) classification of clays as 'active', 'normal', and 'inactive clays', represents that most of the samples are active in nature. The surface sediments of the cores are showing very low shear strength values and exhibit an increasing trend with depth. According to Means and Parcher (1963) classification of clayey soil based on the degree of consolidation it can be concluded that most of the core sediments at depth are normally consolidated clays.

The borehole samples are showing medium plasticity to extremely high plasticity and most of the samples have very high plasticity. The activity of the samples has an average value of 1.41 which says that most of the samples are active. The Paravur borehole samples have a wP/wL ratio ranges from 0.25 to 0.52 and has an average value of 0.44 which accounts the dominance of illite. High value shear strength result suggests that sediments are of high shear strength. Shear strength shows a negative correlation with sand. Sand percentage, in general, shows a decreasing trend with depth. With depth sediments are more compacted and thus have higher shear strength values even in silty and clayey regions.

Illite and kaolinite flocculate at lower salinities than smectite and get deposited near the river mouth stations whereas smectite will be deposited further down the estuaries. This can be attributed to the high illite percentage in the Vembanad lagoon core recovered from a location near to the river mouth. Abundant smectite in the other core sediments indicate the low energy environment in the lagoonal system. Thus the clay minerals present in the Vembanad lagoon is transported by the rivers from the weathering of Precambrian gneisses and laterites and are differentially flocculated based on the salinity. The relative concentration of clay minerals follows the following trend in most of the cores from the nearshore environments: Smectite>Kaolinite>Illite>Chlorite. The major clay mineral groups present in the borehole sediments are: illite, smectite, kaolinite, and chlorite. The temporal relative abundances of clay minerals have varied significantly in the borehole. Illite and smectite are the dominant clay minerals in the borehole showing opposite trends. The source regions for the clay mineral deposition in the central Kerala coast are Deccan trap basalts for illite, chlorite and smectite; gneissic rocks and laterites of the adjacent terrestrial areas for kaolinite. Here, Precambrian gneissic rocks and laterites of Western Ghats are the main source for the kaolinite in the sediments. The detrital chlorite mainly results from chemical weathering of plutonic and metamorphic rocks.

Trace metal variations in the three nearshore cores indicates that the source rock could range from mafic to acidic rocks which are present (Charnockite, Granitic Gneiss) in the surrounding regions. Anthropogenic factor has also enhanced to a very high degree for the trace metal concentration in the sediments. The enrichment factor (EF) reflects the anthropogenic input of the contaminant elements in estuarine or marine sediments. The nearshore sediments show an enrichment of nickel, copper, zinc and lead. Copper shows high enrichment in all the core sediments. There is a spatial increase in the lead enrichment with decreasing water depths. Enrichment in lead concentration is observed at the surface levels of all the three cores. From the Pollution load index (PLI) values it can be concluded that the region is moderately to highly polluted. The Pollution index (PLI > 1) is increasing spatially from G8, G10, and G11 core respectively i.e. with decreasing water depths. The geo-accumulation index reveals that the sediments are uncontaminated – moderately contaminated to strongly – extremely contaminated with decreasing water depths. Sudden increase (at the depth interval of 44-46 cm onwards) in the trace metals like lead, zinc, and nickel in G11 core reveals the anthropogenic contribution in polluting the area. Based on the temporal distribution of trace metals, it is inferred that accumulation of Pb, Zn and Ni in the recent past has increased in the coastal sediments. This is supported by the lesser concentration in the bottom sediment of the cores. The possible reason for this enhancement could be due to ship transportation in the region. In addition the three cores were recovered from locations near to the Munambam fishing harbor and the navigation of the fishing boats also contribute to the increased trace metal concentration. Hence, it can be inferred that not only the natural sources but also industrial effluents, domestic sewage, navigation, fishing and rapid urbanization are the contributors of trace metals which is becoming a major threat for the coastal region in central Kerala.

Radiocarbon ages of the sediments ranging from 11233yrs BP to 4692 yrs BP covers mid to early Holocene time. However, surficial sediments could be of late Holocene. Discontinuity in radiocarbon ages represents reworking of the sediments due to tectonic influences in the coastal region of central Kerala.

Pollen analysis reveals abundant occurrence of Rhizophoraceae pollen (mangrove). Occurrence of other plant taxa assignable to Combretaceae, Anacardiaceae, etc., in low to moderate values, indicates existence of peripheral mangroves. Occurrence of dinocyst and foraminifera in the peat samples reveals a period of marine transgression in the region. The major midland taxa are Oleaceae, Fabaceae (legumes) and Meliaceae whereas Poaceae, Lamiaceae, Urticaceae Asteraceae, Apiaceae and Chenopods, being ubiquitous, are present in low to moderate numbers. The samples represent important sub-environment of the delta during seasonal hydro-periods that indicate variable influence of the brackish water – fresh water revealed by the deposition of various palynomorphs. Humid climate accounts for the development of mangrove vegetation, as the abundant rainfall increases the continental drainage.

****

REFERENCES

Aaby, B. and Jacobsen, J., Changes in biotic conditions and metal deposition in the last millennium as reflected in ombrotrophic peat in Draved Moses, Denmark. Geol. Surv. Denmark, Year book., 1978. 5.

Abdullah, N. A. and Ranjini, A./ Munisamy, P. Mineralogy in Sediments at Kertih Lagoon, Terengganu, Malaysia. *UMT 11th International Annual Symposium on Sustainability Science and Management 09th – 11th July 2012, Terengganu, Malaysia.*

Abrahim, G.M.S. and Parker, R.J. (2008). Assessment of Heavy Metal Enrichment Factors and the Degree of Contamination in Marine Sediments from Tamaki Estuary, Auckland, New Zealand. Environ Monit Assess, v. 136: pp. 227-238.

Adams, F. Interaction of phosphorous with other elements in soils and in plants, in Proc. Symp. : The Role of Phosphorus in Agriculture; Khasawneh, F. E., Ed., Am. Soc. Agron., Madison, WI, 1980, p. 655.

Agrawal, D.P.; Gupta, S.K. and Kusumgar. S. (1970). Radio carbon dates of Quaternary samples. Current Science. V. 30, pp. 219-222.

Agrawal, D.P. and Guzder, S.J. (1974). Quaternary studies on the western coast of India: Preliminary observations. The Paleobotanist. V. 21, pp. 216-222.

Ahmad, E. (1972). Coastal Geomorphology of India. Orient Longman, New Delhi. 222p.

Aidiniyan, N. Kh., (1959). Distribution of rare alkalies in colloids of soils and the participation of vegetation in this process. Geokhimiya, v. 4, p. 346.

Alberic, P., Viollier, E., Jezequel, D., Grosbois, C., Michard, G., (2000). Interactions between trace elements and dissolved organic matter in the stagnant anoxic deep layer of a meromictic lake. Limnology and Oceanography v. 45, pp. 1088-1096.

Alexander, C. R., Domaster, D. J. and Nittrouer, C. A. (1991). Sediment accumulation in a modern epicontinental-shelf setting: The Yellow Sea. Mar. Geol. V. 98., pp. 51-72.

Algeo, T.J. and Maynard, J.B. (2004). Trace element behaviour and redox facies in core shales of upper Pennsylvanian Kansas-type cyclothems. Chemical Geology, v. 206, pp. 289-318.

Algeo, T.J. (2004). Can marine anoxic events draw down the trace-element inventory of seawater? Journal of Geology, 32, 1057-1060.

Alongi, D.M. (2009). The Energetics of Mangrove Forests. Springer Verlag, Heidelberg. xi+ 216 pp.

Álvarez-Iglesias, P., Rubio, B. & Vilas, F. (2003). Pollution in intertidal sediments of San Simón Bay (Inner Ría de Vigo, NW of Spain): total heavy metal concentrations and speciation. Marine Pollution Bulletin, Vol. 46, pp. 491-506.

Anderson, A. (1977). Some aspects on the significance of heavy metals in sewage sludge and related products used as fertilizers; Swed. J. Agric. Res., V. 7(1).

Andrade , F.A., Al-Qureshi, H.A. and Hotza, D. (2011). Measuring the plasticity of clays: A review. Applied Clay Science, v. 51, pp. 1–7.

Angelidis, M.O. and Aloupi, M. (2000). Geochemical study of coastal sediments influenced by river transported pollution: Southern Evoikos Gulf, Greece. Mar. Pollu. Bull., v.40, pp.77-82.

Angulo, E. (1996). The Tomlinson pollution load index applied to heavy metal ‘ Mussel-Watch’ data: a useful index to assess coastal pollution. Science of the Total Environment. v.187, pp. 49-56.

Angulo, R.J., Gianini, P.C.F., Suguio, K., Passenda, L.C.R. (1999). Relative sea-level changes in the last 5500 years in southern Brazil (Laguna-Imbituba region, Santa Catarina State) based on vermitid ^{14}C ages. Marine Geology v. 159, pp. 323-339.

Anguloa, R. J., Lessab, G. C. and de Souzaa, M. C. (2006) A critical review of mid- to late-Holocene sea-level fluctuations on the eastern Brazilian coastline. Quaternary Science Reviews, v. 25 pp. 486–506.

Anirudhan, T.S. (1988). Studies on the nutrient chemistry of a tropical estuary: Cochin University of Science and Technology, India. Ph. D. (unpubl.) Thesis, 98 p.

Anon (1984). Resource Atlas of Kerala. Centre for Earth science Studies Publ., Thiruvananthapuram, Plates 58, 39p.

Anon (1974). Water Resources of Kerala. Government of Kerala Publ., 110p.

Anthony, E.J., Oyede, L.M. and Lang, J. (2002). Sedimentation in a fluvially infilling, barrier-bound estuary on a wave-dominated, microtidal coast: the Oueme River estuary, Benin, West Africa. Sedimentology, v.49, no.5, pp.1095-1112.

ASTM, (1994). Annual Book of ASTM Standards, Soil and Rock (I). American Society of Testing and Materials, 716 pp.

ASTM D 4318 (2000). Standard test methods for liquid limit, plastic limit and plasticity index of soils. Annual book of American Society for Testing and Material standards, West Conshohocken, United States.

ASTM E 691 (2005). Standard practice for conducting an interlaboratory study to determine the precision of a test method. Annual book of American Society for Testing and Material standards, West Conshohocken, United States.

Atterberg, A. (1911). Die Plastizitat der Tone. Internationale Mitteilungen fuer Bodenkunde 1, 10–43.

Babich H, Stotzky G. (1978). Effects of cadmium on the biota: influence of environmental factors. Adv Appl Microbiol. V. 23, pp. 55–117.

Badarudeen, A., Damodaran, K.T., Sajan, K. and Padmalal, D. (1996). Texture and geochemistry of the sediments of a tropical mangrove ecosystem, southwest coast of India. Environ. Geol., v.27, pp.164-169.

Badarudeen, A. (1997). Sedimentology and geochemistry of some selected mangrove ecosystems of Kerala, southwest coast of India. Ph.D. Thesis, Cochin University of Science and Technology, Cochin, 163p.

Bain, J. H. C. (1977) BMR J. Austral. Geol. Geophys. V. 2, pp. 137-147.

Balaram, V. (1997) Microwave dissolution techniques for the analysis of geological materials by ICP-MS. Current Science, v. 73(11), pp. 1019-23.

Balaram, V. and Rao, T.G. (2003) Rapid determination of REEs and other trace elements in geological samples by microwave acid digestion and ICP-MS. Atomic Spectra, v. 24, pp. 206-212.

Baxter, A.E., Meadows, M.E. (1999). Evidence for Holocene sea-level change at Verlorenvlei, Western Cape, South Africa. Quaternary International v. 56, pp. 65-79.

Becker, A., Bucher, F., Davenport, C. A. and Flisch, A. (2004). Geotechnical characteristics of post-glacial organic sediments in Lake Bergsee, southern Black Forest, Germany. Engineering Geology, v. 74, pp. 91–102.

Beets, D. J., Valk, L. V. and Stive M. J. F. (1992). Holocene evolution of the coast of Holland. Marine Geology, v. 103, pp. 423-443.

Benner, R. (2004). What happens to terrestrial organic matter in the ocean? Marine Chemistry, v. 92, pp. 307-310.

Benner, R., Opsahl, S. (2001). Molecular indicators of the sources and transformations of dissolved organic matter in the Mississippi River plume. Organic Geochemistry, v. 32, pp. 597-611.

Bennett, R.H., Bryant, W.R., Keller, G.H. (1981). Clay fabric of selected submarine sediments fundamental properties and models. Journal of Sedimentary Petrology, v. 51, pp. 217-232.

Berglund, B.E., Sandgren, P., Barnekow, L., Hannon, G., Jiang, H., Skog, G., Yu, S.-Y. (2005). Early Holocene history of the Baltic Sea, as reflected in coastal sediments in Blekinge, southeastern Sweden. Quaternary International, v. 130, pp. 111-139.

Berkeley, A., Perry, C.T., Smithers, S.G., Horton, B.P., Cundy, A.B. (2009). Foraminiferal biofacies across mangrove-mudflat environments at Cocoa Creek, north Queensland, Australia. Marine Geology, v. 263, pp. 64–86.

Berner, R.A. (1982). Burial of organic-carbon and pyrite sulphur in the modern ocean — its geochemical and environmental significance. American Journal of Science, v. 282, pp. 451–473.

Bianchi, T. S., Galler, J. J. and Allison, M. A. (2007). Studies on the hydrodynamic sorting and transport of terrestrially derived organic carbon in sediments of the Mississippi and Atchafalaya Rivers. Estuarine, Coastal and Shelf Science, v.73, pp. 211-222.

Bindoff, N.L., Willebrand, J., Artale, V., Cazenave, A., Gregory, J., Gulev, S., Hanawa, K., Le Quéré, C., Levitus, S., Nojiri, Y., Shum, C.K., Talley, L.D., Unnikrishnan, A., (2007). Observations: oceanic climate change and sea level. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC). Cambridge University Press, Cambridge and New York, pp. 385-432.

Birch, G.F., Taylor, S.E., Matthai, C. (2001). Small-scale spatial and temporal variance in the concentration of heavy metals in aquatic sediments: a review and some new concepts. Environ. Pollut., v. 113, pp. 357–372.

Bird, E. (2000). Coastal Geomorphology: An introduction. John Wiley & Sons Ltd. Chichester, England. 322p.

Biscaye, P.E. (1964). Distinction between chlorite and kaolinite in recent sediments by X-ray diffraction. American Mineralogist, v. 49, pp. 1281-1289.

Biscaye, P.E. (1965). Mineralogy and sedimentation in the recent deep-sea clay in the Atlantic and adjacent seas and oceans. Bulletin Geological Society of America, v. 76, pp. 803-832.

Blair, N.E., Leithold, E.L., Brackley, H.L., Trustrum, N.A., Page, M.J., Childress, L. (2010). Terrestrial Sources and export of particulate organic carbon in the Waipaoa Sedimentary System: problems, progress and processes. Marine Geology, v. 270, pp. 108–118.

Blasco, F. (1975). The mangroves of India, Travaux de la Section Scientifique et Technique. French Institute of Pondicherry, v.14, pp. 1 –125.

Blasco, F. (1984). In: Mangroves ecosystem research methods, p.18– 35 (eds. Snedaker, S.C. and Snedaker J.G.), UNESCO, Paris, pp. 36– 49.

Blasco, F., Saenger, P. and Janodet, E. (1996). Mangroves as indicators of coastal change. Catena, v. 27, pp. 167-178.

Blaser, P., Zimmermann, S. and Luster, J. (2000). Critical Examination of Trace Element Enrichments and Depletions in Soils: As, Cr, Cu, Ni, Pb, and Zn in Swiss Forest Soils. *The Science of the Total Environment*, v. 249, pp. 257–280.

Bose, R. N., Kartha, T. D. G., and Padmanabhan, K. (1976). Geophysical investigations in the coastal basin of Kerala. Paper presented to the Workshop on Coastal Sedimentaries of India south of 18°N *ONGC, Madras*.

Botsou, F., Karageorgis, A.P., Dassenakis, E. and Scoullos, M. (2011). Assessment of heavy metal contamination and mineral magnetic characterization of the Asopos River sediments (Central Greece). Marine Pollution Bulletin, v. 62, pp. 547–563.

Bowen, A.J. and Holman, R.A. (1989). Shear instabilities of the mean longshore current: (1). Theory. Jour. Geophy. Res., v.94 (c12), pp. 18023-18030.

Boyd, R., Dalrymple, R., Zaitlin, B.A. (1992). Classification of clastic coastal depositional environments. Sedimentary Geology, v. 80, pp. 139– 150.

Brackley, H.L. (2006). Land to ocean transfer of erosion-related organic carbon, Waipaoa Sedimentary System, East Coast, New Zealand. PhD Thesis, Victoria University of Wellington, Wellington, 129 pp.

Bradley, R.S. (1999). Paleoclimatology: Reconstructing climates of the Quaternary. Academic Press, San Diego, p. 610.

Bruckner, H.(1989). Late Quaternary shorelines in India. In: D.B. Scott, P.A. Pirazzoli and C.A. Honig (Eds.) Late Quaternary sea level correlation and applications. Kluwer Academic Publishers, Dordrecht, pp. 169-194.

Brunn, P. and Nayak, B.U. (1980). Manual on protection and control of coastal erosion in India. NIO Special Publication.

Butman, B., Noble, M., Folger, D. (1979). Long-term observations of bottom current and bottom sediment movement on the mid-Atlantic continental shelf. Journal of Geophysical Research, v. 84 (C3), pp. 1187-1205.

Calcagno, T.H. and Ashley, G.M., (1984). Sedimentation processes in an impoundment, Union Lake, New Jersey. Environmental Geology Water Science, v. 6(4): pp. 237-246.

Caldas, L.H.O., Stattegger, K., Vital, H. (2006a). Holocene sea-level history: evidence from coastal sediments of the northern Rio Grande do Norte coastal, NE Brazil. Marine Geology v. 228, pp. 39-53.

Caldas, L.H.O., Oliveira, J.G., Medeiros Jr., W.E., Stattegger, K., Vital, H. (2006b). Geometry and evolution of the Holocene trangressive and regessive barriers on the semi-arid coast of NE Brazil. Geo-Marine Letters, v. 26, pp. 249-263.

Canfield, D. E. (1994). Factors influencing carbon preservation in marine sediments. Chemical Geology, v. 114, pp. 315-329.

Caratini, C.; Delibrias, G. and Rajagopalan, G. (1990). Paleomangroves of Kanara coast, Karnataka, India and their implications on Late Pleistocene sea-level changes. In. Jain, K.P. and Tiwari, R.S. (Eds.). Proceedings of symposium Vistas in Indian Palaeobotany, Palaeobotanist. V. 38, pp. 370-378.

Caratini, C. and Rajagopalan, G. (1992). Holocene marine transgression marker on the Karnataka coast, India. Indian Journal of Marine Sciences, v. 21, pp. 149-151.

Carr, A.S., Bateman, M.D., Roberts, D.L., Murray-Wallace, C.V., Jacobs, Z., Holmes, P.J. (2010). The last interglacial sea-level high stand on the southern Cape coastline of South Africa. Quaternary Research, v. 73, pp. 351-363.

Castaing, P., Froidefond, J.M., Lazure, P., Weber, O., Prud'homme, R., Jouanneau, J.M. (1999). Relationship between hydrology and seasonal distribution of suspended sediments on the continental shelf of the Bay of Biscay. DeepSea Research Part II: Topical Studies in Oceanography, v. 46 (10), pp. 1979-2001.

Carver, R.E. (1971). Procedures in Sedimentary Petrology (ed.). Wiley-Interscience, NewYork, p. 653.

Casagrande A. (1932). Research on the Atterberg limits of soils. Public Roads 13: pp. 121–130.

Casagrande A. (1948). Classification and identification of soils. Public Roads 13: pp. 121-136.

Casagrande, A. (1958). Notes on the design of the liquid limit device. Geotechnique. 8(2): pp. 84-91.

Chakravarty, M.; Patgiri, A. D. (2009). Metal pollution assessment in sediments of the Dikrong River, NE India. J. Hum. Ecol., v. 27 (1), pp. 63-67.

Chamley, H. (1989). Clay Sedimentology. Springer, Berlin. p. 623.

Chamley, H. (1990). Sedimentology. Springer-Verlag, Berlin, Heidelberg. 285p.

Chang, Y.H., Scimshaw, M.D., Macleod, C.L. and Lester, J.N. (2001). Flood defence in the Blackwater Estaury, Essex, UK: The impact of sedimentological and geochemical changes on salt marsh development in the Tollesbury Managed realignment site. Mar. Pollu.Bull., v. 42, pp. 470-481.

Chassefiere, B. & Monaco, A. (1987). Geotechnical properties and sedimentological processes of the Rhone continental margin. Marine Geology, v. 74, pp. 225–235.

Chatterjee, M., Silva, F. E. V. and Sarkar, S. K. (2007). Distribution and Possible Source of Trace Elements in the Sediment Cores of a Tropical Macrotidal Estuary and Their Ecotoxicological Significance. *Environment International*, v. 33: pp. 346–356.

Chauhan, O.S. (1994). Infuence of macrotidal environment on shelf sedimentation, Gulf of Kachchh, India. Continental Shelf Research, v. 14, pp. 1477-1493.

Che, Y., He, Q. & Lin, W. (2003). The distribution of particulate heavy metals and its indication to the transfer of sediments in the Changjiang Estuary and Hangzhou Bay. Marine Pollution Bulletin, v. 46 (1), pp. 123-131.

Chen, F., Wang, H. & Zheng, Z. (1998). Formation and evolution of underwater delta in Minjiang Estuary. J. Oceanogr. Taiwan Strait v. 17 (4), pp. 396–401.

Chen, X. H., Yin, A, Gehrels, G. E., Cowgill, E. S., Grove, M., Harrison, T. M., & Wang, X. F. (2003). Two phases of Mesozoic north-south extension in the eastern Altyn Tagh Range, northern Tibetan plateau: Tectonics, v. 22, article no. 1053.

Chester R. (1990). Marine geochemistry**,** Unwin Hyman, London, 698 pp.

Church, J.A., White, N.J. & Hunter, J.R. (2006). Sea-level rise at tropical Pacific and Indian Ocean islands. Global and Planetary Change, v. 53, pp. 155-168.

Colonna, M., Casanova, J., Dullo,W.Ch. & Camoin, G. (1996). Sea-level changes and $\delta^{18}O$ record for the past 34,000 years from Mayotte Reef, Indian Ocean. Quaternary Research, v. 46, pp. 335-339.

Comans, R.N.J. & Van Dijk, C. (1988). Role of complexation processes in cadmium mobilization during estuarine mixing. Nature, v. 336, pp. 151-154.

Compton, J.S. (2001). Holocene sea-level fluctuations inferred from the evolution of depositional environments of the southern Langebaan Lagoon salt marsh, South Africa. The Holocene, v. 11 (4), pp. 395-405.

Dagg, M.J., Bianchi, T., McKee, B. & Powell, R. (2008). Fates of dissolved and particulate materials from the Mississippi river immediately after discharge into the northern Gulf of Mexico, USA, during a period of low wind stress. Continental Shelf Research, v. 28, pp. 1443-1450.

Dagg, M.J., Ammerman, J.W., Amon, R.M.W., Gardner,W.S., Green, R.E. & Lohrenz, S.E. (2007). A review of water column processes influencing hypoxia in the Northern Gulf of Mexico. Estuaries and Coasts, v. 30, pp. 735-752.

Dale, B. (1976). Cyst formation, sedimentation, and preservation: Factors affecting dinoflagellate assemblages in recent sediments from Trodheimsfjord, Norway: Review of Palaeobotany and Palynology, v. 22, pp. 39–60.

Dalrymple, R.W., Zaitlin, B.A. & Boyd, R. (1992). Estuarine facies models; conceptual basis and stratigraphic implications. Journal of Sedimentary Petrology, v. 62, pp. 1130-1146.

Dalrymple, R. W. & Choi, K. (2007). Morphologic and facies trends through the fluvial–marine transition in tide-dominated depositional systems: A schematic framework for environmental and sequence-stratigraphic interpretation. Earth-Science Reviews, v. 81, pp. 135–174.

Degens E.T. & Ittekot, V. (1984). A new look at clay-organic interactions. Mitt Geologisch-Palaeontologischen Inst Univ Hamburg, v. 56, pp. 229-248.

de Jonge, V. N., Elliott, M. & Orive, E. (2002). Causes, historical development, effects and future challenges of a common environmental problem: eutrophication. Hydrobiologia, 475/476: pp. 1–19.

De Morais, J. O., Tintelnot, M. Irion, G. & Pinheiro, L. S. (2006). Pathways of clay mineral transport in the coastal zone of the Brazilian continental shelf from Ceará to the mouth of the Amazon River. Geo-Mar. Lett. V. 26: pp. 16–22. (DOI 10.1007/s00367-005-0011-1).

Desikachar, S. V. (1976). Geology and the hydrocarbon prospects of the Kerala west coast basin. Paper presented to the Workshop on Coastal Sedimentaries of India south of 18°N *ONGC, Madras*.

Desikachar, S. V. and Subramanyam, M. (1959). Progress report (Part II) on subsurface geology of parts of coastal Kerala. Madras and Kerala party 1968-59, ONGC Report, Unpublished.

Dias, C.R.R. and Alves, A.M.L. (2009). Geotechnical properties of the Cassino Beach mud. Continental Shelf Research, v. 29, pp. 589–596.

Drake, D.E. & Cacchione, D.A. (1985). Seasonal variation in sediment transport on the Russian River shelf, California. Continental Shelf Research, v. 4 (5), pp. 495-514.

Dyer, K.R. & Moffat, T.J. (1998). Fluxes of suspended matter in the east Anglian plume Southern North Sea. Continental Shelf Research, v. 18 (11), pp. 1311-1331.

Eadie, B.J., McKee, B.A., Lansing, M.B., Robbins, J.A., Metz, S. & Trefry, J.H. (1994). Records of nutrient enhanced coastal productivity in sediments from the Louisiana continental shelf. Estuaries, v. 17, pp. 754-765.

Eggleton, J. & Thomas, K.V. (2004). A review of factors affecting the release and bioavailablity of contaminants during sediment disturbance events. Environment International, v. 30, pp. 973-980.

Elliott, E. T., Paustian, K. & Frey, S. D. (1996). Modeling the measurable or measuring the modelable: A hierarchical approach to isolating meaningful soil organic matter fractionations. In Evaluation of Soil Organic Matter Models. Eds. D. S. Powlson, P. Smith and J. U. Smith. pp 161–179. NATO ASI Series, Springer-Verlag.

Elliot, C. W., Savin, S. M., Dong, H. & Peacor, D. R. (1997), A paleoclimate interpretation derived from pedogenic clay minerals from the Piedmont Province, Virginia. Chem. Geol., v. 142, pp. 201–211.

Ellison, J.C. (2008). Long-term retrospection on mangrove development using sediment cores and pollen analysis: a review. Aquatic Botany, v. 89, pp. 93–104.

Ellison, J.C. & Stoddart, D.R. (1991). Mangrove ecosystem collapse during predicted sea level rise: Holocene analogues and implications. Journal of Coastal Research, v. 7, pp. 151-165.

El Wakeel, C. K. & Riley, J. P. (1957). Determination of organic carbon in marine sediments. Jour. Cons. Perm. Int. Explor. Mar., v. 22, pp. 180-183.

Enos, P. & Perkins, R.D. (1979). Evolution of Florida Bay from island stratigraphy. Bulletin of the Geological Society of America, v.90, pp. 59-83.

Faegri, K. & Iversen, J. (1992). Textbook of pollen analysis, 4th ed. Alden Press, London, pp. 78–83.

Fairbanks, R. G. (1989). Nature, v. **342**, pp. 637–642.

Fang, T.H., Li, J.Y., Feng, H.M. & Chen, H.Y. (2009). Distribution and contamination of trace metals in surface sediments of the East China Sea. Mar. Environ. Res., v. 68, pp. 178 – 187.

Feng, H., Han, X., Zhang, J. & Yu, L. (2004). *Mar. Pollut. Bull.*, v. *49*, pp. 910.

Ferentinou, M., Hasiotis, T. & Sakellariou, M. (2012). Application of computational intelligence tools for the analysis of marine geotechnical properties in the head of Zakynthos canyon, Greece, Computers & Geosciences v. 40 pp. 166–174.

Fernandes, L., Nayak, G.N., Ilangovan, D. & Borole, D.V. (2011). Accumulation of sediment, organic matter and trace metals with space and time, in a creek along Mumbai coast, India. Estuarine, Coastal and Shelf Science, V. 91, pp 388-399.

Fettweis, M., Sas, M. & Monbaliu, J. (1998). Seasonal, neap-spring and tidal variation of cohesive sediment concentration in the Scheldt estuary, Belgium. Estuarine, Coastal and Shelf Science, v. 47 (1), pp. 21-36.

Folk, R.L. & Ward,W.C. (1957). Brazos River bar: a study in the significance of grain size parameters. Journal of Sedimentary Petrology, v. 27, pp. 3-26.

Folk, R.L. (1980). Petrology of Sedimentary Rocks. Austin, Texas, U.S.A. Hemphill Publishing Co., 2nd edition, p. 184.

Foote, R. B. (1883). On the geology of South Travancore. Rec. Geol. Surv., v. 6(1), pp. 90-35.

Forstner, U., & Muller, G. (1973). Heavy metal accumulation in river sediments: A response to environmental pollution. Geoforum 14/17, pp. 53 – 61.

Forstner, U. & Wittman, G. (1981). *Metal pollution in aquatic environment, 2nd* ed., Springer-Verlag, New York. p. 197.

Förstner, U., Ahalf, W., Calmano, W. & Kersten, M. (1990). Sediment criteria development- contributions from environmental geochemistry to water quality management. In: Heling, D., Rothe, P., Förstner, U., Stoffers, P. (Eds.), Sediments and Environmental Geochemistry: Selected Aspects and Case Histories. Springer, Heidelberg, pp. 311-338.

Furukawa, K. & Wolanski, E. (1996). Sedimentation in mangrove forests. Mangroves and Salt Marshes, v. 1, pp. 3 – 10.

Garcia-Montelongo, F., Diaz, C., Galindo, L., Larrechi, M. S. & Rius, X. (1994). Heavy metals in three fish species from the coastal waters of Santa Cruz de Tenerife (Canary Islands). Sci. Mar., v. 58, pp. 179–83.

Gaudette, H.E., Flight, W. R., Toner, L. & Folger, D.W. (1974). An inexpensive titration method for determination of organic carbon in recent sediments. Journal of Sedimentary Petrology, v. 44, pp. 249-253.

Gibbs, R.J. (1965). Error due to segregation in quatitative clay mineral X-ray diffraction mounting techniques. American Mineralogist, v. 50, pp. 741-751.

Gibbs, R. J. (1977). Clay mineral segregation in the marine environment. Journal of Sediment Petrology, v. 47: pp. 237-243.

Goni, M.A., Ruttenberg, K.C. & Eglinton, T.I. (1997). Sources and contributions of terrigenous organic carbon to surface sediments in the Gulf of Mexico. Nature, v. 389, pp. 275-278.

Goñi, M.A., Yunkerb, M.B., Macdonald, R.W. & Eglinton, T.I. (2005). The supply and preservation of ancient and modern components of organic carbon in the Canadian Beaufort Shelf of the Arctic Ocean. Marine Chemistry, v. 93, pp. 53-73.

Griffin, G.M. & Ingram, R.L. (1955). Jour. Sed. Petrol., v. 25, pp.194- 198.

Grim, R.E. (1968). Clay mineralogy. McGraw-Hill, New York, p. 600.

Grim, R.E., Dietz, R.S. & Bradley, W.F. (1949). Clay mineral composition of some sediments from the Pacific Ocean off the California coast and the Gulf of California. Bull. Geol. Soc. Am., V. 60, pp. 1785-1808.

GSI (1995). Geological and Mineral map of Kerala. Hyderabad, India: Geological Survey of India, Scale 1:500,000, 1 sheet.

Guo, L. D. & Santchi, P. H. (1997). Composition and cycling of colloids in marine environments. Rev Geophys., v. 35: pp. 17–40.

Guo, L., Santschi, P.H. & Bianchi, T.S. (1999). Dissolved organic matter in estuaries of the Gulf of Mexico. In: Bianchi, T.S., Pennock, J.R., Twilley, R.R. (Eds.), Biogeochemistry of Gulf of Mexico Estuaries. John Wiley and Sons, New York, pp. 269-299.

Hajjaji, W., Hachani, M., Moussi, B., Jeridi, K., Medhioub, M., López-Galindo, A., Rocha, F., Labrincha, J. A. & Jamoussi, F. (2010). Mineralogy and plasticity in clay sediments from north-east Tunisia. Journal of African Earth Sciences, v.57, pp. 41–46.

Händle, F. (Ed.). (2007). Extrusion in Ceramics. Springer, New York.

Hashimi, N.H., Kidwai, R.M. & Nair, R.R. (1978). Grain size and coarse fraction studies of sediments between Vengurla and Mangalore on the western continental shelf of India. *Indian* J. *Mar. Sci., v.* 7: pp. 231-238.

Hashimi, N.H. (1981). Comparative study of the moment and graphic size parameters of the sediments of the western continental shelf of India. *Mathematical Geology, v.*13(4): pp. 291-301.

Hashimi, N.H. & Nair, R.R (1981). Surficial sediments of the continental shelf off Karnataka. *Journal of Geological Society of India, v.* 22: pp. 266-273.

Hashimi, N. H., Nigam, R., Nair, R. R. & Rajagopalan G. (1995). Holocene sea level fluctuations on Western India Continental Margin: An update; *J. Geol. Soc. India*, v. **46**, pp. 157–162.

Hayes, M.O. (1975). Morphology of sand accumulations in estuaries. In: Estuarine research , 2nd Edition, v. 2, New York : Academic press, pp.2-22.

Hayes, M.O. (1980). General morphology and sediment patterns in tidal inlets. Sediment. Geol., v. 26, pp. 139–156.

Hedges, J.I. & Keil, R.G. (1995). Sedimentary organic-matter preservation - an assessment and speculative synthesis. Marine Chemistry, v. 49, pp. 81-115.

Hedges, J.I., Keil, R.G. & Benner, R. (1997). What happens to terrestrial organic matter in the ocean? Organic Geochemistry, v. 27, pp. 195-212.

Hobday, D.K. (1976). Quaternary sedimentation and development of the lagoonal complex, Lake St. Lucia, Zululand. Annuals of the South African Museumm v.71, pp.93-113.

Holland, K. T. & Elmore, P. A. (2008). A review of heterogeneous sediments in coastal environments. Earth-Science Reviews, v. 89. pp. 116–134.

Hori, K., Tanabe, S., Saito, Y., Haruyama, S., Nguyen, V. & Kitamura, A. (2004). Delta initiation and Holocene sea-level change: example from the Song Hong (Red River) delta, Vietnam. Sedimentary Geology, v. 164, pp. 237–249.

Hori, K. & Saito, Y. (2007). Classification, architecture, and evolution of large-river deltas. In: Gupta, A. (Ed.), Large Rivers: Geomorphology and Management. Wiley, Chichester, pp. 75-96.

Horowitz, A.J. & Elrick, K.A. (1987). The relation of stream sediment surface area, grain size and composition to trace element chemistry. Appl. Geochem., v. 2, pp. 437–451.

Huu, H. H., Rudy S. & An Van Damme. (2010). Distribution and contamination status of heavy metals in estuarine sediments near Cau Ong harbor, Ha Long Bay, Vietnam. *Geology Belgica, v.*13(1-2): pp. 37 – 47

Inman, D. L. & Nordstorm, C. E. (1971). On the tectonic morphologic classification of coasts: Jour. Geol., v. 79, pp. 1-21.

Ip, C.C.M., Li, X.D., Zhang, G., Wai, O.W.H. & Li, Y.S. (2007). Trace metal distribution in sediments of the Pearl River Estuary and the surrounding coastal area, South China. Environ. Pollut., v. 147, pp. 311 – 323.

Irion, G., Morais, J. & Bungenstock, F. (2012). Holocene and Pleistocene sea-level indicators at the coast of Jericoacoara, Ceará, NE Brazil. Quaternary Research, v. 77, pp. 251–257.

IS:2720 (Part – II). (1964). Methods for tests for soils – Part II. Determination of moisture content.

IS: 2720 (Part V) (1970). Methods of Test for Soils-Part V, Determination of Liquid and Plastic limits.

IS: 2720 (part XXX) (1980). Methods of test for soils (Part XXX) laboratory vane shear test.

IUCN (1983). Global status of mangrove ecosystems. *Commission on Ecology. Paper No.* 3, Switzerland, IUCN.

Jacob, K. & Sastri, V. V. (1952). Miocene foraminifera from Chavara near Quilon, Travancore. Rec. Geol. Surv. Ind., v. 82(2), pp. 342-459.

Jayalakshmi, K., Nair, K.M. & Kumai, H. (2001). Quaternary Geology of Kerala Coast: An Overview. Gondwana Research, V.4, No.4.

Jayalakshmi, K., Nair, K. M., Kumai, H. & Santhosh, M. (2004). Late Pleistocene–Holocene palaeoclimatic history of the southern Kerala basin, southwest India. Gondwana Res., v. 7, pp. 585–594.

Jayasurya, P.K.; Kaladharan, P., Rajagopalan, M.S., Dam Roy, S. & Sadhu, A.K. (2000). Mangrove vegetation. In Parayannilam. G.J. (Ed.). Mangrove ecosystems – A manual for the assessment of biodiversity, CMFRI, pp. 1-15.

Jordao, C.P., Perewira, M. G. & Perreira, J. L. (2002). Metal contamination of river waters and sediments from effluents of Kaotin Processing in Brazil. *J. Wat. Air. and Soil. Pollut*., v. 140: pp. 119-138.

Joseph, S. & Thrivikramaji, K.P. (2002). Kayals of Kerala coastal land and implication to Quaternary sea level changes. *In:* Narayana, A.C. (Ed.), Late Quaternary Geology of India and Sea level changes. Memoir Geol. Soc. of India, pp. 51-64.

Kabata-Pendias A. (2001). Trace Elements in Soils and Plants; CRC Press.

Kale, V.S. & Rajaguru, S.N. (1983). Mid-Holocene fossil wood from Colva, Goa. Current Science, v. 32, pp. 778-779.

Keller, G. H. (1982). Organic matter and the geotechnical properties of submarine sediments. Geo-Mar. Lett., v. 2: pp. 191-198.

Keil, R.G., Tsamakis, E., Giddings, J.C. & Hedges, J.I. (1998). Biochemical distributions (amino acids, neutral sugars, and lignin phenols) among size-classes of modern marine sediments from the Washington coast. Geochimica Et Cosmochimica Acta, v. 62, pp. 1347-1364.

Kessarkar, P.M., Rao, V.P., Ahmad, S.M. & Anil Babu, G. (2003). Clay minerals and Sr-Nd isotopes of the sediments along the western margin of India and their implication for sediment provenance. Marine Geology, v. 202, pp. 55-69.

Khadge, N.H. (2002). Geotechnical properties of siliceous sediments from the Central Indian Basin. Current Science, v. 82(3), pp. 338-342.

King, W. (1882). The Warkalli beds and associated deposits at Quilon in Travancore. Rec. Geol. Surv. Ind., v. 15(2), pp.93-102.

Knebel, H. J., Rending, R. R. & Bothner, M. H. (1991). Modern ssedimentary environments in Boston Harbour, Massachusetts. J. Sedimentary Petrology, v. 61(5), pp. 791-804.

Kolla, V., Henderson, L. & Biscaye P.E. (1976). Clay mineralogy and sedimentation in the western Indian Ocean. Deep-Sea Res., v. 23, pp. 949-961.

Kolla ,V., Kostecki, J. A., Robison, F., Biscaye, P.E. & Ray, P.K. (1981). Distribution and origin of clay-minerals and quartz in surface sediments of the Arabian Sea. Journal of Sedimentary Petrology, v. 51(2), pp. 563-569.

Kolla, V. & Kidd, R. B. (1982). Sedimentation and sedimentary processes in the Indian Ocean, In: The ocean basins and margins (eds.) A. E. Nair and F.G. Stehli (New York Plenum Press) v. 6. Pp. 1-50.

Krauss, K.W., Allen, J.A. & Cahoon, D.R. (2003). Differential rates of vertical accretion and elevation change among aerial root types in Micronesian mangrove fossils. Estuarine Coastal Shelf Science, v. 56, pp. 251– 259.

Krishnan M.S. (1968). The evolution of the coasts of India. Bull. Nat. Inst. of Sci. India, v. 38, pp.398-404.

Kumaran K.P.N. (2001). Quaternary sea level oscillations, geological and geomorphological evolution of south Kerala sedimentary basin, Report on the palynological analysis of subsurface samples. DST project interim report, 1 – 22 p.

Kumaran, K.P.N., Shindikar, M.R. & Mudgal, T.R. (2004a). Floristic composition, palynology and sedimentary facies of Hadi mangrove swamp (Maharashtra). Journal of Indian Geophysical Union, v. 8, pp. 55–63.

Kumaran, K.P.N., Shindikar, M.R. & Limaye R.B. (2004b). Mangrove deposits as archives of palaeomonsoon records: the palynological assessment of the Late Quaternary deposits of West Coast of India. Pollen, v. 14, pp. 246–247.

Kumaran, K.P.N. & Nair, K.M. (2005). Tracing Palaeoclimatic signatures in fossil woods and sub-fossil logs of Kerala, Southwestern India. Pages News, v. 13, pp. 15–17.

Kwon, T., Lee, K., Cho, G. & Lee, J. (2011). Geotechnical properties of deep oceanic sediments recovered from the hydrate occurrence regions in the Ulleung Basin, East Sea, offshore Korea. Marine and Petroleum Geology, v. 28, pp. 1870 – 1883.

Liu, J. T. & Zarillo, G. A. (1990). Shoreface dynamics: evidence from bathymetry and surficial sediments. Mar. Geol., v. 94. pp. 37-54.

Laluraj, C. & Nair, S. (2006). Geochemical index of trace metals in the surficial sediments from the western continental shelf of India, Arabian Sea, Environmental Geochemistry and Health, v. 28, pp. 509-518.

Lambeck, K. & Chappel, J. (2001). Sea level change through the last glacial cycle. Science, v. 292, pp. 679-686.

Lario, J., Spencer, C., Plater, A.J., Zazo, C., Goy, J.L. & Dabrio, C.J. (2002). Particle size characterisation of Holocene back-barrier sequences from North Atlantic coasts SW Spain and SE England. Geomorphology, v.42, pp.25-42.

Laughman, F .C. & Craig, D.C. (1962). Aust. Jour. Marine. Fresh Water. Res., V. 13, pp. 48-56.

Li, C.X., Zhang, J.Q., Fan, D.D. & Deng, B. (2001). Holocene regression and the tidal radial sand ridge system formation in Jiangsu coastal zone, East China. Mar. Geol., v. 173, pp. 97-120.

Limaye, R.B. (2004). Contribution to Palaeopalynology and palaeoecology of the coastal deposits of Maharashtra, India, Ph.D. Dissertation, University of Pune, Pune.

Limaye, R.B. & Kumaran, K. P. N. (2012). Mangrove vegetation responses to Holocene climate change along Konkan coast of south-western India, Quaternary International, v. 263, pp. 114-128.

Liu S.Y., Shao, G. H., Du, Y. J. & Cai, G. J. (2011). Depositional and geotechnical properties of marine clays in Lianyungang, China. Engineering Geology, v. 121 pp. 66–74.

Liu, W.X., Li, X.D., Shen, Z.G., Wang, D.C., Wai, O.W.H. & Li, Y.S. (2003). Multivariate statistical study of heavy metal enrichment in sediments of the Pearl River Estuary. Environmental Pollution, v. 121, pp. 377-388.

Louda, J. W., Loitz, J. W., Melisiotis, A. & Orem, W. H. (2004). Potential Sources of Hydrogel Stabilization of Florida Bay Lime Mud Sediments and Implications for Organic Matter Preservation. Journal of Coastal Research, V. 20, pp. 448 – 463.

Lyons, T.W., Werne, J.P., Hollander, D.J. & Murray, R.W. (2003). Contrasting sulfur geochemistry and Fe/Al and Mo/Al ratios across the last oxic-to-anoxic transition in the Cariaco Basin, Venezuela. Chemical Geology, v. 195, pp. 131–157.

Macias, C., Schifter, I., Liuch-cota, D.B., Mendez-Rodriguez, L. & Hernandez-Vazquez, S. (2006). Distribution, enrichment and accumulation of heavy metals in coastal sediments of Salina Cruz Bay, Mexico. Environmental Monitoring and Assessment, v. 118, pp. 211-230.

Magesh N. S., Chandrasekar, N. & Roy, D. V. (2011). Spatial analysis of trace element contamination in sediments of Tamiraparani estuary, southeast coast of India. Estuarine, Coastal and Shelf Science, V. 92, pp. 618-628.

Mallik, T.K. & Suchindan, G.K. (1984). Some sedimentological aspects of Vembanad Lake, Kerala, west coast of India. Ind. Jour. Mar. Sci., v.13, pp.159-163.

Mann, D.H. & Streveler, G.P. (2008). Post-glacial relative sea level, isostasy, and glacial history in Icy Strait, southeast Alaska, USA. Quaternary Research, v. 69, pp. 201-216.

Manojkumar, P., Narayana, A. C. & Tatavarti, R. (1998). Mudbank dynamics: physical properties of sediments. J. Geol. Soc. India, v. 51, pp. 793–798.

Martin, L., Dominguez, J.M.L. & Bittencourt, A.C.S.P. (2003). Fluctuating Holocene sea-levels in eastern and southeastern Brazil: evidence from multiple fossil and geometric indicators. Journal of Coastal Research v. 1, pp. 101–124.

Mascarenhas, A.; Paropakari, A.L. & Babu, C.P. (1993). On the possibility of allochthonous peat on the inner shelf off Karwar. Current Science, v. 64, pp. 684-687.

Mascarenhas, A. (1997). Significance of peat on the western continental shelf of India. Journal Geological Society of India, v. 49, pp. 145-152.

McKee, B.A. (2003). RiOMar: the Transport, Transformation and Fate of Carbon in River dominated Ocean Margins. Report of the RiOMar Workshop, 1–3 November 2002. Tulane University, New Orleans, LA. 48 pp.

McLaren, P. (1981). An interpretation of trends in grain size measures. Journal of Sedimentary Petrology, V. 51, No.2, pp. 0611-0624.

McLaren, P. & Bowles, D. (1985). The effect of sediment transport on grain-size distributions, Journal of Sedimentary Petrology, V. 55, No. 4, pp. 0457-0470.

McLennan, S.M. (2001). Relationships between the trace element composition of sedimentary rocks and upper continental crust. Geochem. Geophys. Geosyst., 2(4), doi: 10. 1029/2000GC000109.

Means, W.E. & Parcher, J.V. (1963). Physical properties of soils. Charles E. Merril, Columbus, Ohio, 476 pp.

Mehta. A.J., Parchure, T.M., Dixit, J.G. & Ariathurai, R. (1982). Re-suspension potential of deposited cohesive sediment beds. In Kennedy, V. (Ed.). Estuarine Comparisons, Academic Press, New York, pp. 591-609.

Menon, K.K. (1966). Sequence and source of the Tertiaries near Karruachal, Trivandrum district, Bull. Geol. Surv. Ind, v. 3, pp, 75-78.

Meunier, A., (2005). Clays. Springer, Berlin, 472 pp.

Meybeck, M. (1982). Carbon, nitrogen, and phosphorus transport by World Rivers. American Journal of Science, v. 282, pp. 401-450.

Meybeck, M., Cauwet, G., Dessery, S., Somville, M., Gouleau, D. & Billen, G. (1988). Nutrients (Organic C, P, N, Si) in the eutrophic river Loire (France) and its estuary. Estuarine Coastal and Shelf Science, v. 27, pp. 595-624

Mil-Homens, M., Stevens, R.L., Abrantes, F. & Cato, I. (2006). Heavy metal assessment for surface sediments from three areas of the Portuguese continental shelf. Continental Shelf Research, v. 26, pp. 1184-1205.

Miller, C.V., Foster, G.D. & Majedi, B.F. (2003). Baseflow and stormflow metal fluxes from two small agricultural catchments in the Coastal Plain of the Chesapeake Bay Basin, United States. Appl. Geochem., v. 18, pp. 483 – 501.

Milliman, J.D. & Meade, R.H. (1983). World-wide delivery of river sediment to the oceans. J. Geol., v. 91, pp. 1 – 21.

Mitchell, S.B. & West, J.R. (2002). Particle size distribution in an estuarine turbidity maximum region. In: Winterwerp, J.C., Kranenburg, C. (Eds.), Fine Sediment Dynamics in the Marine Environment. Proceedings in Marine Science, v. 5, pp. 251-263.

Morton, R.A., Paine, J.G. & Blum, M.D. (2000). Response of stable bay-margin and barrier-island system to Holocene sea-level highstands, western Gulf of Mexico. Jour. Sed. Res., v. 703, pp.478-490.

Muller, J. (1964). A palynological contribution to the history of the mangrove vegetation in Borneo. In: Cranwell, L.M. (Ed.), Ancient Pacific Floras. University of Hawaii Press, Honolulu, pp. 33–42.

Muller, G. (1969). Index of geoaccumulation in sediments of the Rhine River. Geo. J., v. 2 (3), pp. 108-118.

Muller, J. & Caratini, C. (1977). Pollen of Rhizophora (Rhizophoraceae) as a guide fossil. Pollen et Spores, v. 19, pp. 361–389.

Muller, P.J. & Suess, E. (1979). Productivity, sedimentation rate and sedimentary organic matter in the oceans: I. Organic carbon preservation. Deep – Sea Research, v. 26A, pp. 1347-1362.

Murty, C.S., Sastry, J.S. & Varadachari, V.V.R. (1980). Shoreline deformations in relation to shore protection structures along Kerala coast. Indian Jour. Mar. Sci., v. 9, pp. 77-81.

Murty, P.S.N. (1977). Evolution of the Netravati drainage, Karnataka State. South India. Indian Journal of Earth Sciences, v. 4, pp. 197-202.

Murty, P.S.N. & Veerayya, M. (1972a). Studies on the sediments of Vembanad lake, Kerala state: Part I - Distribution of organic matter. Ind. Jour. Mar. Sci., V. 1, pp. 45-51.

Murty, P.S.N. & Veerayya, M. (1972b). Studies on the sediments of Vembanad lake, Kerala state Part 11 Distribution of phosphorus. Ind. Jour. Mar. Sci., V. 1, pp. 106-115.

Murty, P.S.N. & Veeryya, M. (1981). Studies on the sediments of Vembanad lake, Kerala State. Part IV Distribution of trace elements. Ind. Jour. Mar. Sci., V. 10, pp. 165-17.

Murty, Y. G. K., Varadan, G. N., Nair, K. B. Gopalakrishnan, L. S. & Chauhan, R. P. S. (1976). A review of the sedimentary formations of Kerala. Paper presented to the Workshop on Coastal Sedimentaries of India south of 18°N *ONGC, Madras*.

Muslim, I. & Jones, G. (2003). The seasonal variation of dissolved nutrients, chlorophyll a and suspended sediments at Nelly Bay, Magnetic Island. Estuarine, Coastal and Shelf Science, v. 57 (3), pp. 445-455.

Naidu, A.S., Mowatt, T.C., Somayajulu, B.L.K. & Rao, K.S. (1985). Characteristics of clay minerals in the bed loads of major Rivers of India. Mitteilungen aus dem geologischpalaontologischen Institut, University of Hamburg. Heft, v. 2, pp. 559-568.

Nair, A.S.K., Sankar, G. & Nalinakumar, S. (1993). Remote sensing application in the study of migration and offset of coastal inlets. Jour. of the Indian Soc. of Remote sensing, v. 21, no.3, pp. 151-156.

Nair, A.S.K. & Sankar, G. (1995). Classification and Evaluation of coastal wetlands of Kerala using Indian Remote Sensing Satellite IRS-1A LISS II data. Proceedings of the seventh Kerala Science congress, Palakkad, pp.46-48.

Nair, A.S.K. & Nalinakumar, S. (1997). Remote sensing application in the study of sediment deposition on a microtidal environment. Proceedings of the ninth Kerala science congress, Thiruvanathapuram, pp.30-31.

Nair, C.K., Balchand, A.N. & Chacko, J. (1993). Sediment characteristics to changing hydrography of Cochin estuary. Ind. Jour. Mar. Sci., v.22, pp.33-36.

Nair, K.K. (1987). Geomorphological and Quaternary geological studies along the coastal plains in parts of Cannanore and Kasargod districts, Kerala. J. Geol. Soc. India, v.29, pp.433-439.

Nair, K.M. & Padmalal, D. (2003). Quaternary Sea level Oscillations, Geological and Geomorphological Evolution of South Kerala Sedimentary Basin. PCR ESS/23/VES/6/98; Report submitted to DST, GOI., 1 –60 p.

Nair, K.M., Padmalal, D. & Kumaran, K.P.N. (2006). Quaternary geology of South Kerala sedimentary basin – an outline. Journal of Geological Society of India, v. 67, pp. 165–179.

Nair, M.M. (1987). Coastal geomorphology of Kerala. Jour. Geol. Soc. India, v.29, pp.450-458

Nair, M.M. (1999). Quaternary coastal geomorphology of Kerala. Indian Jour. Of Geomorphology, v. 4, pp.51- 60.

Nair, R. R, Hashimi, N. H., Kidwai, R. M., Guptha, M. V. S., Paropakari, A. L., Ambre, M. V., Muralinarh, A. S., Mascarenhos, A. & D'Costa, G. P. (1978). Topography and sediments of the western continental shelf of India - Vengurla to Mangalore. *Indian.J.Mar.Sci.,v.* 7: pp. 224-230.

Nair, R.R., Hashimi, N.H. & Rao, V.P. (1982). Distribution and dispersal of clay minerals on the western continental shelf of India. Marine Geology, v. 50, M1-M9.

Nair, S.M., Balchand, A.N. & Nambisan, P.N.K. (1990). Metal concentrations in recently deposited sediments of Cochin backwaters, India. The science of the total environment, v.97-98, pp.507-524.

Nair, M. & Ramachandran, K.K. (2002). Textural and trace metal distribution in sediments of Beypore estuary. Indian J. of Mari Science, v. 31: pp.295-304.

Nameroff, T.J., Calvert, S.E. & Murray, J.W. (2004). Glacial interglacial variability in the eastern tropical North Pacific oxygen minimum zone recorded by redox-sensitive trace metals. Paleoceanography, 19(1).

Narayana, A.C. & Pandarinath, K. (1991). Sediment transport direction derived from grain size statistics on the continental shelf off Mangalore, west coast of India. Jour. Geol. Soc. India, v. 38, pp. 293-298.

Narayana, A.C., Naidu, P.D., Shinu, N., Nagabhushanam, P. & Sukhija, B.S. (2009). Carbonate and organic carbon content changes over last 20 ka in the Southeastern Arabian Sea: Paleoceanographic implications; Quaternary International, V. 206 pp. 72–77.

Narayana, A.C., Priju, C.P. & Chakrabarti, A. (2001a). Identification of paleodelta near the mouth of Periyar River in central Kerala. Journal Geological Society of India. V. 57, pp. 545 - 547.

Narayana, A.C., Priju, C.P. & Rajagopalan, G. (2001b). Late Quaternary peat deposits from Vembanad lagoon, Kerala, SW Coast of India. Current Science, v. 83, pp. 318-321.

Narayana, A.C., Priju, C.P. & Rajagopalan, G. (2002). Late Quaternary peat deposits from Vembanad Lake lagoon, Kerala, SW coast of India. Curr. Sci., v.83, no.3, pp.318-321.

Narayana A. C. & Priju C. P. (2004). Spatial and Temporal Variability of Trace Element Concentrations in a Tropical Lagoon, SouthwestCoast of India: Environmental Implications; Journal of Coastal Research; ICS (Proceedings); Brazil ISSN 0749-0208; PP- 1053 – 1057.

Narayana, A.C. & Priju, C.P. (2004). Evolution of Coastal landforms and Associated Sedimentary Environments of Central Kerala, Southwest Coast of India. 8th International Coastal Symposium, Proceedings ICS 2004, Santa Catarina, Brazil. Jour. Coast. Res., Special Issue, 39 *(in press)*.

Narayana A.C. & Priju C.P. (2006). Landform and shoreline changes inferred from satellite images along the Central Kerala coast; Journal Geological Society of India, v.68, pp.35-49.

Narayana, A.C. (2007). Peat Deposits of the West Coast of India: Implications for Environmental and Climate Changes during Late Quaternary, Journal of Coastal Research, Special Issue v. 50, pp. 683-687.

Narayana A.C., Jago, C.F., Manojkumar, P. & Tatavarti, R. (2008). Nearshore sediment characterstics and formation of mudbanks along the Kerala coast, southwest India; Estuarine coastal and shelf science, v.78, pp. 341-352.

Narayanan, K. (1958). Report on the geological reconnaissance of Kerala. ONGC Report, Unpublished.

Narayanan, V. & Anirudhan, S. (2003). Coastal geomorphology of Kerala from Vizhinjam to Kochi; Indian Journal of Geomorphology, v.8, no. 1&2, pp.35-42.

Nasnolkar, C.M., Shirodkar, P.V. & Singbal, S.Y.S. (1996). Studies on organic carbon, nitrogen and phosphorous in the sediments of Mandovi estuary, Goa. Ind. Jour. Mar. Sci., v.25, pp.120-124.

Nayak, S. & Bahuguna, A. (2001). Application of Remote Sensing Data to Monitor Mangroves and other Coastal Vegetation of India. Indian Journal of Marine Sciences, v. 30, pp. 195– 213.

Nichols, M.M. & Biggs, R.B. (1985). Estuaries. In: Davis Jr. R.A. (Ed.), Coastal Sedimentary Environments, second ed. Springer-Verlag, New York, pp. 77-186.

Nichols, M.M., Johnson, G.H. & Peebles, P.C. (1991). Modern sediments and facies model for a microtidal coastal plain estuary, The James estuary, Virginia. Jour. Sed. Petrol., v.61, no.6, pp. 883-899.

Norström, E., Risberg, J., Gröndahl, H., Holmgren, K., Snowball, I., Mugabe, J. A. & Sitoe, S. R. (2012). Coastal paleo-environment and sea-level change at Macassa Bay, southern Mozambique, since c 6600 cal BP. Quaternary International, v. 260, pp. 153-163.

Nriagu, J.O. & Pacyna, J.M. (1988). Quantitative assessment of worldwide contamination of air, water and soils by trace metals. Nature, v. 333, pp.134–139.

Odell, R. T., Thornburn, I. H. & McKenzie, L. J. (1960). Relationships of Atterberg limits to some other properties of Illinois soils. Soil Sci. Soc. Am. Proc., v. 24: pp. 297-300.

Pache, T., Brockamp, O. & Clauer, N. (2008). Varied pathways of river-borne clay minerals in a near-shore marine region: a case study of sediments from the Elbe and Weser rivers, and the SE North Sea. Estuarine Coastal and Shelf Science, v. 78, pp. 563-575.

Padmalal D. (1992). Mineralogy and geochemistry of the sediments of Muvattupuzha river and central Vembanad estuary, Kerala, India, Ph.D. thesis submitted to Cochin University of Science and Technology, Cochin, Kerala.

Padmalal D., Kumaran K.P.N., Nair K.M., Baijulal B., Limaye Ruta B. & Mohan, S.V. (2011). Evolution of coastal wetland system of SW India during the Holocene: Evidence from marine and terrestrial archives of Kollam coast, Kerala. Quaternary International, V. 237, pp. 123-139.

Panda, D., Subramanian, V. & Panigrahy, R.C. 1995. Geochemical fractionation of heavy metals in Chilka lake east coast of India – a tropical coastal lagoon. Environ. Geol., v.26, pp.199-210.

Pandarinath, K. (1991). Sedimentological, Mineralogical and geochemical studies of riverine, estuarine and nearshore sediments of Gangolli area, Karnataka, India. Ph.D. thesis (unpubl.), 194p.

Pandarinath, K. & Narayana, A.C. (1998). Elemental distribution in the innershelf sediments off Coondapur, west coast of India. Journal Geological Society of India, v.51, pp. 493-508.

Pandarinath, K., Shankar, R. & Yadava, M.G. (2001). Late quaternary changes in sea level and sedimentation rate along the sea coast of India: evidence from radiocarbon dates. Current Science, v. 81, pp. 594–600.

Parkinson, R.W. (1989). Decelerating Holocene Sea-level Rise and its Influence on SW Florida Coastal Evolution: A Transgressive/Regressive Stratigraphy. Journal of Sedimentary Petrology, v. 59, pp. 960-972.

Patchineelam S.M. & De Figueiredo A. M. (2000). Preferential settling of smectite on the Amazon continental shelf. Geo-Mar. Lett. V. 20: pp.37–42.

Paulose, K.V. & Narayanaswami, S. (1968). The Tertiaries of Kerala coast. Mem. Geol. Soc. India., no.2, pp.300-308.

Pedersen, T.F. & Calvert, S.E. (1991). Anoxia vs productivity: What controls the formation of organic-carbon-rich sediments and sedimentary rocks? Am. Assoc. Pet. Geol., v. 74: pp. 454-466.

Peltier, W.R. (1988). Lithospheric thickness, Antarctic deglaciation history, and ocean basin discretization effects in a global model of postglacial sea level change: a summary of some sources of nonuniqueness. Quaternary Research, v. 29, pp. 93–112.

Peltier, W.R. (1998). Global glacial isostatic adjustment and coastal tectonics. In: Stewart, I., Vita-Finzi, C. (Eds.), Coastal Tectonics—Geological Society London: Special Publication, v. 146, pp. 1–29.

Pisharody, P.R. (1992). Keralathile Kalavasta (Climate in Kerala). State Institute of Language Publ. 26p. (in Malayalam).

Powar, S.D., Venkataramana, B., Mathai, T. & Mallikarjuna, C. (1983). Progress Report (unpubli.)., Geological Survey of India, Trivandrum.

Prakash, K. & Sridharan, A. (2004). "Free Swell Ratio and Clay Mineralogy of Fine-grained Soils", Geotechnical Testing Journal, ASTM, 27(2): 220–225.

Prakash, T.N. & Prithviraj, M. (1988). A study of seasonal longshore transport direction through grain size trends: An example from the Quilon coast, Kerala, India. Ocean and Shoreline Management, v. 11, pp. 195-209.

Priju, C. P. (2004). Studies on recent landform changes and Sedimentological aspects of Central Kerala Coast, Ph.D. thesis submitted to Cochin University of Science and Technology, Cochin Kerala.

Priju, C.P. & Narayana, A.C. (2004). Spatial and temporal variability of trace element concentrations in a Tropical lagoon, Southwest coast of India: Environmental Implications. Proc. 8th International Coastal Symposium, Santa Catarina, Brazil. Jour. Coast. Res. Special Issue, 39 *(in press).*

Priju, C. P. & Narayana A.C. (2007). Heavy and Trace Metals in Vemanad Lake Sediments; Int. J. Environ. Res., V. 1(4) Autumn, pp. 280-289.

Priju, C.P. & Narayana A.C. (2007). Particle size characterization of late Holocene depositional processes in Vembanad lagoon, Kerala: Inferences from suite statistics; Journal Geological Society of India, v. 69, pp. 311-318.

Quasim, S.Z. & Reddy, C.V. (1967). The estimation of plant pigments of Cochin backwater during the monsoon months. Bull. Mar. Sci., v.17, pp. 95-110.

Quasim, S.Z., Bhattathiri, P.M.A. & Abidi, S.A.H. (1968). Solar radiation and its penetration in a tropical estuary. Jour. Exp. Mar. Bio. Ecol., v.2, pp.87-103.

Radakovitch, O., Roussiez, V., Ollivier, P., Ludwig, W., Grenz, C. & Probst, J.L. (2008). Input of particulate heavy metals from rivers and associated sedimentary deposits on the Gulf of Lion continental shelf. Estuar. Coast. Shelf Sci., v. 77, pp. 285 – 295.

Raj, S.M. & Jayaprakash, M. (2007). Distribution and enrichment of trace metals in marine sediments of Bay of Bengal, off Ennore, south-east coast of India. Environmental Geology, v. 56,pp. 207-217.

Rajendran, C.P.; Rajagopalan, G. & Narayanaswamy. (1989). Quaternary geology of Kerala: Evidence from radiocarbon dates. Journal Geological Society of India, v. 33, pp. 218-222.

Ramsay, P.J. & Cooper, J.A.G. (2002). Late Quaternary sea-level change in South Africa. Quaternary Research, v. 57, pp. 82-90.

Rao, B. R. J. & Datta, P. M. (1976). Foraminiferal and Ostracod Fauna of the early Miocene beds of Kerala coast and their stratigraphic significance. Paper presented to the Workshop on Coastal Sedimentaries of India south of 18°N *ONGC, Madras*.

Rao, P.S., Ramaswamy, V. & Thwin S. (2005). Sediment texture, distribution and transport on the Ayeyarwady continental shelf, Andaman Sea, Marine Geology, V. 216, p. 239– 247.

Rao, V.P. (1991). Clay mineral distribution in the continental and slope of Suarashtra, west coast of India. Indian Journal of Marine Science, v.20, pp.1-6.

Rao, V.P. & Rao B. R. (1995). Provenance and distribution of clay minerals in the sediments of the western continental shelf and slope of India. Continental Shelf Research, V. 15, No. 14, pp. 1757-1771.

Rasheed, K., Joseph, K.A. & Balchand, A.N. (1995). Impacts of harbour dredging on the coastal shoreline features around Cochin. Proceedings of the international conference on ‘Coastal Change 95’ (Bordomer-IOC, Bordeaux, France). pp. 943-948.

Rashid, M. (1974). Adsorption of metals on sedimentary and peat humic acids: Chemical Geology, v. 13, p. 115-123.

Rashid, M.A. & Brown, J.D. (1975). Influence of Marine organic compounds on the engineering properties of remoulded sediment. Engineering Geology, v. 9, pp.141-154.

Raymond, P.A. & Bauer, J.E. (2001a). Use of 14C and 13C natural abundance for evaluating riverine, estuarine, and coastal DOC and POC sources and cycling: a review and synthesis. Organic Geochemistry, v.32, pp.469-485.

Regnier, P. & Wollast, R. (1993). Distribution of trace metals in suspended matter of the Scheldt estuary. Marine Chemistry, v.43 (1-4), pp.3-19.

Resmi T.R. (2004). Hydrogeochemical Evaluation of Inorganics and Bioorganics in selected Aquatic Environments, Ph.D. thesis submitted to Cochin University of Science and Technology, Cochin, Kerala.

Rimmer, S.M. (2004). Geochemical paleoredox indicators in Devonian-Mississippian black shales, Central Appalachian Basin (U.S.A). Chemical Geology, v. 206, pp. 373-391.

Rimmer, S.M., Thompson, J.A., Goodnight, S.A. & Robl, T. (2004). Multiple controls on the preservation of organic matter in Devonian-Mississippian black shales: Geochemical petrographic evidence. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 215, pp. 125-154.

Risberg, J., Miller, U. & Brunnberg, L. (1991). Deglaciation, Holocene shore displacement and coastal settlement in eastern Svealand, Sweden. Quaternary International, v.9, pp. 33-37.

Romankevich, E. A. (1984). Geochemistry of organic matter in the ocean. Springer.

Rominger, J.F. & Rutledge, P.C. (1952) Use of soil mechanics data in correlation and interpretation of Lake Agassiz sediments. J Geol., v.60(2), pp.160–180.

Rosenquist, I. Th. (1962). The influence of physicochemical factors upon the mechanical properties of clays. Clays and Clay Minerals, v.9, pp.12-27.

Roy, P. S., Thom, B. G. & Wright, L. D. (1980). Holocene sequences on an embayed high-energy coast: an evolutionary model. Sedimentary Geology, v. 26: pp.1-19.

Rubio, B. Nombela, M.A. & Vilas, F. (2000). Geochemistry of major and trace metals in sediments of the Ria De Vigo NW Spain: an assessment of metal pollution. Mar. Pollu. Bull., v.40. no.11, pp.968-980.

Sajan, K. & Damodaran, K.T. (1981) Studies on the distribution of organic matter content in the sediments of the Ashtamudi lake, Kerala. Bull. Dept. Mar. Sci., University of Cochin, Vol. 12, pp. 155-160.

Sakellari, A., Marta Plavsic, M., Karavoltsos, S., Dassenakis M. & Scoullos, M. (2011). Assessment of copper, cadmium and zinc remobilization in Mediterranean marine coastal sediments. Estuarine, Coastal and Shelf Science, v.91, pp.1-12.

Sampath, S. & Vinayak, P.V.S.S.K. (1989). Rainfall in Kerala. Report, CESS, Thiruvanathapuram, 17p.

Sankaranarayanan, V.N., Udayavarma, P., Balachandran, K.K., Pylee, A. & Joseph, T. (1986). Estuarine characteristics of the lower reaches of the river Periyar (Cochin backwater). Indian Jour. Mar. Sci., v.15, p.166-170.

Sarika, P.R. (2005). Biogeoorganics and trace metal speciation in mangroves, Ph.D. thesis submitted to Cochin University of Science and Technology, Cochin, Kerala.

Satyanarayana, D., Panigrahy, P.K. & Sahu, S.D. 1994. Metal pollution in harbour and coastal sediments of Visakhapatnam, east coast of India. Ind. Jour. Mar. Sci., v.23, pp.52-54.

Schmitz, R.M., Schroeder, Ch. & Charlier, R. (2004). Chemo–mechanical interactions in clay: a correlation between clay mineralogy and Atterberg limits. Applied Clay Science, v.26, pp.351– 358.

Schoellhamer, D. (1996). Factors affecting suspended-solids concentrations in south San Francisco Bay, California. Journal of Geophysical Research, v.101 (C5), pp.12,087-12,095.

Selvam, V. (2003). Environmental classification of mangrove wetlands of India. Current Science, v.84, pp.757–765.

Seralathan, P., Meenakshikutty, N.R., Asarafe, K.V. & Padmalal, D. (1993). Sediment and organic carbon distributions in the Cochin harbour area. Ind. Jour. Mar. Sci., v.22, pp.252-255.

Seralathan, P. & Padmalal, D. (1994). Textural studies of the sediments of Muvattupuzha river and central Vembanad estuary, Kerala. Jour. Geol. Soc. India, v.43, pp.179-190.

Seshan, B. R. R., Natesan, U. & Deepthi, K. (2010). Geochemical and statistical approach for evaluation of heavy metal pollution in core sediments in southeast coast of India. Int. J. Environ. Sci. Tech., v.7 (2), pp.291-306.

Seshan B. & Ranga Rama. (2010). Vertical profile of heavy metal concentration in core sediments of Buckingham canal, Ennore; Indian Journal of Geo Marine Sciences, V. 40(1), pp.83-97

Shajan, K.P. (1998). Studies on the Late Quaternary sediments and sea level changes of Central Kerala Coast, India. PhD thesis, Cochin University of Science and Technology, Kochi.

Sharma, G. S. (1978). Upwelling off the southwest coast of Indian. Journal of Marine Sciences, v. 7(4), pp. 209-218.

Shetye, S.S., Sudhakar, M., Mohan, R. & Tyagi, A. (2009). Implications of organic carbon, trace elemental and $CaCO_3$ variations in a sediment core from the Arabian Sea. Indian Journal of Marine Sciences, v.38, pp.432-438.

Shiller, A.M. & Boyle, E.A. (1991). Trace elements in the Mississippi River Delta outflow region: behavior at high discharge. Geochim. Cosmochim. Acta, v.55, pp.3241–3251.

Shinu, N. (2008). Environment and climate changes during the Late Quaternary: Inferences from sedimentary records of Southeastern Arabian Sea. Thesis submitted to Cochin University of Science & Technology, Cochin, Kerala.

Silva, A. J. (1974). Marine geomechanics: overview and projections. In Deep Sea Sediments, Physical and Mechanical Properties (ed. A. L. Inderbitzen). London: Plenum Press, pp. 45-76.

Simoneit, B.R.T. (1978). The organic chemistry of marine sediments. *In:* J.P. Riley and R. Chester Eds., Chemical Oceanography, v.7, 2nd Edition, p.233.

Singer, A. (1984). The paleoclimatic interpretation of clay minerals in sediments–a review. Earth-Science Review, v. 21, pp.251-293.

Singh, A. & Punmia, B. C. (1970). Soil Mechanics and Foundations. Standard Book-house. Delhi.

Skempton, A.W. (1953). The colloidal “Activity” of clays. Proceedings of the 3rd International Conference of Soil Mechanics and Foundation Engineering. V.(1) pp.57-60.

Sleighter, R.L. & Hatcher, P.G. (2008). Molecular characterization of dissolved organic matter (DOM) along a river to ocean transect of the lower Chesapeake Bay by ultrahigh resolution electrospray ionization Fourier transform ion cyclotron resonance mass spectrometry. Marine Chemistry, v.110, pp.140-152.

Smerdon, E.T. & Beasley, R.P. (1961). “Critical tractive forces in cohesive soils”. Agricultural Engineering, v.42(1), pp 26-29.

Smith, D.E. (1968). Strandline. The Encyclopedia of Geomorphology. v.III, Edited by Fairbridge, R.W., Reinhold Book corporation. New York, p.1053.

Sodorblom, R. (1966). Chemical aspects of quick clay formation. Engineering Geology, v.1, pp.415-431.

Soman, K. (2002). Geology of Kerala. Bangalore, India: Geological society of India, 336p.

Soman, K., Chattopadhyay, M., Chattopadhyay, S. & Potti, G.K. (2002). Occurrence and water resource potential of fresh water lakes in south Kerala and their relation to the Quaternary geologic evolution of the Kerala coast. In: Narayana, A.C. (Ed.), Late Quaternary Geology of India and Sea Level Changes, vol. 49. Memoir of Geological Society of India, pp. 17-29.

Spagnoli G, Rubinos D, Stanjek H, Ferna´ndez-Stegeer T, Feinendegen M. & Azzam, R. (2012). Undrained shear strength of clays as modified by pH variations. Bull. Eng. Geol. Environ., v.71(1): pp.135–148.

Spencer, R.G.M., Ahad, J.M.E., Baker, A., Cowie, G.L., Ganeshram, R., Upstill- Goddard, R.C. & Uher, G. (2007). The estuarine mixing behaviour of peatland derived dissolved organic carbon and its relationship to chromophoric dissolved organic matter in two North Sea estuaries (U.K.). Estuarine Coastal and Shelf Science, v.74, pp.131-144.

Spitzy, A. & Ittekkot, V. (1991). Dissolved and particulate organic matter in rivers. In: Mantoura, R.F.C., Martin, J.M., Wollast, R. (Eds.), Ocean Margin Processes in Global Change. John Wiley and Sons, New York, pp. 5-17.

Sridharan, A., Rao, S.M. & Murthy, N.S. (1986). Compressibility behaviour of homoionised bentonites. Geotechnique, v.36(4): pp.551-564.

Sridharan, A., Rao, S.M. & Murthy, N.S. (1988). Liquid limit of kaolinitic soils. Geotechnique, v.38(2): pp.191-198.

Sridharan, A. & Prakash, K. (2000). Classification procedures for expansive soils, Proc. Instn Civ. Engrs Geotech. Engng, v.143, pp.235-240.

Srivastava, P., Parkash, B. & Pal, D.K. (1998). Clay minerals in soils as evidence of Holocene climatic change, Central Indo-Gangetic Plains, North-Central India. Quaternary Research, v.50, pp.230-239.

Stanley, D.J. & Warne, A.G. (1994). Worldwide initiation of Holocene marine deltas by deceleration of sea-level rise. Science, v.265, pp.228– 231.

Sternberg, R. (1986). Transport and accumulation of river-derived sediment on the Washington continental shelf, USA. Journal of the Geological Society, v.143 (6), pp.945-956.

Steffen, H. & Kaufmann, G. (2005). Glacial isostatic adjustment of Scandinavia and northwestern Europe and the radial viscosity structure of the Earth´s mantle. Geophysical Journal International, v.163 (2), pp.801–812.

Stewart, R. H., Pilkey. O. H. & Nelson, B. W. (1965). Sediments of the northern Arabian Sea. *Mar. Geol., v.*3: pp.411-427.

Stiros, S.C. (2001). Subsidence of the Thessaloniki (northern Greece) coastal plain, 1960–1999. Eng. Geol., v.61, pp.243–256.

Subramanian. K.N. (2002). Wet land resources with particular reference to the mangroves of Kerala. In: *Compendium on the Focal Theme of* 14 *th Kerala Science Congress.* Published by State committee on Science, Technology and Environment (Kamalakshan, K., Premachandran, P. N., Biju, K., Eds.) pp. 119 - 130.

Subramanian, V. (1980) Mineralogical input of suspended matter by Indian rivers in to the adjacent areas of the Indian Ocean. Marine Geology, v.36, pp.M29-M34.

Sun, D.H., Bloemendal, J., Rea, D.K., Vandenberghe, J., Jiang, F.C., An, Z.S. & Su, R.X. (2002). Grain-size distribution function of polymodal sediments in hydraulic and aeolian environments, and numerical partitioning of the sedimentary components. Sedimentary Geology, v.152, pp.263-277.

Sunil Kumar, R. & Antony, A. (1994). Impact of environmental parameters on polychaetous annelids in the mangrove swamps of Cochin, south west coast of India. Ind. Jour. Mar. Sci., v.23, pp.137-142.

Sutherland, R. A. (2000). Bed Sediment-Associated Trace Metals in an Urban Stream, Oahu, Hawaii. *Environmental Geology*, v.39: pp.611–627.

Svendsen, J.I. & Mangerud, J. (1987). Late Weichselian and Holocene sea-level history for a cross-section of western Norway. Journal of Quaternary Science, v.2, pp.113-132.

Swift, D.J.P. (1976). Coastal sedimentation. In: Stanley, D.J., Swift, D.J.P. (Eds.), Marine Sediment Transport and Environmental Management. Wiley, New York, pp. 255–311.

Tatavarti, R., Narayana, A.C., Manojkumar, P. & Ravishankar, M. (1996). Mudbank dynamics: Field evidence of edge waves and FIG waves. Current Science, v.70, pp. 837-843.

Thamban, M., Rao V. P. & Schneider, R. R. (2002). Reconstruction of late Quaternary monsoon oscillations based on clay mineral proxies using sediment cores from the western margin of India, Marine Geology, v. 186 pp. 527-539.

Thrivikramji, K.P., Joseph, S. & Narayana, A.C. (2007). Coastal wetlands of Kerala: origin and evolution. Proceedings Kerala Environment Congress, pp.36-48.

Thom, B.G. (1984). Coastal landforms and geomorphic processes. In: Snedaker, S.C., Snedaker, J.G. (Eds.), The Mangrove Ecosystem: Research Methods. Unesco, pp. 3– 17.

Thrivikramaji, K.P. (1987). On the evolution of the lagoon barrier system between N. Lats. 12° 5′ and 12° 15′, Cannanore district, Kerala. Jour. Geol. Soc. India, v.29, pp.405-411.

Tissot, C. (1990). Late Holocene environment in Coondapur area, Karnataka. Preliminary palynological results. In. Jain, K.P. and Tiwari, R.S. (Eds.). Proceedings of symposium Vistas in Indian Palaeobotany, Palaeobotanist, v.38, pp.348-358.

Tomilson, D. C., Wilson, D. J., Harris, C. R. & Jeffrey, D. W. (1980). Problem in assessment of heavy metals in estuaries and the formation of pollution index. Helgol. Wiss. Meeresunlter, v.33 (1-4), pp.566-575.

Torricelli, S., Knezaurek, G. & Biffi, U. (2006). Sequence biostratigraphy and paleoenvironmental reconstruction in the Early Eocene Figols Group of the Tremp-Graus Basin (south-central Pyrenees, Spain). Palaeogeography, Palaeoclimatology, Palaeoecology, v.232, pp.1–35.

Tribovillard, N., Averbuch, O., Devleeschouwer, X., Racki, G & Riboulleau, A. (2004a). Deep-water anoxia over the Frasnian-Famennian boundary (La Serre, France): a tectonically-induced oceanic anoxic event? Terra Nova, v.16, pp.288-295.

Truong, M. H., Nguyen, V. L., Ta, T. K. O. & Takemura, J. (2011). Changes in late Pleistocene–Holocene sedimentary facies of the Mekong River Delta and the influence of sedimentary environment on geotechnical engineering properties, Engineering Geology, v. 122, pp. 146–159.

Tucker, R.W. & Vacher, H.L. (1980). Effectiveness of descriminating beach, dune and river sands by moments and cumulative weight percentages. Jour. Sed. Petrol., v. 50. pp.165-172.

Turekian, K. K. & Wedepohl, K. H. (1961). Distribution of the elements in some major units of the earth's crust. Geol. Soc. Am. Bull., v.72, pp.175 – 192.

Valdiya, K. S. & Narayana, A. C. (2007). River response to neotectonic activity: Example from Kerala, India. J. Geol. Soc. India, v.70, pp.427–443.

Vanoni, V. A. (ed). (1975). Sedimentation Engineering. Manuals & Reports on Engineering Practice, no. 54, ASCE, New York, USA.

Varadarajan, K. & Balakrishnan, M. K. (1976). Geomorphology of the Kerala Coastal Belt. Paper presented to the Workshop on Coastal Sedimentaries of India south of 18°N *ONGC, Madras*.

Veerayya, M., Murty, C.S. & Varadachari, V.V.R. (1973). Effect of Artificially Dumped Material on the configuration of Baina Beach, Goa. Ind. Jour. Mar. Sci., v.2, pp.78-83.

Veerayya, M. & Varadachari, V.V.R. (1975). Depositional environments of coastal sediments of Calangute, Goa. Sed. Geol., v.14, pp.63-74.

Veerayya, M. & Murty, P.S.N. (1974). Studies on the sediments of Vembanad lake, Kerala state: Part III- Distribution & Interpretation of Bottom Sediments. Ind. Jour. Mar. Sci., v.3, pp.16-27.

Velde, B**.** (1992). The stability of clays, in *The stability of minerals*, edited by C. D. Price and N. L. Ross, The Mineralogical Society Series, Chapman and Hall, London v.3, pp. 329-351.

Venkataramaiah, C. 1993. Geotechnical Engineering, Wiley Eastern Limited. New Delhi, p.892.

Verma,K.K & Mathur, U.B. (1979). Quarternary sea level changes along west coast of India. Geol. Soc.India, no.45. pp.263-272.

Vernal, A & Rochon, A. (2011). Dinocysts as tracers of sea-surface conditions and sea-ice cover in polar and subpolar environments. IOP Conf. Series: Earth and Environmental Science, v.14.

Vink, A., Steffen, H., Reinhardt, L. & Kaufmann, G. (2007). Holocene relative sea-level change, isostatic subsidence and the radial viscosity structure of the mantle of northwest Europe (Belgium, the Netherlands, Germany and southern North Sea). Quaternary Science Reviews, v.26, pp.3249–3275.

Visher, G.S. (1969). Grain size distributions and depositional processes. Jour. Sed. Petrol., v.39, pp.1074-1106.

Wakeham, S.G., Canuel, E.A., Lerberg, E.J., Mason, P., Sampere, T.P. & Bianchi, T.S. (2009). Partitioning of organic matter in continental margin sediments among density fractions. Marine Chemistry, v.115, pp.211-225.

Wang, Z., Xu, H., Zhan, Q., Saito, Y., He, Z., Xie, J., Li, X. & Dong, Y. (2010). Lithological and Palynological evidence of late Quaternary depositional encironments in the subaqeous Yangtze delta, China; Quaternary Research, v. 73, pp. 550-562.

Wang, X., Cai, Y. & Guo, L. (2010). Preferential removal of dissolved carbohydrates during estuarine mixing in the Bay of St Louis in the northern Gulf of Mexico. Marine Chemistry, v.119, pp.130-138.

Weaver, C. E. (1989). Clays, muds and shales. Developments in Sedimentology. 44. Elsevier, Amsterdam.

Wedepohl, K.H. (1971). Environmental influences on the chemical composition of shales and clays. In: Ahrens LH, Press F, Runcorn SK, and Urey HC (Eds) Physics and chemistry of the Earth. Oxford: Pergamon, v.8, pp.307-331.

Wedepohl, K.H. (1991). Chemical-Composition and Fractionation of the Continental-Crust. Geologische Rundschau, v.80(2), pp.207-223.

Werne, J.P., Lyons, T.W., Hollander, D.J., Formolo, M.J. & Damste, S. (2003) Reduced sulphur in euxinic sediments of the Cariaco Basin: sulphur isotope constraints on organic sulphur formation: Chemical Geology, v.195, pp.159-179.

Whitehouse, U. G., Jeffrey, L. M. & Debbrecht, J. D. (1960) Differential settling tendencies of clay minerals in saline waters: Seventh Nat. Con. Clays and Clay Minerals, pp. 1-79.

Wilson, M.J. (1999). The origin and formation of clay minerals in soils: past, present and future perspectives. Clay Minerals, v.34, pp.7-25.

Wolanski, E. (1995). Transport of sediment in mangrove swamps. Hydrobiologia, v.295, pp.31– 42.

Wollast, R. (1991). The coastal organic carbon cycle: fluxes, sources, and sinks. In: Mantoura, R.F.C., Martin, J.M., Wollast, R. (Eds.), Ocean Margin Processes in Global Change, Dahlem Workshop Report. Wiley, Chichester, West Sussex, England, pp. 365–381.

Woodroffe, C.D. (1983). Development of mangrove forests from a geological perspective. In: H.J. Teas (Ed.) Biology and ecology of mangroves, pp.1-18.

Woodroffe, C.D.; Thom, B.G. & Cappell, J. (1985). Development of widespread mangrove swamps in mid Holocene times in Northern Australia. Nature, v.317, pp.711-713.

Woodroffe, C.D. (1993). Late Quaternary evolution of coastal and lowland riverine plains of southeast Asia and northern Australia: an overview. Sedimentary Geology, v.83, pp.163–175.

Woodroffe, C. D., Chappell, J., Thom, B.G. & Wallensky, E. (1989). Depositional model of a macrotidal estuary and floodplain, South Alligator River, northern Australia: Sedimentology, v. 36, p. 737–756.

Woodroffe, S.A. & Horton, B.P. (2005). Holocene sea-level change in the Indo-Pacific. Journal of Asian Earth Sciences, v.25, pp.29-43.

Yadava, M.G. & Ramesh, R. (1999). Speleothems–useful proxies for past monsoon rainfall. Journal Science India Research, v.58, pp.339-348.

Yan, W. M. & Ma, Y. (2010). Geotechnical characterization of Macau marine deposits. Engineering Geology, v.113 pp.62–69.

Yasar, D., Aksu, A.E. & Uslu, O. (2001). Anthropogenic pollution in Izmit Bay: heavy metals in surface sediments. Turkish Journal of Engineering and Environmental Sciences, v.25, pp.299-313.

Yoon, H. I., Yoo, K., Bak, Y., Lee, Y. I. & Lee, J. I. (2009). Core-based reconstruction of paleoenvironmental conditions in the southern Drake Passage (West Antarctica) over the last 150 ka, Geo-Mar. Lett., v.29: pp.309–320.

Zhang, J. (1999). Heavy metal compositions of suspended sediments in the Changjiang (Yangtze River) estuary: significance of riverine transport to the ocean. Continental Shelf Research, v.19 (12), pp.1521-1543.

Zhang, L. P., Ye, X. & Feng, H. (2007). Heavy Metal Contamination in Western Xiamen Bay Sediments and Its Vicinity, China. *Marine Pollution Bulletin*, v.54: pp.974–982.

Zhou, J.L., Liu, Y.P. & Abrahams, P.W. (2003). Trace metal behaviour in the Conwy estuary, North Wales. Chemosphere, v.51, pp.429–440.

Zuther, M., Brockamp, O. & Clauer, N. (2000). Composition and origin of clay minerals in Holocene sediments from the south-eastern North Sea. Sedimentology, v.47, pp.119-134.

Zwartz, D., Bird, M., Stone, J. & Lambeck, K. (1998). Holocene sea-level change and ice-sheet history in the Vestfold Hills, east Antarctica. Earth and Planetary Science Letters, v.155, pp.131-145.

****