Geology of sediment core retrieval from borehole No. 60
Sotrant, Cortalim, Mormugao Taluka, Goa

A Dissertation report for
GEO-651: Dissertation
Credits: 16
Submitted in partial fulfilment of Master's Degree
MSc in Applied Geology
by

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May 2024

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DECLARATION BY STUDENT

I hereby declare that the data presented in this Dissertation report entitled, "Geology of sediment core retrieval from borehole No.60 Sotrant, Cortalim, Mormugao Taluka, Goa" is based on the results of investigations carried out by me in the M.Sc. Applied Geology at the School of Earth, Ocean and Atmospheric Sciences, Goa University under the Supervision of Dr. A.A.A.A Viegas, Associate Professor, and the same has not been submitted elsewhere for the award of degree or diploma by me. Further, I understand that Goa University or its authorities will be not be responsible for the correctness of observations / experimental or other findings given the dissertation.

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CHAPTER I : INTRODUCTION

PROLOGUE

A tiny emerald green country on the west coast of India, Goa is a small coastal state known for its beautiful beaches, picturesque landscapes and rich cultural heritage. The state also has various geological formations that provide insight into the geological history of the area. It lies between latitudes 14°53’54”N and 15°40’00”N and longitudes 73°40’ 33”E and 74°20’13”E. Goa is spread over an area of 3702 sq km and has two revenue districts: North Goa and South Goa. The boundaries of Goa are defined in the north by the Terekhol River, which separates the state from Maharashtra, and in the east and south by Karnataka and the Arabian Sea in the west. Goa is located on the west coast of India and has a coastline of 104 km.

Goa’s landscape is topographically complex from west to east. The region can be divided into three major topographic units: the vast coastal plain, the central hilly region, and the Western Ghats. The coastal and estuarine plains consist of linear coasts with sand dunes, river-crossed wetlands, and intersecting rocky outcrops. The central plateau consists of mountainous lateritic plateau hills, usually covered with trees. The high peaks of the forested Ghats lie further inland.

Goa is located in the northwestern part of the Western Dharwar Craton, where the Shimoga-Goa belt on the crust extends continuously until it is hidden beneath the Arabian Sea and the Deccan Traps to the north. This belt probably continues below the trap to Narmada, where the Narmada-Son lineament terminates this supracrustal belt. The Shimoga-Goa belt extends from north-northwest to south-southeast for about 250 km and its maximum width at Dharwar is about 120 km.

The western end of the belt is characterized by large main blocks of gneisses such as Chandranath granite gneiss and Canacona granite. A fault can be seen at the eastern end, but the southern part appears to have been deposited in shallow water. The southern section is generally characterized by an orthoquartzite/mafic lava combination, indicating a subsurface environment and intermittent volcanic activity in a coastal sand environment under stable conditions. The Goa-North Canara limestone-iron-manganese-ore sequence (Castlerock zone) in the northern part of the Shimoga belt (McLearn, 1904) is physically continuous with that in the southern part and serves as a marker horizon indicating continuity. It forms part of the rock along the entire length of the belt.

Granitic gneisses from Goa have been assigned either syntectonic or posttectonic status (Harinadha Babu et al. 1981a; Gokul et al. (1985) on deformation episodes in crustaceans. However, there are considerable uncertainties about the interactions between granitic gneisses, on the one hand, and between gneisses and upper crustal greenstones, on the other hand. In particular, the upper crustal sequence is composed of metavolcanic rocks and clastics, each with its own characteristics, but both assigned to one stratigraphic group, namely the Goa Group (Gokul et al.1985).

COASTAL SEDIMENTS

Coastal sediments comprise pieces of solid material that may be moved due to water motion (waves or currents) but do not float (van Rijn 1998). Coastal sediments, commonly sand grains, occur along the coast, with a wide range of origins, structures, sizes and chemical compositions. Sediment may develop through weathering of rock, shells or shell fragments; organic debris or chemical precipitation. They may be delivered to the coast by river flows, wave action or currents. At the coast, the decline of river currents as they enter the ocean and the landward push by waves determines that the coastal margin is a preferred location for coastal sediments to deposit.

Material accumulating near the coast is redistributed by waves, currents and tides. However, sediment accumulations may affect these water motions, and with sufficient feedback this provides the basis for development of coastal landforms (Woodroffe 2003). The most significant type of coastal landform-process interaction occurs at the shoreline itself, where the reduced mobility of sediment out of the water determines that accumulation often occurs horizontally, building soft shores, including beaches and low-relief coastal landforms, for example foredunes, banks and coastal terraces.

Coastal sedimentary landforms are naturally dynamic, responding to variation of waves, winds, currents and water levels over a wide range of time scales. This sensitivity determines that they are among the first coastal features affected by natural or artificial change to existing conditions. The ability of soft shores to adjust to different conditions may provide a high degree of resilience to coastal change, provided land-use planning allows sufficient scope for coastal movement. The high amenity of soft shores and, often, the value of adjacent infrastructure, determine that perceived adverse impacts of coastal dynamics give high sociopolitical pressure for management responses. However, the spatial connections of soft shores determine that action without an appropriate understanding of coastal dynamics often leads to propagation or an increase of the problem.(Eliot 2016).

AIM AND OBJECTIVE

The purpose of this study was to become familiar with borehole data of Sotrant, Cortalim sediments, study sediment properties, gain insight into sediment mineralogy with particular emphasis on clay minerals, and decipher probable paleoenvironment conditions.

- Understanding the borehole data of Sotrant, Cortalim sediments.
- Investigating sediment properties to gain into sediment mineralogy, particularly focusing on clay minerals.
- Megascopically and microscopically examining rock remains associated with the sedimentary deposits to identify the rock type.
- Delineating sediment fractions at different depths through sediment characterization.
- Using XRD examination to identify different minerals present in samples at various depths.
- Plotting sediment fractions on a Perjup diagram to understand hydrodynamic conditions.
- Indexing sediment color according to the Munsell color chart.
- Combining all generated data to prepare a log.

CHAPTER II : LITERATURE REVIEW

GEOLOGY OF GOA

The supracrustals that constitute the Goa Group of Gokul et al. (1985) can be divided into two lithostratigraphic sequences namely the Barcem Group and the Ponda Group (Dessai 2011). The former comprises predominantly greenstones (metabasalts) and rests on a basement of the 3300-3400 Ma Anmode Ghat trodhjemite gneiss with a crudely developed quartz-pebble conglomerate at the base, and shows lithological similarities with the lower part of the Bababudan Group.

The younger sequence is dominated by clastics, and is assigned to a new stratigraphic Group formally termed the Ponda Group which is equivalent to the Chitradurga Group of the Dharwar Super Group. This Group rests on a basement of the 2700-2900 Ma Chandranath granite gneiss with a distinct unconformity marked by a polymict, granite- clast metaconglomerate. The conglomerate displays many similarities with the Talya conglomerate that occurs at the base of the Chitradurga Group. It is overlain by a psamolitic sequence which is followed in ascending order by the chemogenic sediments that host the BIF and by the deep water turbidite sequence (argillite-graywacke association) with intercalations of mafic volcanics. The supracrustal sequence is intruded by the Bondla layered mafic-ultramafic complex. The Ponda Group makes up the three formations namely the Sanvordem, Bicholim and Vagheri.

Residual rocks		Laterites
Basic intrusives (late)	65-56 Ma	Dolerites
Deccan traps	67-64 Ma	Basalts
Basic intrusives		Metadolerites
Bondla mafic-ultramafic layered complex		Dunite-peridotite-gabbro and equivalents
Ponda Group	Vagheri Formation	Metabasalts Argillites and metagreywack
	Bicholim Formation	Banded ferruginous quartzites, Manganiferous chertbreccia with pink ferruginous Phyllites Limestones Pink ferruginous Phyllites Quartz-Chlorite-Amphibole schists
	Sanvordem Formation	Metagreywack Argillites, quartzites, tilloid
	Unconformity	
Barcem Group	Barcem Formation	Metagabbros Peridotite, talc-chlorite schists Quartzites, quartz-sericite schists, Red Phyllites, Quartz porphyry Massive, schistose and vesicular metabasalts
Cancona Granite	2979 + 4 Ma	Potassic porphyritic granite
Chandranath Granite Gneiss	2900-2700 Ma	Granodiorite gniess
Basement: Anmode Ghat Trondhjemite Gneiss	3400-3300 Ma	Tonalite-Trondhjemite-Granodiorite (TTG) Gniess

Table no 2.1: Lithostratigraphic classification of supracrustal rocks from Goa after Dessai(2011)

BARCEM GROUP

The Barcem Group is best developed to the southeast of Barcem village and account for a thickness of over 2 km, however, in the northwest between Margao and Vasco-da-Gama the rocks attain a thickness of about 2.5 km (Ram Das and Ramprasad Rao, 2001). It has a quartz-pebble conglomerate at the base and the lithounits consist of metavolcanics with intercalations of quartzites and pelites. The volcanics are represented by basic and felsic lavas, agglomerates and tuffs. The metasediments consist of quartzites, quartz-sericite-schists, quartz-chlorite-schists and minor phyllites.

Non-vesicular metabasalts are exposed around Astagal, Polem, Padi and Subdalem in the south and Tisk (Usgao)- Dharbandora in central Goa. Vesicular metabasalts are exposed to the north of Gulem, at Barcem and along the Saleri-Vagon road in Canacona taluka. At places phyllitic intercalations are noticed. These probably represent the metamorphosed tuffs. Well preserved pillow lavas are described from Gokuldem-Maina section (Gokul et al.1985). The metabasalts at the base of the Mormugao headland are schistose and at places vesicular. Those from Bogmalo are vesicular, amygdaloidal and are dominated by pyroxene and at places contain olivine (Devaraju et al. 2010). The metabasalts in general, consist of saussuritised plagioclase and uralitised pyroxene forming actinolite, chlorite, epidote, clinozoisite along with opaques.

Metavolcanics that could be silicic lavas or pyroclasts are exposed at Barcem. They are represented by quartz- Sericite-schists and quartz porphyry. At places they are crushed and contain porphyroclasts of quartz in a pulvurised quartz-sericite matrix.

PONDA GROUP

The Ponda Group is best developed around Ponda town where most of the lithologies are well exposed. It comprises three formations which in ascending order are the Sanvordem Formation, Bicholim Formation and the Vagheri Formation, all of which are dominated by clastics with very subordinate mafics only in the Vagheri Formation. Around Ponda, the total thickness of this group is about 8.8 km (Ram Das and Ramprasad Rao, 2001).

Sanvordem Formation:

The Sanvordem Formation rests on the Chandranath granite gneiss with a polymict metaconglomerate at the base and comprises metagreywacke, and argillites. The conglomerate consists of stretched and elongated pebbles of quartzite and gneiss in a schistose chlorite matrix. The proportion of lithologies varies considerably from place to place, so much so that one single lithology may be present to the complete exclusion of the other members. The best section of this formation is exposed along the railway tract between Sanvordem railway sation and Periudoc. In this section it has a thickness of over 1.2 km. To the east of Kalay railway station the litho unit is a metagreywacke which contains intercalations of metaconglomerate. It is also well exposed between Bati and Keri south of river Tiracol. Exposures occur at Ribandar in the vicinity of Panaji where it consists of metagreywack with subordinate metaconglomerate The metaconglomerate consists of stretched lenticular pebbles and boulders of quartzite and granite gneiss (particularly at Sanvordem) that vary in size from 10 x15 x 25 cm along the longer dimension and are enclosed in a schistose matrix dominated by chlorite. The exposures across metaconglomerate. The exposures across river Mandovi between Aguada and Baga are dominated by thinly bedded (laminated) argillites.The metagreywackes are faintly schistose and show graded bedding. The rock consists of sub-angular quartz, plagioclase and lithic fragments in a matrix of sericite, chlorite and quartz. The argillites are light to deep grey and consist of quartz in a sericite-chlorite matrix with opaques. rarely biotite has formed at the expense of chlorite.

Bicholim Formation:

This formation can be traced over the entire length of Goa (~ 185 km) in a NW-SE direction from Naibag in the northwest to Salgini in the southeast. The average true thickness of the Formation is about 1.4 Km (Gokul et al. 1985). It consists of amphibole schists, ferruginous and manganiferous phyllites, limestones and banded ferruginous quartzites (BHQ) that occur as intercalations within the phyllites. The BHQs serve as the protores for the iron ore deposits that are extensively developed in this formation. The BHQs are thinly laminated and consist of alternate laminae of hematite/magnetite and chert.

Magnetite is invariably martitised. The banded iron formation (BIF) consists of two subfacies- the haematite sub-facies and the magnetite sub-facies, both of which show interdigitated relationship. The hematite sub-facies predominates in the northwestern and central part (from Advalpal to Dharbandora) of the State whereas the magnetite sub-facies is largely confined to the central and the Southeastern part (Costi to Barazana-Villiena). Calcareous (carbonate facies of BIF) and carbonaceous (sulphide Facies) intercalations are common. Manganese being an inseparable associate of iron in almost all BIF the world over, both these exhibit a zonal distribution pattern. Broadly the iron-rich protores are confined to the northern part of the formation, northwest of Sanguem, whereas the manganese-rich protores are confined to the southern part of the State, largely to the south of Sanguem between Rivona and Salgini.

The dominant lithology is represented by quartz-chlorite- tremolite schists followed by ferruginous phyllites which is a ubiquitous lithology of this formation. The rock consists of tremolite and chlorite with quartz and opaques as accessories. Limestone exposures are rare and occur to the northeast of Goa. Calcareous intercalations also occur in association with the phyllites and are encountered in bore holes below the manganiferous phyllites. They are also intercalated by cherts. The rock consists of calcite and dolomite with some tremolite.

Vagheri Formation:

The topmost formation of this Group is represented by Vagheri Formation which conformably overlies the Bicholim Formation. It is best exposed to the northeast of Valpoi. It comprises metagreywacke-argillite with intercalated metabasalts. The metagreywackes are grey to greyish green, compact and exhibit poorly developed schistosity. The rock consists of angular to sub-angular crystic- and lithic-fragments in a fine grained mesostasis made up of quartz, feldspar and chlorite. The rocks are immature with poorly sorted angular and sub-angular clasts of andesitic tuffs (Devaraju et al. 2010). The metabasalts occur as narrow, lenticular intercalations within the metagreywacke and exhibit a poorly developed foliation at places. The rock is grey to greenish grey, hard, compact and has faint schistosity. It consists of chlorite, tremolite/actinolite, plagioclase, epidote, zoisite, opaques and quartz which may be secondary.

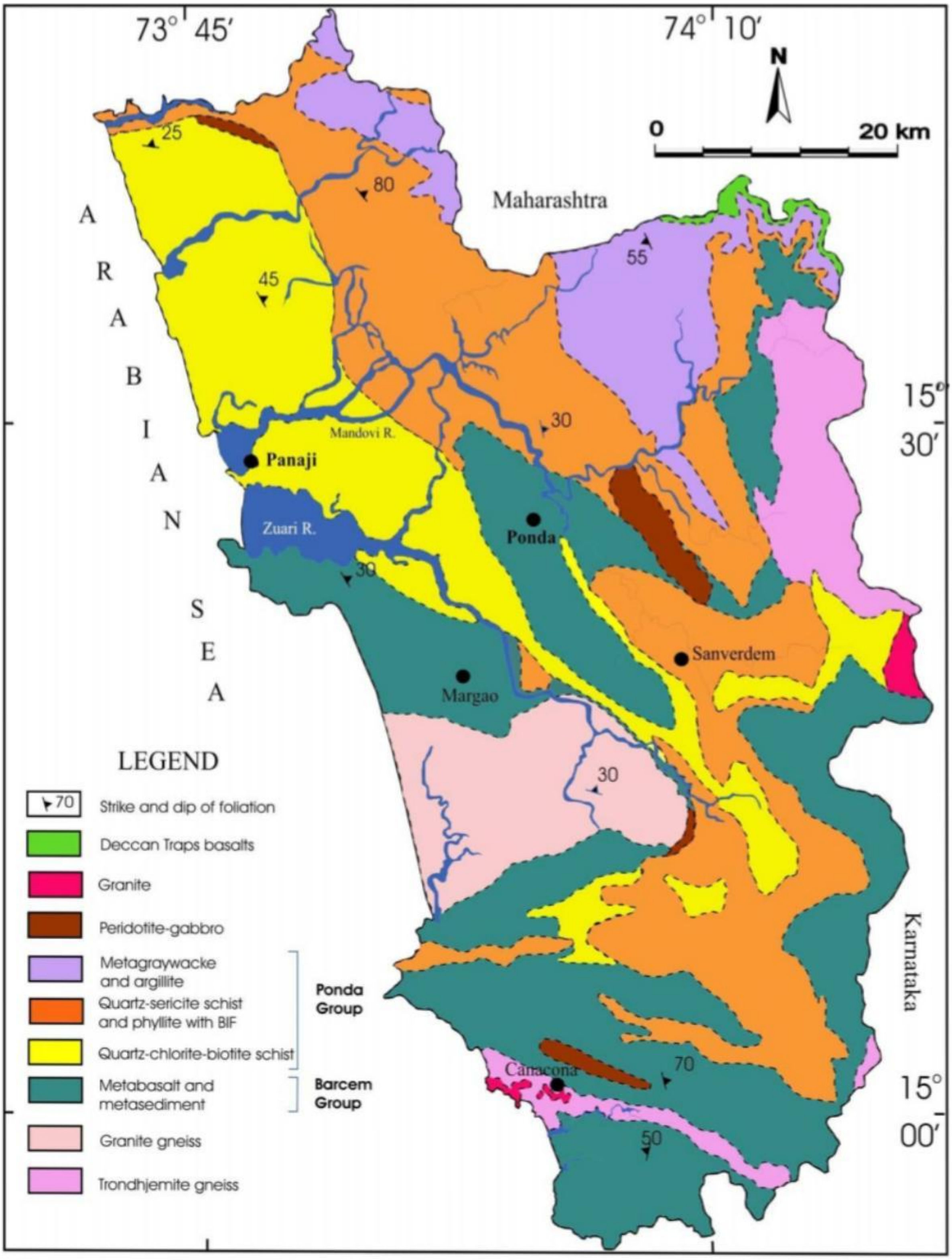


Figure 2.1: Geological sketch map of Goa after Dessai (2011)

STRUCTURE AND TECTONICS

The rocks of the Goa Group are oriented in a general NW-SE direction and have experienced three folding cycles in the past. Initially, during the first folding phase (F1), the rocks acquired a WNW-ESE orientation, particularly noticeable in the southern part of the state. The Chandranath granite, dated around 2650 million years ago, is thought to have formed concurrently with this folding phase. In the second folding phase (F2), the rocks aligned in a NW-SE direction, with overturned southwestern limbs, mainly prominent in the north-central and northeastern regions of Goa. Intense folding in the eastern parts of the state led to schistosity and axial plane cleavage in many rock formations, with the Western Ghat granite gneiss forming during this phase. The third folding phase (F3) (Gokul et al.1987) resulted in northwesterly plunging broad synclinal folds in the northeastern parts of Goa.

The straight coastline suggests a major fault along the west coast, associated with several weak planes. In South Goa, folding intensity is high, and metamorphic layering often exhibits vertical or steeply dipping characteristics with a similar WNW-ESE strike. In contrast, rocks along the north Goa coast gently dip without significant schistosity, aligning almost parallel to the coastline in a NW-SE direction. The Western Ghat, extending from NS to NNW-SSE direction, represents a notable fault zone.

GEOMORPHOLOGY

Goa is a land of rolling hills, beaches and parts of Western Ghats. The landforms of Gon exhibit a polycyclic landscape that has been evolved largely during the Cenozoic under humid tropical climatic condition. The geomorphic features and landforms are a consequence of complex interplay of drifting of the Indian plate from the east Gondwana followed by epeirogenic upheaval and subsidence aided by change in climatic conditions and eustatic variation in the sea level. The 100km long coastline of Goa is complex. The coasts of Salcete, Bardez, Pernem are made up long sandy beach stretches whereas the coasts of Canacona-Quepem and Mormugao are crenulated and indented. Coastal sandy plains and littoral terraces less than 10m above mean sea level. Low level tablelands between 10to 100m above mean sea level. Intermediate level tableland, low denudational hills, contiguous ridges with intervening weathered plains 150 to 300m above mean sea level. High level Western Ghats plateau 600 to 700m above mean sea level.

The major rivers flowing through Goa are namely Tiracol, Mandovi, Zuari, Sal, Talpona and Galgibaga. River Mandovi and Zuari are relatively major as compared to others. All the rivers that flow across the state are estuarine and tidal waters penetrate several kilometers inland. Zuari rises in the Western Ghats in Karnataka and flows over a distance of over 70 km until it meets the Arabian Sea at Mormugao. The major tributaries are Guloli, Sanguem and Kushavati.

Telles (1911) and Feio (1956), described the geomorphic problems of Goa paying special attention to the Ghat Escarpment. It is a sinuous border at whose front there is relief which is often nearly as high as the plateau further inland. The escarpment was considered by them as an erosional. They believed that the difference of levels of the order of 1000 m between the

Deccan Plateau and the littoral region indicate a high probability of tectonism. They proposed a fault or a group of faults exist, positions of which not exactly known but which is probably not very far from the present shoreline.

De Souza (1965) physiographically classified the Goa area into three regions: 1) the coastal tract consisting of beaches, rugged sea-cliffs, small isolated pocket beaches, recent broad alluvial plains, vast estuaries, older spits, dunes, beach ridges, broad hard wave cut platforms 2) the tract between the coast and the ghats, ie, the sub-ghats region and 3) the high ranges of the western ghats rising to elevations of 900 to 2500 m above m.s.l.

The drainage network is largely dendritic. The average drainage density in river Zuari basin is 2.1km/ sq km and the average stream frequency is 2.5 stream/sq km. The constant of channel maintenance is 1.0491 sq km. Average relative relief in the Zuari basin is 147.8m. Geostatistical analysis carried out in the Zuari shows that relief is more mature which is supported by gentle slopes in the basin of 5- 6 degrees. The pelitic schist of the Zuari basin may have facilitated the development of these gentle slopes. The morphological condition in the Zuari basin is more mature.

Major lineament that represents fracture zone extends from the village Panchwadi to Suktoli over a length of 22.5 km and controls the course of river Zuari in this stretch. The channel of the river appears more having many curves and turns, although overall controlled by the structure. The apparent curving is related to the discharge and the sediment load which depend on the gradient and therefore on the velocity. The low gradient in Zuari allows accommodation of excess water and sediment load which reduces the erodibility capacity and also leads to deposition which forces the river to follow a meandering course. The meanders are incised indicating that they were formed in paleo-environment and have still persisted (Dessai, 2018).

The Western Dharwar Craton, whichh is part of the Archacan-Proterozoic granite-greenstone region of western peninsular India, is covered in its northern region by the Late-Cretaceous-Early Cenozoic (64-67 Ma) Deccan Traps CFBP.

A rather uniform macro geomorphological landscape has resulted from denudation's effects on the pre-Deccan basement and lavas. This terrain is found, from west to cast, along much of the western peninsular edge.

- The Konkan and Kanara low-lands: a low -lying costal zone characterized by short westerly flowing rivers.
- The Western Ghats: A coast-parallel continental-scale escarpment typically of 600-1000m elevation.
- The summit zone of the Western Ghats: A narrow zone of discrete, elevated mesas in the basalt region, or of ridges and domes in the cratonic basement.
- The Karnataka and Maharashtra plateaus: Areas of extensive elevated upland of 600-800m elevation, and displaying a gentle eastward slope, and characterized by long, easterly or south-easterly flowing rivers.

Since numerous rivers flowing on the east side of the main escarpment empty into the Bay of Bengal at a distance of between 800 and 1000 kilometers, the Western Ghats serve as the primary drainage divide for much of Peninsular India. Although rivers rarely cross this barrier in the Deccan region, some have done so between 14° and 15°N and at a number of other less significant but nonetheless significant locations further south. Where this has happened, they drop the Ghats as a series of breathtaking waterfalls, like Jog Falls (74°49E-14°13'N) in northern Karnataka, and thereby replenish the geographically constrained drainage basins of the quick-moving Konkan -Kanara rivers. However, the majority rivers which cross the coastal lowlands originate near, or at, the Western Ghats face in the both the Deccan Trap and Pre-Deccan basement areas, and have deeply dissected the laterite belt resulting in a mosaic of aerially extensive, but discrete, ‘tablelands’. This intense dissection has conferred a good deal of relief to the coastal lowlands landscape (i.e. up to 180 m) in the form of steep-sided meandering valleys (Sriram and Prasad, 1979).

In Goa, the laterite belt has been further divided into mosaic of smaller “table-land” section over time as a result of drainage development and extension (Sriram and Prasad, 1980). A knickpoint (or change in gradient) occasionally occurs at the base of the Ghats escarpment in the lower 20 to 40 km stretches of the major rivers and estuaries that cut through the laterite tablelands (eg, Tiracol, Chapora, Mandovi, and Zuari) (Dikshit, 1976, Brückner, 1989).

SEA LEVEL CHANGES

Goa’s coastline is a scenic combination of bays and headlands broken by large estuaries of the Mandovi and Zuari rivers. Of the bays, the Baga, Calangute, and Colva are long curved stretches of white beach sands and palm fringes. Successive changes in sea level have resulted in the emergence of diverse coastal topography due to degradation and deterioration. Ahmed (1972) generally classified the west coast of India as a submerged coast. Several features, such as sea cliffs, abrasive platforms, rocky headlands, tidal flats, anastomosing drainage channels, and laterites at death of 20–30 m below mean sea level, suggest a flooded coastline. In contrast, extensive coastlines, the presence of offshore rock islands, sea caves, coastal cliffs, and coastal terraces support recent coastal formation (Dessai 2018).

D’Souza (1968) classified this coast as a ria coast, characterized by wide estuaries, representing a coastline that is largely surfaced and partially submerged. Since the current sea level was set in the mid-Holocene, the coastal morphology mainly shows evolutionary features after the Flemish transgression (Goudie, 1973). The Goa coast has abundant evidence of sea-level changes associated with the Middle Holocene and beyond. Sea-level fall occurred along the paleocoastline, as indicated by shelf-edge ridges of 1–12 m height 7.5 and 110 Ma BC; (Nair and Hashim, 1980; Rao and Nair, 1992).

It is known that sea level during the late Quaternary period (transition from Pleistocene to Holocene) was 6090 meters lower than today’s sea level. This is supported by the presence of high cliffs at sea level, river terraces, laterites, and cliffs along the estuary at depths of 20–34 m below the MSL. (Feio, 1956). The associated deforestation created deep-seated river valleys lined with uplifted laterite plateaus. Submarine laterites near the Cansaulim coast (occurring 10 m below MSL) provide evidence of this regression. This was followed by a gradual rise in sea level, probably from the early Holocene until 6.0 ka BP, reaching a peak of 6 m above modern sea level at about 3.0 ka BP. Cansaulim and Dapoli (3.5-2.6 ka BC; Kumaran et al.,2012) were possibly deposited in fluvio-lacustrine and/or lagoonal environments, broadly contemporaneous with this trangressive phase.

Similar occurrences of clayey and silty sediments, at places carbonaceous, at others with carbonized plant remains, are also encountered in bore wells in other parts of Goa, for example the Sangod iron ore mine near Sanvordem, Dabolim railway station, Goa University Campus, Taleigao, Tivim and many other places. These find support in the lithiiied beach rock that occurs 4-6 m above MSL around Korlai-Borlai and other areas along the Konkan coast, to the north of Goa (Dessai 2018).

Marine shell beds in intertidal regions of Chapora, Mandovi, Zuari, and other rivers suggest ancient sea inlets possibly linked to sea level oscillations between approximately 2.3-1.5 ka BP (e.g., Bruckner, 1989; Karlekar, 1996). These oscillations likely formed tidal flats along the west coast. Features like abrasional platforms, sea stacks, sea caves at Aguada, and beach rocks at Anjuna (1.5 meters above MSL) and Bogmalo (2.5 meters above MSL) suggest past sea level rises. Palaeodunes indicate higher past sea levels, while fossil wood in calcareous Aeolianite, 6.5 meters below dune surfaces, hints at sea level lowering by 1.5-2.5 meters below present-day MSL (Kale, 1983). Fluctuations in sea level are indicated by beach rock and aeolianite, with today's sea level approximately corresponding to levels 6.0 ka BP. However, modern local sea level rise is observed along the west coast, particularly in Goa, likely linked to neotectonic activity and/or eustatic base level changes (Dessai, 2018). The coastal zone has been alternately flooded and exposed during the sea-level fluctuations during the late Quaternary period as well as earlier. In general, lower sea levels were associated with the glacial periods while warmer interglacial periods were characterized by higher sea levels and transgression. During the Last Glacial Maximum (LGM) the sea level along the Indian coastline was lower by more than 120 m. Available evidence suggests that the sea level rose rapidly during the early Holocene flooding low lying areas and penetrated inland via rivers creating estuaries (Bruckner 1988; Hashimi et Al. 1995; Banerjee 2000). Evidence of higher sea levels in the past in the form of raised marine terraces, beach rocks, oyster beds, corals, tidal deposits, etc. have been reported from some parts of Saurashtra, Konkan, Malabar, Coromandal and Andhra coasts. Marine shells, beach rocks and coral reefs occurring ~1-2 m above the high tide level have been reported from Rameswaram and Kanyakumari (Bruckner 1988; Banerjee 2000). Beach rocks at Manori (Mumbai) and Mirya Bhatti (Ratnangi) are present ~3-6 m above present sea level (Agrawal and Guzder 1972).

Palaeoplain	Age	Process/lithology
	Mid-/Early-Holocene	Sea level 1.5-2.5 m below MSL/Clays with carbonized plant remains/Colva fossil wood
	Late-Pliocene/Pleistocene	Low sea level (60-90 m below MSL), shelf- edge ridges, rapid rise in sea level, river terraces, Sea-Level Laterites, Coastal Laterites
Mopa- Verna	Mid- Miocene/Early Pliocene	Low- level Laterites
	Mid- Miocene	Transgressive cycle/shales with carbonaceous streaks e.g. Ratnagiri carbonaceous shales
Unconformity		
Mollem	Late Oligocene/ Early Miocene	Intermediate - level Laterites
	Late Oligocene	Seaward tilt with marine transgressions
Unconformity		
	Early Oligocene	Downwarping of the crust, submergence of Bombay High platform
	Late Eocene	Marine transgression Tarapur and Telwa Shales
Unconformity		
Anmode	Mid Eocene	In-land lagoon Formation/change in drainage/ dissection of High-Level laterites/continental sediments-organic black shales e.g. Cambay shales
	Late Palaeocene/ Early Eocene	Partly inland partly offshore basin formation/ terrigenous clastics on eastern flank of Laccadive ridge, High-Level Laterites

Table 2.2: Summary of the geomorphological evolution of Goa (after Dessai 2018)

PREVIOUS WORK

Sediment dynamics of tropical open coastal beaches of the central west coast of India: Impact of spatiotemporal variations, by Lynda-Uta Edet Okon (2023). Buried channels provide keys for inferring Quaternary stratigraphic and paleoenvironmental changes: A case study from the west coast of India by Dubey and Chaubey (2019). Tidal-related changes in the environmental characteristics of the Zuari and Mandovi estuaries (Dehadrai, P.V.) (1970). Monitoring of some environmental parameters of the Zuari River estuary by De Sousa (1977). Diurnal variations of some physicochemical factors in the Zuari estuary Singbal (1973). Origin and fate of organic matter in suspended solids and sediments of the Mandovi and Zuari estuaries of western India, by R. Shynu. Kessarkar (2013) Geochemistry of suspended sediments in the Mandovi and Zuari river estuaries of the central west coast of India.

Seasonal-scale suspended sediment dynamics in Mondovi and Zuari estuaries, Purnachandra Rao, central west coast of India (2010). Vulnerability of mangroves to changing coastal regulation areas: A case study of Mandovi and Zuari rivers in Deshpande, Goa. (2023)

Clay minerals in determining the origin of sediments in the Mandovi estuary, Goa, west coast of India by Bhukhari and Nayak (1996). Design of marine sediment traps and accessories by Janakiraman G. et al. (1986). Textural characteristics of Mormugao Gulf sediments, by Rao, D.G.; Rao, T.C.S. (1974). Anomalous behavior of uranium isotopes in backwater sediments of the Zuari River by Joshi, L.U. et al. (1983). Changes in the physical properties of water in the Zuari Estuary, Cherian, T. et al (1975).

CHAPTER III : METHODOLOGY

STUDY AREA

Latitude: 15°23’15”N

Longitude: 73°55’11”E

The area being examined is situated between 16 km and 20 km south of Goa University. It is bounded by Sancoale in the west, Agacaim in the north, Nagoa and Verna in the south and Madkai in the east. The primary towns near the study area are Panjim (16 km) and Vasco (14 km) which are connected to the National Highway (NH-66). The site is approachable by all weather road. The region experiences tropical monsoon climate.The study area is slope at Sotrant. Humid climate, abundant precipitation, large source of plant debris, retardation of organic decomposition and low influx of inorganic sedimentary material are some of the prerequisites which promote vigorous plant growth, with consequent peat accumulation. Avicennia officinalis L. species of mangroves is profuse. In the study area, mangroves are the most remarkable and the only intertidal trees which inhabit bays and lagoons influenced by sea water. (Mascarenhas and Chauhan, 1998). The country rock is encountered at a depth of 30m.

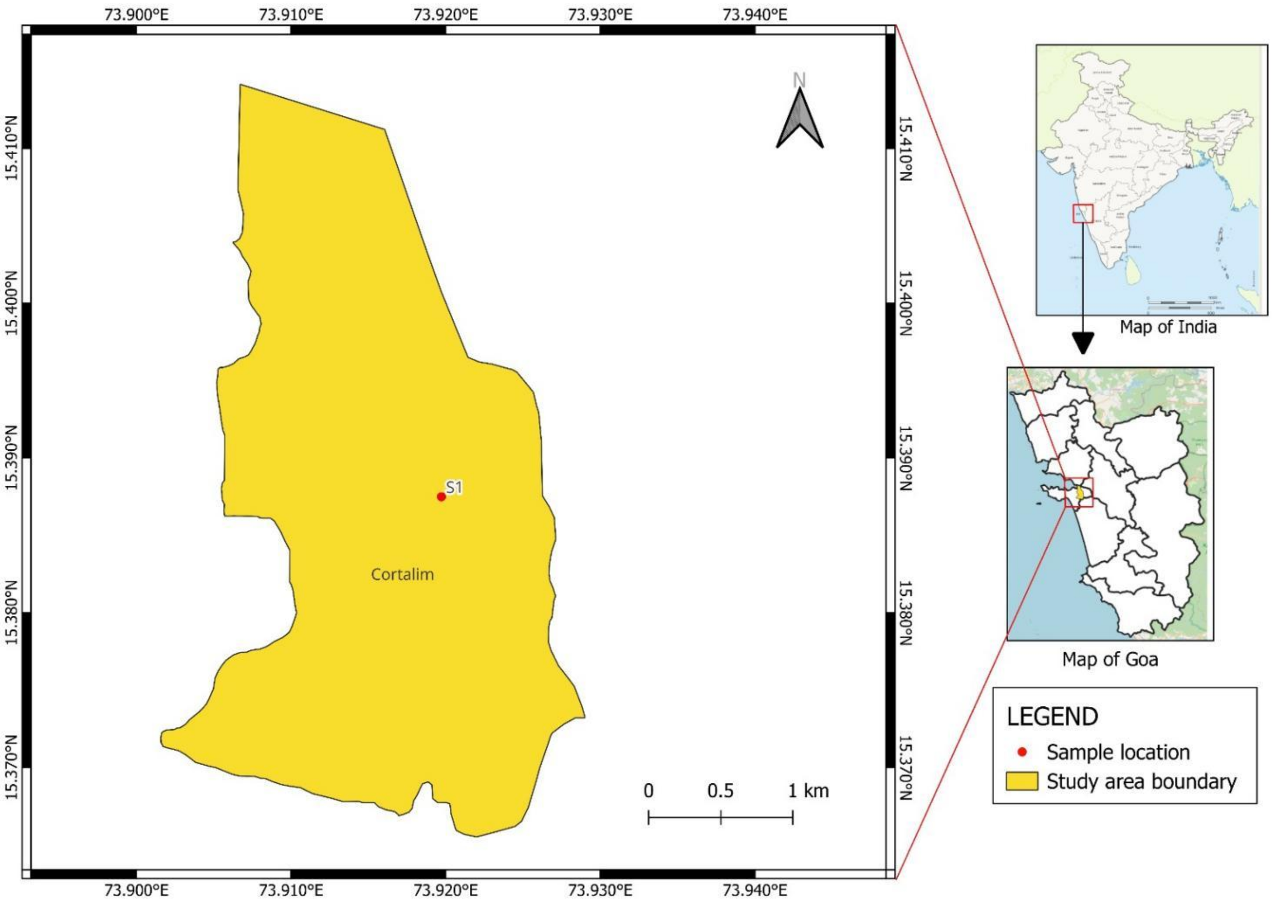


Figure 3.1: Map of the study area.

GRAIN SIZE

Granulometry is a basic analytical technique that has wide applications within the earth and archaeological sciences. Particle or grain size is a fundamental attribute or physical property of particulate samples or sediments and sedimentary rocks (Folk, 1980; Friedman and Sanders, 1978). Much can be said from analyzing not only the size of clastic or detrital (inorganic), bioclastic (organic), or chemical particles but also from the overall size distribution, size fraction percentages, textural maturity of the sediment or sorting, surface texture attributes of a particle, and sphericity/angularity and shape of a particle (Krumbein and Sloss, 1963; Syvitsky, 2007). Several sediment, soil, or material properties are directly influenced by the size of its particles, as well as their shape (form, roundness and surface texture or the grains) and fabric (grain-to-grain interrelation and grain orientation), such as texture and appearance, density, porosity, and permeability.

The size of particles is directly dependent on the type of environmental setting, transporting agent, length and time during transport, and depositional conditions, and hence it possesses significant utility as an environmental proxy (McManus, 1988; Stanley-Wood and Lines, 1992). Grain size is related to a multitude of external factors acting on a local or regional scale. For example, in the coastal and marine setting, grain size is related to the bathymetry and geometry of the basin, nutrient regime, biogeochemical oceanography, coastal processes, net sedimentary inputs from land sources, and outputs.

Traditionally, sediments were divided into three principal categories: gravel, sand, and mud. The latter is further divided into silt and clay, mostly based on mineralogical distinction rather than (hydro-)dynamic properties. Since the early 1900s, standardization of such size ranges has been defined based upon different grade scales constrained by particle size limits or range boundaries. The size of the particles is based on their nominal diameter, traditionally reported in millimeters (mm), micrometers (µm) or phi (Ø) units. The Wentworth or Udden-Wentworth scale (Udden, 1914; Wentworth, 1922) divides the size ranges into textural classes with specific terminology, from boulders (> 200 mm) to clay (< 0.004 mm). It is a geometric scale in which each size limit is ½ or twice the millimetre value of the next. The Krumbein Ø scale (Krumbein and Sloss, 1963) is a logarithmic scale modified from the Udden-Wentworth one and based on conveniently calculated round values, to avoid dealing with mm fractions . Classification of detrital sediments is based upon the quantification of, or relationship between, the proportions/percentages/ratios of different particle size fractions or textural classes within a mixed sediment (Shepard, 1954; Folk, 1980).

GRAIN SIZE SCALES

Geometric scales:

The grade scale most widely used by sedimentologists is the Udden–Wentworth scale (Wentworth, 1922). Each value in this scale is two times larger than the preceding value. The Udden–Wentworth scale extends from< 1/256 mm (0.0039 mm) to > 256mm and is divided into four major size categories (clay, silt, sand, and gravel).

Logarithmic phi scale:

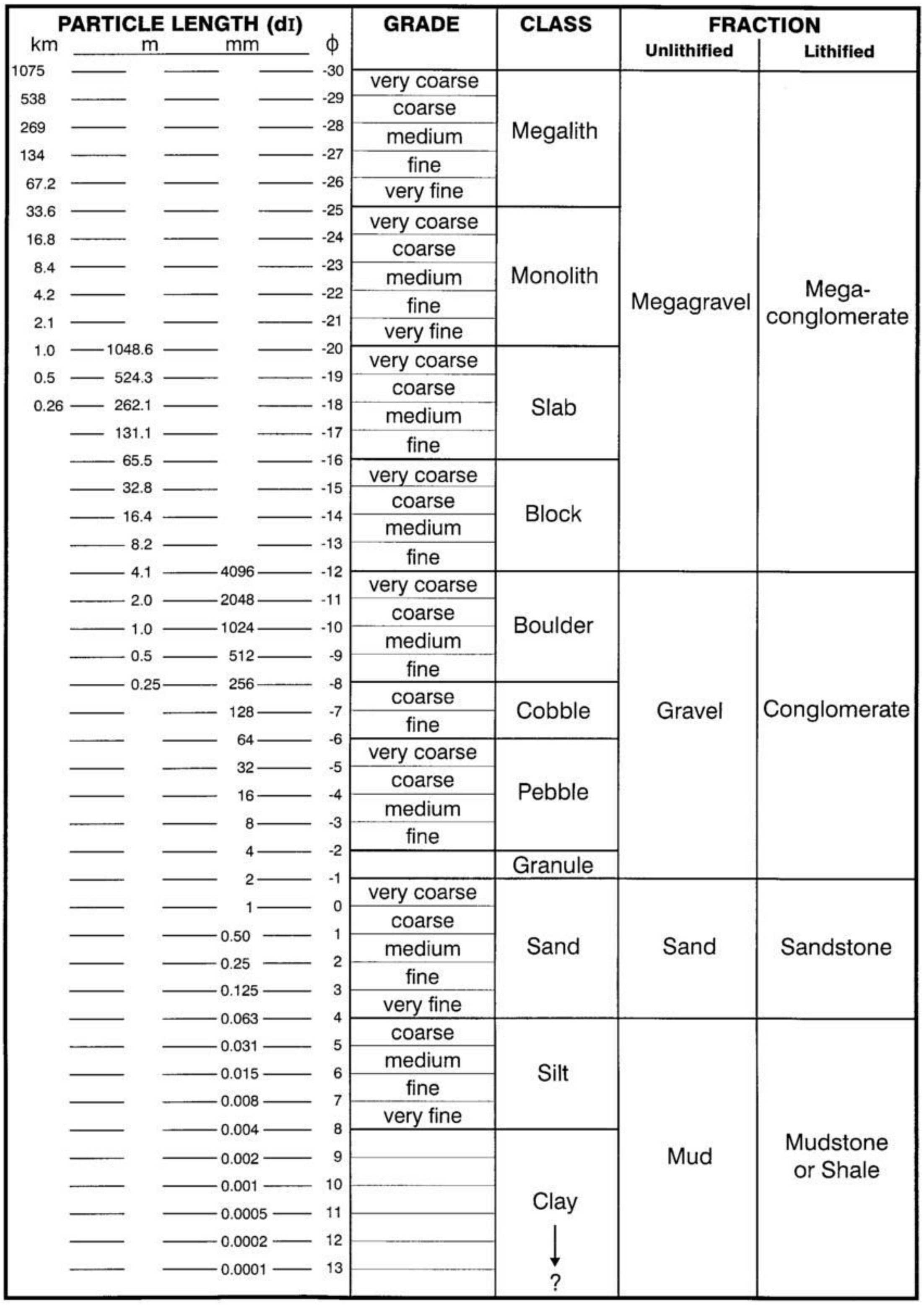
A modification of the Udden–Wentworth scale is the logarithmic phi (ϕ) scale proposed by Krumbein (1934). This scale, is based on the following relation:

Phi (ϕ) = -log2 d

where d is the diameter (mm).

The principle advantage of the (ϕ) scale are:

1. The phi (ϕ) scale allows grain-size classes to be expressed in integers (size class become whole numbers instead of fractions).
2. The phi (ϕ) scale simplifies statistical calculations and graphic plotting of grain size data.
3. This scale is consistent with common practice of plotting coarse sizes to the left and fine sizes to the right in the graphs.



PARTICLE LENGTH (d_I) km	m	mm	ϕ	GRADE	CLASS	FRACTION Unlithified	FRACTION Lithified
1075			-30	very coarse	Megalith	Megagravel	Mega-conglomerate
538			-29	coarse			
269			-28	medium			
134			-27	fine			
67.2			-26	very fine			
33.6			-25	very coarse	Monolith		
16.8			-24	coarse			
8.4			-23	medium			
4.2			-22	fine			
2.1			-21	very fine			
1.0	1048.6		-20	very coarse	Slab		
0.5	524.3		-19	coarse			
0.26	262.1		-18	medium			
	131.1		-17	fine			
	65.5		-16	very coarse	Block		
	32.8		-15	coarse			
	16.4		-14	medium			
	8.2		-13	fine			
	4.1	4096	-12	very coarse	Boulder	Gravel	Conglomerate
	2.0	2048	-11	coarse			
	1.0	1024	-10	medium			
	0.5	512	-9	fine			
	0.25	256	-8	coarse	Cobble		
		128	-7	fine			
		64	-6	very coarse	Pebble		
		32	-5	coarse			
		16	-4	medium			
		8	-3	fine			
		4	-2		Granule		
		2	-1	very coarse	Sand	Sand	Sandstone
		1	0	coarse			
		0.50	1	medium			
		0.25	2	fine			
		0.125	3	very fine			
		0.063	4	coarse	Silt	Mud	Mudstone or Shale
		0.031	5	medium			
		0.015	6	fine			
		0.008	7	very fine			
		0.004	8		Clay ↓ ?		
		0.002	9				
		0.001	10				
		0.0005	11				
		0.0002	12				
		0.0001	13				

Table 3.1: Modified Udden Wentworth grain size scale (Wentworth 1922)

MEASURING GRAIN SIZE

Unconsolidated Sediments: Boulder, cobble and pebble are manually measured with a caliper or tape. Granule to silt-size particles are commonly measured by sieving through a set of nested, wire-mesh screens. Standard sieves of different mesh sizes that are kept one over, coarser at the top and finer downwards are mechanically shaken. The amount trapped in each sieve is weighed and taken as an estimate of the particle size range in the sample. Finer sediments like silt and clay are more difficult to measure. The commonest method is elutriation, where a sample is treated with a dispersant and placed in cylinder (settling tube). The finest fractions stay in suspension longest, whereas the larger/heavier fractions generally sink more rapidly. Size is measured by determining their settling.

Consolidated Sediments: Consolidated granules and hard indurated sandstones are measured by thin-section method. Grains are visually counted and measured under petrological microscope with an eyepiece having a graduated scale and point counter. Size measurement of silt and clay in consolidated rock is possible by using a high magnification electron microscope. We can also use digital image processing method for grain size measurement.

Gravel and Sand size:

Measurement of sand and gravel grains are done by means of mechanical sieving. Sieves are screens of standard size and specified by the American Society of Testing Materials (A.S.T.M). For size measurement, a bank of sieve is taken with downward decreasing opening (mesh) size . Sediment with measured weight is kept on the topmost sieve and then the sieve bank is mechanically or electrically shaken. Depending on the size of grains, the sediment get distributed among the sieves used; grains larger than the sieve size will remain on the sieve and grains with smaller size pass on to the sieves further down. By weighing fraction of sample lying on each sieve, a grain size distribution curve is constructed. The lower limit of sieve analysis is 0.04-0.03 mm, grains finer than this size show large amount of cohesion and do not allow measurement o individual grain by this method.

Silt and Clay size fraction:

As discussed earlier, sediments of this size fraction cannot be measured by seive analysis because of their high cohesive character. In lieu, measurement of size for these sediments are done through number of other methods, most of which deals with settling velocity of sediment grain in liquids following Stoke's law.

Hydrometer:

A suspension of sample is created with a mixer in a cylinder so that at time T_0 there is an even distribution of grains in the solution and density of solution is same throughout the cylinder. A hydrometer will float in the solution and register density of the solution on a scale on the upper part of the tube. If the suspension is allowed to stand for a time T_1, grains will settle in the bottom of the cylinder. At the level at which the bulb of the hydrometer is floating the largest of the grain sizes already must have settled resulting decrease in density of the solution. This reduction in density of solution gets registered by further submergence of hydrometer and recorded in the scale at the upper part of tube. By taking successive readings of the density we plot a density variation curve with respect to time. Since density reduction is a function of settling velocity, the density variation curve can be calibrated as a grain-size distribution curve. In case the sediment contains a large clay fraction, use of centrifuge may be a more desired option. The acceleration term g in the Stoke's law increased in the centrifuge, which can be calculated from the velocity of rotation and length of rotating arm.

Clay separation

Clay particles can also be separated by sedimentation techniques on the basis of the settling velocity of the particles. In these techniques, grains are allowed to settle through a column of water at a specified. Temperature in a settling tube, and the time required for the grains to settle is measured. For coarser particles (granules, sand, silt), the settling time of the particles is related empirically to a standard size-distribution curve (calibration curve) to obtain the equivalent millimeter or phi size. Settling velocity of particles is affected by particle shape. Spherical particles settle faster than non-spherical particles of the same mass. Therefore, determining the grain sizes of natural, non-spherical particles by sedimentation techniques may not yield exactly the same values as those determined by sieving. The grain size of fine silt and clay particles can be determined by sedimentation methods based on Stokes's Law. Stokes law pertains to the terminal fall velocity of a sphere in a fluid. If the settling velocity is known for a particular temperature, diameter (D) of the sphere can be calculated as

$D = \sqrt{V} \sqrt{C}$

V is the terminal velocity of the sphere.

Where C is a constant equaling $C = (ds-df)/ 18\mu$

ds = 2.65 gm/cm3 (density of quartz)

df= the density of distilled water at the particular temperature

g= 980 cm/sec and μ= the viscosity of distilled water at a particular

Stokes law is only valid under the following conditions and limitations:

- Particle must have reached terminal fall velocity

- Particles must be rigid
- Particles must be smooth
- No slippage or shear may take place between the particle and the fluid
- The fluid must be of infinite extend in relation to the particle
- Particle concentration must be less than 1%
- Particle must be greater than 0.5 micron in diameter
- Particle must be sphere (Carver, 1971)

Sedimentation Balance:

In this method sediments suspended in a solution in a cylinder fall through the water column and settle on a balance pan kept at any depth level within the cylinder. The balance pan is kept connected with a weigh meter. With passing time more and more sediment will get settled on the pan and the increasing weight of the pan get recorded in the weigh meter. The increase in weight is plotted against time, which indicated suspension settlement from solution as a function of time. This gives a direct cumulative curve that can be calibrated to give a grain-size distribution curve.

CLAY MINERALOGY

Georgius Agricola (1494–1555), the founder of geology, was seemingly the first who gave the definition of clay in 1546. According to Joint Nomenclature Committees (JNCs) of the Association Internationale pour l'Etude des Argiles (AIPEA) and the Clay Minerals Society (CMS), clay, a naturally occurring material, composed mainly of fine-grained minerals, become plastic in presence of water and become hard when dried or fired. By this definition of clay, engineered clays and clay-like materials can be distinguished as clay (fine grained minerals) exhibiting plasticity in presence of water and become hard on drying and firing.

Clay is a soft, freely bound, fine grained natural rock or earthy material having diameter less than 0.005 mm and composed essentially of clay particles. Based on the standard definition of mineral, clays are mainly inorganic materials except peat, muck, some soils, etc. that contain huge amount of organic/natural materials. The clay particles are formed due to the weathering and erosion of rocks containing soil, ceramic clays, clay shales, glacial clays (including great volume of detrital and transported clays) the mineral group feldspar (known as the 'mother of clay') over vast spans of time. During weathering, the content of feldspar is distorted by hydrolysis process results in formation of clay minerals such as kaolinites the primary minerals in kaolin clays and smectite the primary minerals in bentonite clays. Clay can incorporate with one or more clay minerals even in presence of minute quantities of quartz (SiO2), metal oxides (Al2O3, MgO) and organic matter. The plasticity of clays are due to their particle size, geometry as well as content of water and become hard, stiff, coherent and non: plastic upon drying or firing. Plasticity and hardness are greatly affected by the chemical composition of the material present in the clay. Clays can be molded in any form when they retain water. For example, some species of chlorite and mica are found to be non-plastic while grinding macroscopic flakes even where more than 70% of the material is <2 µm esd (equivalent spherical diameter). Whereas some species of chlorites and micas become plastic on grinding the macroscopic flakes where 3% of the materials is <2 µm . Clays are easily molded into a form that they retain when dry, and they become hard and lose their plasticity when subjected to heat.

In all definition of clays, the particle size is a key parameter, no generally upper limit is accepted till now. Although clays can be distinguished from other fine-grained soils on the basis of their size difference and mineralogy. The particle sizes of silts (fine-grained soils that do not consist of clay minerals) is larger than clays. Individual clay particles are always smaller than 0.004 mm. The difference between silt and clay varies by discipline. Geologists and soil scientists usually consider a particle size of 2 µm (clays being finer than silts) for the separation, sedimentologists apply 4–5 µm, and colloid chemists use 1 µm. According to Geotechnical engineers, differentiation between silts and clays can be done on the basis of the plastic characteristics of the soil, as measured by the soils' Atterberg limits. The combination of silts, sand and clay (<40%) are called loam.

ORIGIN OF CLAY MINERALS

Clay minerals have their origin in the weathering and alteration of rocks and minerals over long periods of time. The formation of clay minerals is a complex process that involves the interaction of various geological and chemical factors. Here's a general overview of the origin of clay minerals:

Weathering: The process begins with the weathering of primary minerals, such as feldspars, micas, and volcanic glass, which are present in rocks. Weathering occurs due to the exposure of rocks to various physical, chemical, and biological agents, including temperature fluctuations, water, wind, and organisms. These agents break down the rocks into smaller particles and initiate the transformation of primary minerals into clay minerals.

Chemical alteration: During weathering, water acts as a solvent, infiltrating the rocks and facilitating chemical reactions. The water interacts with the minerals, leading to ion exchange, dissolution, and precipitation reactions. The breakdown of primary minerals releases ions into the water, which can react with other dissolved species to form new minerals, including clay minerals.

Hydrolysis: One of the key processes in clay mineral formation is hydrolysis. Hydrolysis occurs when water molecules react with the silicate minerals, breaking the chemical bonds and altering the mineral structure. This process releases ions such as silicon, aluminum, and various cations into the solution, which can later participate in the formation of clay minerals.

Ion exchange: As the primary minerals weather and release ions, these ions can be exchanged with other ions present in the environment. For example, cations like calcium, sodium, and potassium can be exchanged with hydrogen ions in the presence of water. This ion exchange process further contributes to the formation of clay minerals.

Sedimentation and digenesis: The weathered particles, including the newly formed clay minerals, can be transported by wind or water and eventually settle as sediments. Over time, these sediments can undergo compaction and lithification, transforming into sedimentary rocks. The process of compaction and lithification, known as diagenesis, plays a crucial role in the preservation and transformation of clay minerals. It’s important to note that there are various types of clay minerals, including kaolinite, montmorillonite, illite, and others. The specific conditions and mineral compositions during the weathering and alteration process influence the type and characteristics of the resulting clay minerals

OCCURENCES OF CLAY

Sedimentary rocks only make up 5% of the Earth's crust, but cover about 80% of the surface of the earth in which clays (including shales) form well over 40% of the sedimentary rocks. The raw material for sedimentary rocks comes from weathering. If we look at the volume of material at the earth's surface, we see that clay minerals constitute about 16% of its total. 20 km is considered the surface of the earth because it is the region from which we extract natural resources(and dump our waste). Clay sediments are collected by the agencies of water (e.g. marine clays, alluvial clays, lacustrine clays), wind (Aeolian clays), or ice (e.g. glacial clay, till or boulder clay, as most clays in Finland). The majority of the common sedimentary clays, however, are the marine deposits typically comprising mixtures of coarser material with clay in which the clay mineral, illite, usually predominant.

Clays are divided into two classes:

1. Residual clay: Residual clays are found in the place of origin and formed by surface weathering which gives rise to clay in three ways:
    - Chemical decomposition of rocks, such as granite, containing silica and aluminia
    - Solution of rocks, such as limestone, containing clayey impurities, which, being insoluble, are deposited as clay
    - Disintegration and solution of shale

2. Transported clay: also known as sedimentary clay, removed from the place of origin by erosion and deposited in a new and possibly distant position

CLASSIFICATION OF THE CLAY MINERALS

Allophane: Allophane is a series name used to describe clay-sized, short-range ordered aluminosilicates associated with the weathering of volcanic ashes and glasses. Allophane commonly occurs as very small rings or spheres having diameters of approximately 35 – 50 Å. This morphology is characteristic of allophane, and can be used in its identification. Allophanes have a composition of approximately Al2Si2O5·nH2O.

Kaolin Minerals: The kaolin group minerals comprise kaolinite, nacrite, dickite and halloysite, and are among the most common clay minerals in nature. They have a 1:1 layered structure, that is, each layer consists of one tetrahedral silicate sheet and one octahedral sheet, with two-thirds of the octahedral sites occupied by aluminum. Kaolinite, nacrite and dickite all have the ideal chemical composition: Al2Si2O5(OH)4 and the theoretical chemical composition is SiO2, 46.54%; Al2O3, 39.50%; and H2O, 13.96%.

Kaolinite differs from the other three members of the group by including molecular water in the inter layer Kaolins sorted into a size range of 0.5 – 2.0 mm typically have a specific surface of about 5 m2 g-1 and a cation exchange capacity of 10meq./100 gm or less. These and other properties, such as opacity and viscosity, make kaolins suitable for a wide range of uses including paper coatings and fillers, pottery, porcelain and sanitary ware production and fillers in paints and rubbers.

Kaolinite: This clay mineral is the weathering product of feldspars. It has a white, powdery appearance. Kaolinite is named after a locality in China called Kaolin, which invented porcelain (known as china) using the local clay mineral. The ceramics industry uses it extensively. Because kaolinite is electrically balanced, its ability of adsorb ions is less than that of other clay minerals.

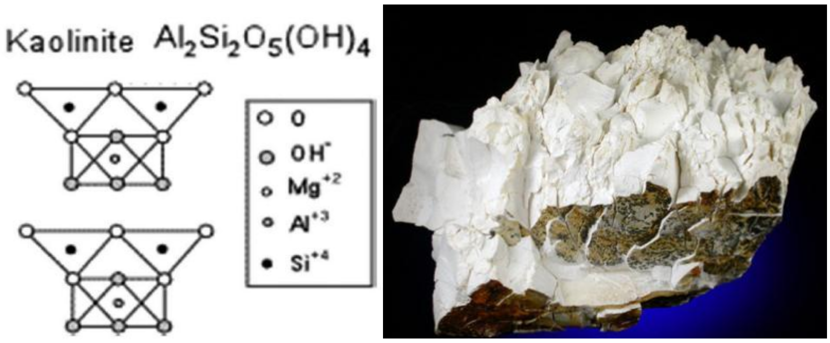


Figure 3.2 : Kaolinite structure and mineral

Dickite and nacrite: result from hydrothermal or pneumatolytic alteration (pneumatolytic is the process by which ores and minerals are formed from the action of vapours produced by igneous magmas). They are rare in clay materials and virtually of no importance commercially. They differ from kaolinite in the way the silicate units are stacked.

Halloysite: is similar to kaolinite but the microscopic crystalline particles are elongate rather than equidimensional in shape. Halloysite occurs in two forms: one hydrated, in which there is a layer of water molecules between the layer, and one dehydrated. The hydrated form has a basal spacing of 10 Å and dehydrated form, 7.2 Å.

Smectite Minerals

Smectite: This clay mineral is the weathering product of mafic silicates, and is stable in arid, semi-arid, or temperate climates. It was formerly known as montmorillonite. Smectite has the ability to adsorb large amounts of water, forming a water-tight barrier. It is used extensively in the oil drilling industry, civil and environmental engineering (where it is known as bentonite), and the chemical industry.

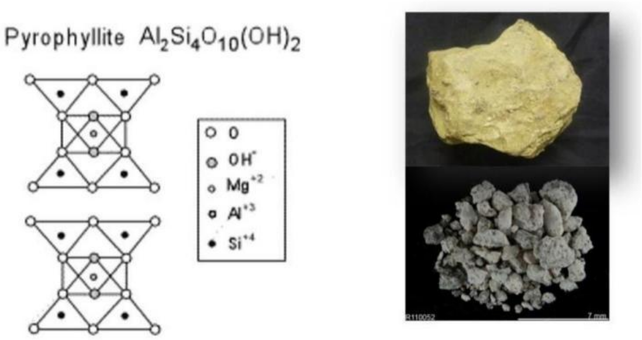


Figure 3.3 : Smectite structure and mineral

There are two main varieties of smectite, described in the following: Sodium smectite: This is the high-swelling form of smectite, which can adsorb up to 18 layers of water molecules between layers of clay.

Calcium smectite: The low-swelling form of smectite adsorbs less water than does sodium smectite, and costs less. Calcium smectite is used locally for drilling muds.

The major smectite minerals are Na-montmorillonite, Ca-montmorillonite, saponite (Montmorillonite), nontronite (Fe-montmorillonite), hectorite (Li-montmorillonite), and Beidellite (Al-montmorillonite). Smectite minerals are composed of two silica tetrahedral sheet with centraloctahedral sheet and are designated as a 2:1-layer mineral (Fig. 12). Water molecules and cations occupy the space between the 2:1 layers. The theoretical formula is (OH)4Si8Al4O20.NH2O (interlayer) and the theoretical composition without the interlayer material is SiO2, 66.7%; Al2O3, 28.3%; and H2O, 5%. However, in smectite, there is considerable substitution in the octahedral sheet and some in the tetrahedral sheet. If the octahedral positions are mainly filled by Al, the smectite mineral is be Illite; if filled by Mg, the mineral is saponite; and if by Fe, the mineral is Nontronite. The most common smectite mineral is Ca-montmorillonite, which means that the layer charge deficiency is balanced by the interlayer cation calcium and water. The basal spacing of the Ca montmorillonite is 14:2Å.

Na-Montmorillonite occurs when the charge deficiency is balanced by sodium ions and water and basal spacing is 12.2Å. Ca-montmorillonite has two water layers in the interlayer position and Na-montmorillonite has one water layer. The smectite mineral particles are very small and because of this, the X-ray diffraction data are sometimes difficult to analyze.

Illite minerals

Resembles muscovite in mineral composition, only finer-grained. It is the weathering product of feldspars and felsic silicates. It is named after the state of Illinois, and is the dominant clay mineral in mid-western soils. The structure is a 2:1 layer in which the interlayer cation is potassium. The size, charge, and coordination number of K is such that it fits snugly in hexagonal ring of oxygen of the adjacent silica tetrahedral sheets. This gives the structure a strong interlocking ionic bond which holds the individual layers together and prevents water molecules from occupying the interlayer position as it does in the smectite. Illites, which are the dominant clay minerals in argillaceous rocks, form by the weathering of silicates (primarily feldspar), through the alteration of other clay minerals, and during the degradation of muscovite (Deer etal., 1975). Formation of Illite is generally favored by alkaline conditions and by high concentrations of Al and K

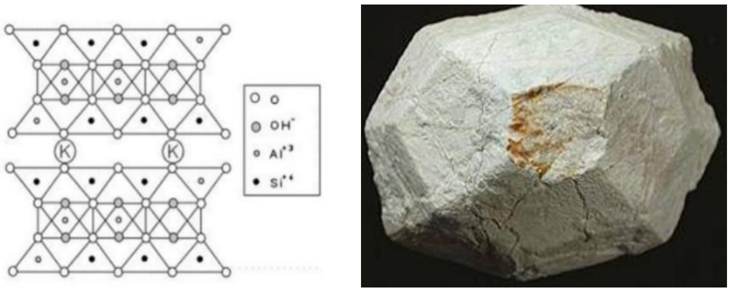


Figure 3.4: Illite structure and mineral

Chlorite Minerals

This clay mineral is the weathering product of mafic silicates and is stable in cool, dry, or temperate climates. It occurs along with illite in mid-western soils. It is also found in some metamorphic rocks, such as chlorite schist. The chlorite group members contain a 2:1 layer with variable x and an interlayer hydroxide sheet. The octahedral sheets may both be dioctahedral (di/di) or trioctahedral (tri/tri), or mixed (di/tri, or tri/di). The interlayer hydroxide sheet may have a positive charge. Dioctahedral chlorite is dioctahedral in both the 2: 1 layer and the interlayer hydroxide sheet. An example is donbassite (Brindley and Brown, 1984). Trioctahedral Chlorites should be named according to the dominant divalent octahedral cation present.

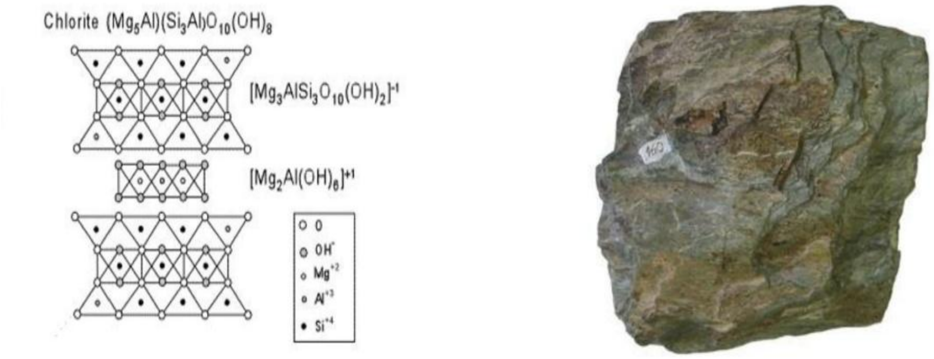


Figure 3.5: Chlorite structure and mineral

CLAY WATER SYSTEM

In the natural environment, clays are often formed under water. In water, positively charged ions (cations) and positive edges of dipoles (water molecules) are attracted to the clay surface. Several layers of water molecules are attracted on the clay surface in a very ordinary way. The water layer with thickness of 10 angstrom is called adsorbed water layer. The layer is very rigidly, electrically attracted to the clay surface and behaves as if a part of the particle itself. In the outer part of water, there are distributions of mobile cations and anions. Those cations and anions are from resolved minerals and other matters in natural water. Cations further attract dipoles around them. This type of water is called electrostricted water, and they move together when a cation moves. The rest of the space is filled with regular water, which is called free water. There is a boundary within which a clay particle has an influence electricity. The layer extending from the clay particle surface to the limit of attraction is known as Diffuse Double Layer.

X-RAY DIFFRACTION

X-rays are a form of electromagnetic radiation, X-ray wavelengths are 10^{-2} to 10^{-12} m long wavelength X-rays grade into ultraviolet light; shorter wavelengths grade into cosmic and gamma rays. Mineralogists usually give X-ray wavelengths in angstroms (1A equals 10^-10 m). The copper radiation commonly used in X-ray studies has λ=1.5418A. Because of their short wavelengths and high frequencies, X-rays have high energy compared with visible light and most other forms of electromagnetic radiation. High energy allows X-rays to penetrate many natural materials, as observed by Rontgen in 1895. X-rays of highest energy, called hard radiation, are used in many manufacturing and industrial applications, such as checking steel for flaws. X-rays of relatively low energy, called soft radiation, are used by numerologists and for medical diagnoses.

As the target electrons return to a lower energy level, they emit energy as X-rays. The energy difference between the two levels is proportional to the energy and frequency, and inversely proportional to the wavelength, of X-rays emitted. Because some electrons are elevated to higher levels than others, and because they do not all return to the same levels, typical X-ray tubes emit characteristic radiation having several different wavelengths. The wavelengths of the characteristic radiation depend on the metal in the target of the X-ray tube. So, X-ray tubes emit polychromatic radiation, radiation having a range of wavelengths, but most of the energy is channeled at specific wavelengths. We designate the different characteristic X-ray wave lengths using combinations of English (K, L, M) and Greek letters. The most intense peaks are designated Ka (Dexter Perkins 2002). Copper Ka radiation is used for most routine X-ray studies. Copper has several characteristic wavelengths, but interpretation of X-ray diffraction is easiest if we use one wavelength (monochromatic radiation). To isolate Ka radiation from the other wavelengths, most X-ray machines have filters, monochromators, or solid-state monochromatic detectors. X-ray tubes emit two nearly equal wavelengths of Cu - Ka radiation Ka_1 = 1.5401A and Ka_2 = 1.5443A. They are so similar that, even if both wavelengths are present, for most applications the radiation is effectively monochromatic, and we take a weighted average of Ka_1 and Ka_2 wavelengths and assume a λ value of1.5418A.

Von Laue (age) derived equations that describe diffraction by three-dimensional structures call these the Laue equations. W. L. Bragg developed a simpler and more easily understood mathematical treatment. His final equations are just as valid as von Laue's, but the derivations avoided some unnecessary complexities. Although diffraction and reflection are two different processes, Bragg noted that diffracted X-rays behave as if they were reflected from planes within a crystal. To model this "reflection," Bragg considered two parallel planes of atoms separated by distance d (hkl). ABC and DEF represent monochromatic parallel X-ray beams striking and reflecting from the planes. The angle of incidence and the angle of reflection are both ϴ. Because path lengths of the beams must vary by an integral number of wavelengths if diffraction is to occur, the sum of distances must equal n(λ), So $N\lambda = 2d\sin(\Theta)$.

For conventional X-ray diffraction studies, we assume the n in the Bragg law is 1. Because first-order diffraction by planes with spacing d occurs at the same angle as second-order diffraction by a set of planes spaced twice as far apart, we cannot distinguish the two, and assuming first order is simplest when we are talking about X-ray diffraction. In other applications of the Bragg Law, the order of diffraction is important. In standard X-ray diffraction studies, we allow the incident X-ray beam to strike the sample at many different angles so that many different d-values will satisfy the Bragg law and cause diffraction. After being diffracted, the X-ray beam travels at an angle of 2ϴ from the incident beam. Because a crystal contains many differently spaced planes with atoms on them, diffraction occurs at many 2Θ angles. In most X-ray devices used today, the sample is at a fixed location (although it may rotate), and the detector, which measures Xray intensity, moves through an entire range of angles from near 0 to some high angle. In some diffractometers, the X-ray tube moves as well. If a diffracted beam is to hit the detector, two requirements must be met:

A family of planes with d (hkl) must be oriented at an angle (Θ) to the incident beam so the Bragg law is satisfied.

The detector must be located at the correct angle (2Θ) from the incident beam to intercept the diffracted X-rays. Because the angle between the detector and the X-ray beam is 2Θ, mineralogists usually report X-ray data in terms of 2Θ (Dexter Perkins 2002).

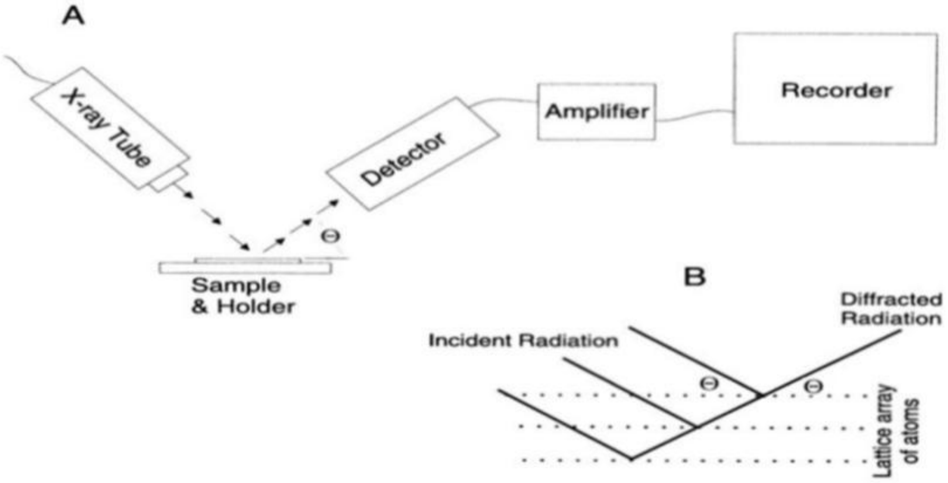


Figure 3.6: Schematic diagram for XRD analysis

INTENSITY OF DIFFRACTION

Diffraction at different Angles: The Bragg law allows us to calculate the angle at which diffraction occurs for any particular d value. However, intense diffraction will only occur if many atoms occupy the (hkl) planes; without atoms no electrons are present to scatter X-rays as the locations and natures of atoms are different in different minerals, diffraction patterns depend on atomic arrangement and composition. But symmetry also affects diffraction patterns. High crystal symmetry means identical spacing of atoms in multiple directions, so fewer d-values cause diffraction, and thus diffraction occurs at fewer angles with relatively uniform intensity. Low crystal symmetry means diffraction occurs in many directions, with variable intensity because atoms and atomic spacing differ in all directions. X-ray diffractometer generally use monochromatic radiation and a solid state X-ray detector instead of film. The diffractometer rotates the crystal and moves the detector to measure intensities of all diffraction peaks. Computers controlling the diffractometer store data directly on disk. Because the process is completely automatic, the d-values and intensities of thousands of different peaks may be measured for a single crystal. Single-crystal X-ray studies provide information necessary for structure determination. Simpler techniques exist for mineral identification). The most common is powder diffraction, the X-raying of a finely powdered sample mounted on a slide or in a holder. A powdered sample consists of a near infinite number of small crystals in random orientations. X-ray diffractometer generally use monochromatic radiation and a solid state X-ray detector instead of film. The diffractometer rotates the crystal and moves the detector to measure intensities of all diffraction peaks. Computers controlling the diffractometer store data directly on disk. As the process is completely automatic, the d-values and intensities of thousands of different peaks may be measured for a single crystal. We use the data obtained from single-crystal diffraction studies to determine crystal structures. If we know the composition of a mineral, we know how many atoms of which elements are present.

Computer programs can determine where atoms are in the structure, based on the diffraction data. We call this process a crystal structure determination. Crystal structure determination involves many complexities, but automation and high-speed computers have simplified the process considerably since the Braggs determined the first crystal structure in 1913. Mineralogists originally used film to record powder diffraction patterns (but today most mineralogists use an X-ray powder diffractometer instead of a camera. We mount the powdered sample on a slide or in a well in a sample holder). A focused X-ray beam strikes the sample, and the detector moves in a circular arc, usually vertically, to measure diffraction intensities for 2(Θ) values from very low angles to more than 150°. In practice, most diffractometer cannot measure peaks at angles below 1 or 2° 2Θ because the direct X-ray beam bombards the detector. The upper 2Θ limit of measurement usually depends on the need of the mineralogist, but for most purposes we do not need data above 60° to 70° 2Θ.

In the past, diffractometer was connected to strip chart recorders to produce diffraction patterns that typically contained 10 to 50 identifiable peaks. A mineralogist would then measure the diffraction angles on the chart and calculate d-values. Today we connect diffractometer to computers that can analyze data and print diffraction patterns. For simple mineral identification, powder diffractometer connected to computers generally require a few to 30 minutes. For extremely precise results, however, data collection may take hours or even days (Dexter Perkins 2002).

IDENTIFICATION OF MINERALS FROM XRD DATA

Powder patterns are not normally used to determine crystal structures because powder diffraction data are more difficult to interpret than single crystal data. Different X-ray peaks correspond to different crystals in the powdered sample, and the data do not reveal the orientations of planes causing diffraction. A relatively new technique for interpreting powder patterns, called the Rietveld method, sometimes overcomes these complications and is gaining popularity today. However, powder patterns do yield a list of d-values and peak intensities that mineralogists routinely use to identify minerals. Each d-value corresponds to sets of planes, and the intensity is a measure of how many atoms are on those planes. Because each mineral has a different atomic arrangement, each yields a different pattern.

For mineral identification, we compare measured d-values and intensities to reference data sets. The most complete reference, the X-ray Powder Diffraction File (PDF) compiled by the International Centre for Diffraction Data, is available in computerized databases and contains in Formation on almost 6,000 minerals and tens of thousands of other inorganic and organic crystalline compounds. Different samples, different methods of sample preparation, and many other things can lead to deviations. d- values corresponding to strong peaks, however, will always match closely, even if the intensities are not quite in the same order. Once we learn d-values and intensities for an unknown mineral, it may seem an overwhelming task to find a match in a large reference file. It sometimes is, however, often we need only consider a few of the most intense diffraction peaks. Today, instead of doing the matching manually, most mineralogists use computer databases to match unknown patterns with PDF data. This has two main advantages: It is rapid, and many more reference patterns can be considered.

On the other hand, computer searches are “black box” procedures; they do not improve our knowledge or understanding of powder diffractometry. In addition, computers sometimes give absurd answers that a human would have rejected at a glance. Powder diffraction is not an exact process, and many things cause X-ray patterns to deviate from those in reference files. Proper sample preparation is crucial to obtaining accurate patterns. Complications arise when we do not grind a sample properly, or when powdered crystals are not in truly random orientations. A second problem, more so for natural materials than for synthetics, is that small compositional and crystal structural variations can affect X-ray patterns. Despite these complications, the PDF data contain adequate reference patterns for most mineralogical uses (Dexter Perkins 2002).

SmartLab SE X-ray Diffractometer from Rigaku

The Smart Lab SE is a highly versatile multipurpose X-ray diffractometer with built-in intelligent guidance. It offers continued refinement of the original ease of use features that enabled the original Smart Lab to receive the R&D 100 Award in 2006: automatic alignment, component recognition, cross beam optics and advanced photon counting hybrid pixel array detectors (HPAD).

Features.

- Smart Lab Studio II software based on a new architecturally integrated modular platform.
- Cross-beam optics module switches between Bragg-Brentano and parallel beam without the need to change optics.
- HyPix-400 2D detector enables seamless switch between 0D, 1D and 2D detection mode depending on application type.

- D/teX Ultra 250 1D detector accelerates powder diffraction by a factor of 250 in speed and provides adjustable energy resolution of approximately 20% or 4% depending on sample type.
- Integrated intelligent guidance software enables fully automated measurement including optics and sample alignment.

Figure 3.7: XRD machine, SmartLab X-ray Diffractometer from Rigaku at Goa University.

SAMPLE PREPARATION

Each of the five samples, weighing approximately 25g, was placed in a 250ml beaker with a small amount of water for soaking. Subsequently, the soaked samples were gently crushed using a glass rod. After crushing, 100ml of 6% hydrogen peroxide was added to each sample while stirring continuously, followed by heating at 40°C for one hour. Once sedimentation occurred, the samples were decanted to the minimum level and washed with distilled water. Next, 40ml of 2M sodium carbonate was added to each sample with continuous stirring to remove organic matter, silica content, and salinity. The resulting saline-free samples were decanted, and 250ml of 10% sodium hexametaphosphate was added, followed by a minimum 2-hour incubation period. Subsequently, the samples were sieved using a 230 mesh size sieve, and the sand fraction collected in pre-weighed 100ml beakers was dried in an oven. The sieved sample solution was collected in a 1-liter measuring cylinder, thoroughly stirred, and left undisturbed for settling. Samples were pipetted out at different time intervals and depths, specifically 20 seconds, 4 minutes and 28 seconds, and 3 hours and 41 minutes, from depths of 20cm, 10cm, and 5cm, respectively. These samples were collected in pre-weighed beakers and dried in an oven. After drying, the beakers were weighed for grain size calculations. Finally, the samples were further ground for XRD analysis. Here's the procedure for making slides for XRD analysis.

- Begin with a few tenths of a gram of clay mineral after drying.
- Grind the sample into a fine powder.
- Place the powdered sample onto a glass slide within a sample holder, ensuring a uniform smear and a flat surface. Secure it with double sticky tape in a sample container.

- Ensure the upper surface is flat and there’s a random distribution of lattice orientations, unless an oriented smear is needed.
- For clay samples requiring single orientation, refer to specialized techniques provided by USGS.
- For unit cell determinations, add a small amount of a standard with known peak positions that don’t interfere with the sample to correct peak positions.

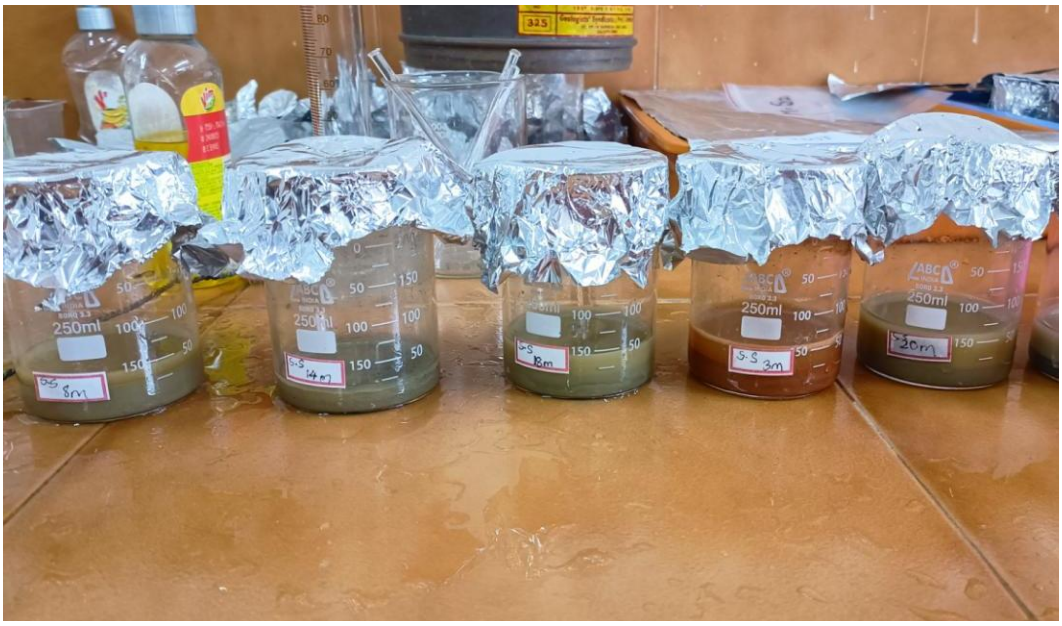


Figure 3.8 : Samples kept for soaking in water.

Figure 3.9 : Samples reacting with hydrogen peroxide

Figure 3.10 : Sieves used for sieving sand fraction

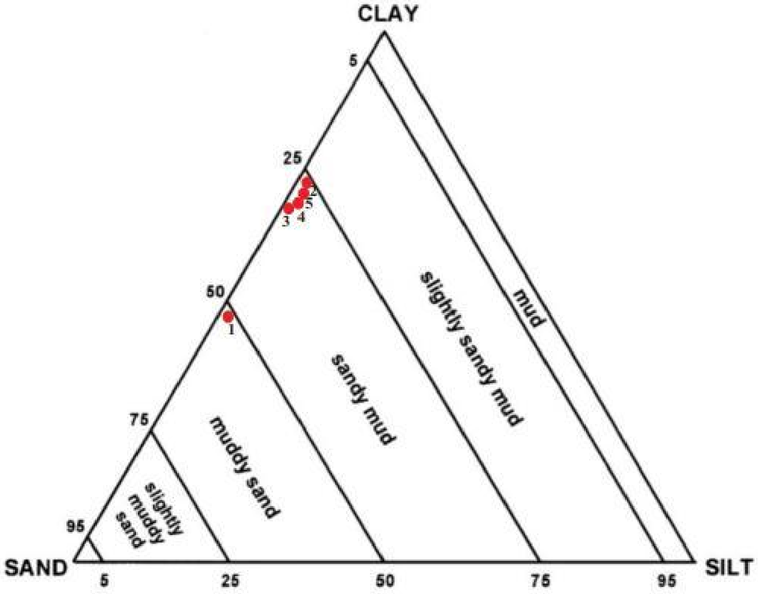
Figure 3.11: Sand fraction collected from the sieves kept for drying in oven

CHAPTER IV : OBSERVATIONS AND DISCUSSION

GRAIN SIZE DATA OF BOREHOLE

SAMPLE DEPTH	SAMPLE ID	GRAVEL %	SAND %	SILT %	CLAY %
3m	1	-	50.34	2.02	47.64
8m	2	-	26.052	0.916	73.032
14m	3	-	29.452	1.696	68.852
18m	4	-	26.88	2.196	70.924
20m	5	-	26.252	3.276	70.488

Table 4.1: Percentage of sediment fraction



(Refer table no.4.1)

Figure 4.1: Ternary diagram for sand-silt-clay mixture (after Flemming, 2000)

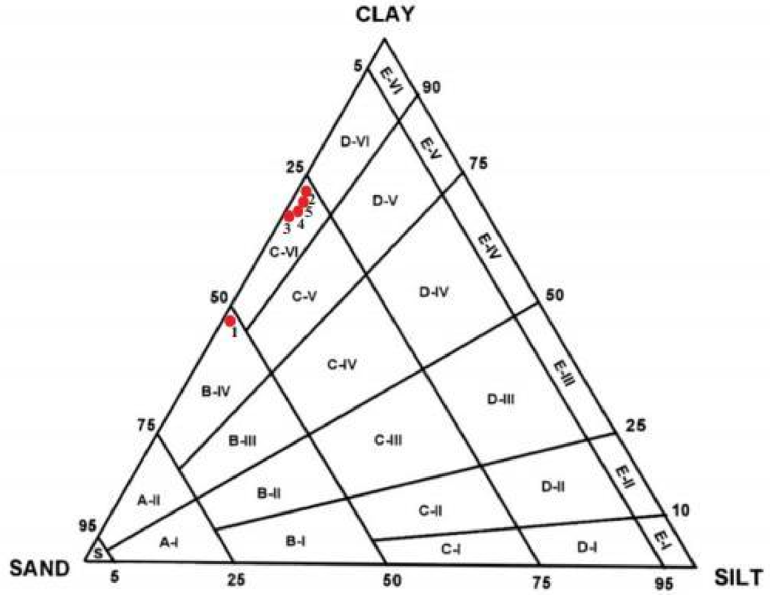
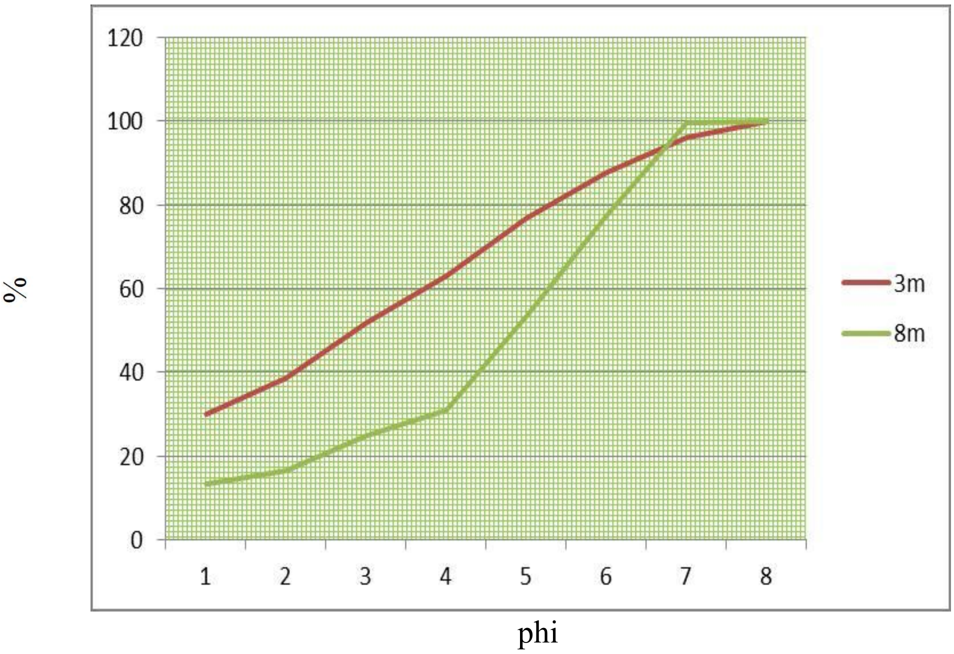


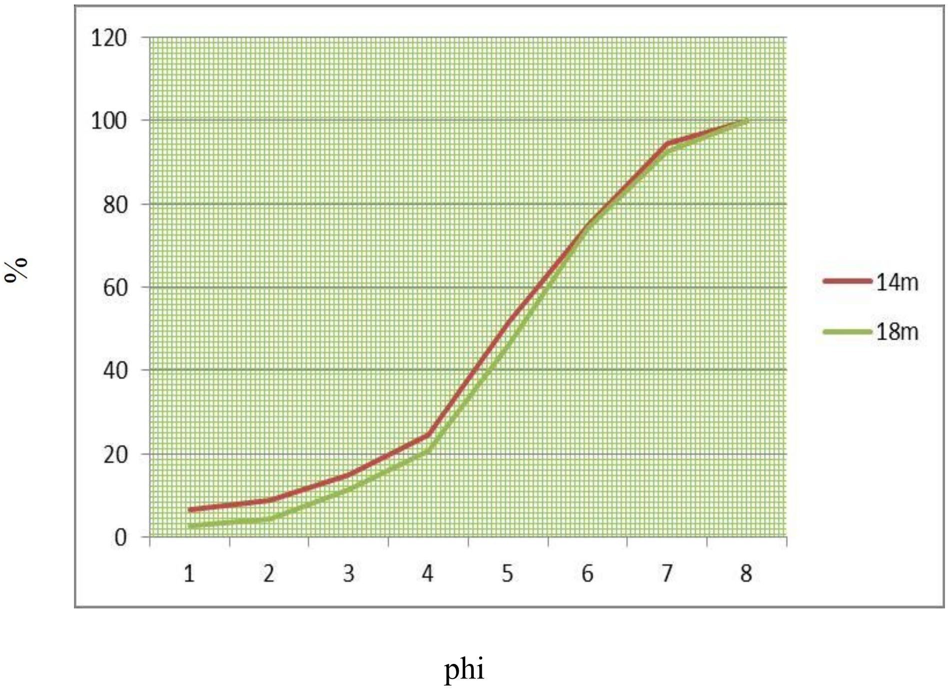
Figure 4.2: Ternary diagram representing detailed distinction for sand-silt-clay mixture (after Flemming, 2000)

SAMPLE ID	CODE	TEXTURAL CLASS NAME
3m	B-IV	VERY CLAYEY SAND
8m	C-VI	EXTREMELY CLAYEY SANDY MUD
14m	C-VI	EXTREMELY CLAYEY SANDY MUD
18m	C-VI	EXTREMELY CLAYEY SANDY MUD
20m	C-VI	EXTREMELY CLAYEY SANDY MUD

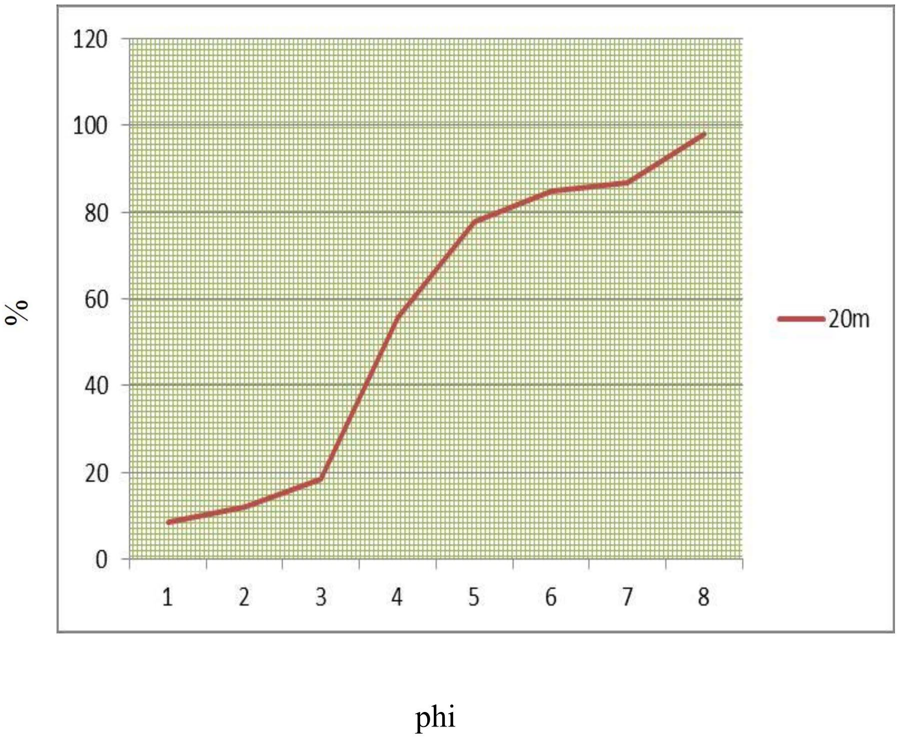
Table 4.2: Descriptive terminology of the three component sand-silt-clay


120
100
80
60
40
20
0
%
1
2
3
4
5
6
7
8
phi
3m
8m

4.3 (a)


120
100
80
60
40
20
0
%
1
2
3
4
5
6
7
8
phi
14m
18m

4.3 (b)



4.3 (c)

Figure 4.3 : (a) – (c) Cumulative graph of sediment

Sample no	Sample depth	Mean		Standard deviation		Skewness		Kurtosis	
1	3m	1.6166	Medium	1.3556	Poorly sorted	0.007	Positively Skewed	0.627	Very Platykurtic
2	8m	2.417	Fine	1.155	Poorly sorted	-0.414	Strongly negative Skewed	0.956	Mesokurtic
3	14m	2.666	Fine	0.9299	Moderately sorted	-0.263	Negatively Skewed	1.331	Leptokurtic
4	18m	2.783	Fine	0.791	Moderately sorted	-0.141	Negatively Skewed	1.252	Leptokurtic
5	20m	2.43	Fine	1.0208	Poorly sorted	-0.197	Negatively skewed	1.972	Very Leptokurtic

Table 4.3: Graphical grain size parameters for sediment samples

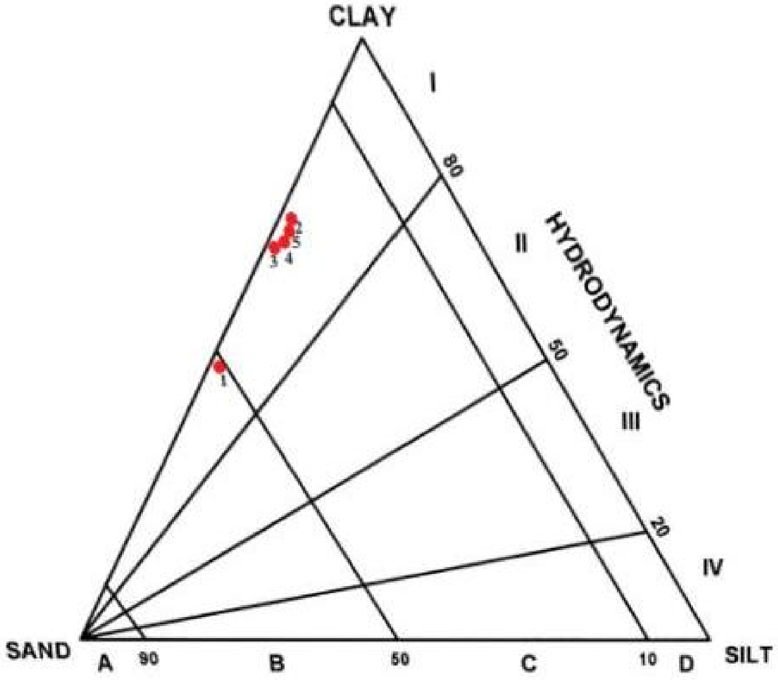
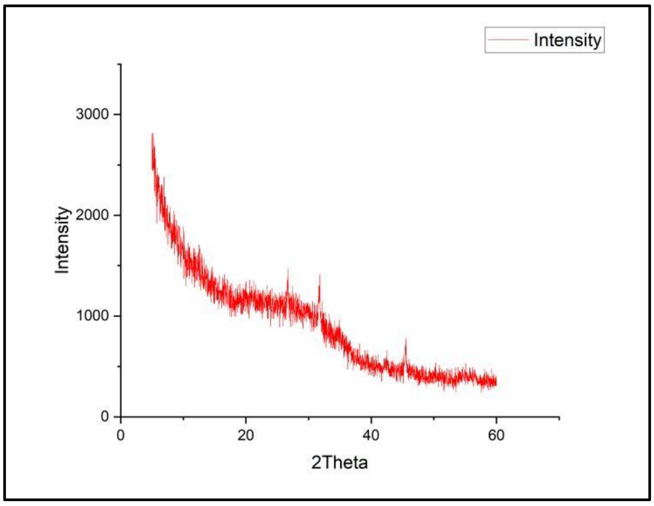


Figure 4.4 : Ternary diagram for classification of hydrodynamic conditions of depositonal environment (after Perjup, 1998)

Textural classification of each sample is given in table 4.2. Among the samples, slightly rough texture is seen in sample of 3m depth whereas samples from 8m to 20m has highest clay content, along with sand particles creating muddy texture. Values and description of grain size parameters namely mean, standard deviation, skewness and kurtosis is mentioned in Table 4.3. Based on ternary diagram (Perjup 1998), hydrodynamic conditions get stronger from I to IV. As seen in figure no.4.4, all samples lie in category I which had low hydrodynamic conditions. Categories A to D show the textural roughness with respect to increasing sand fraction.

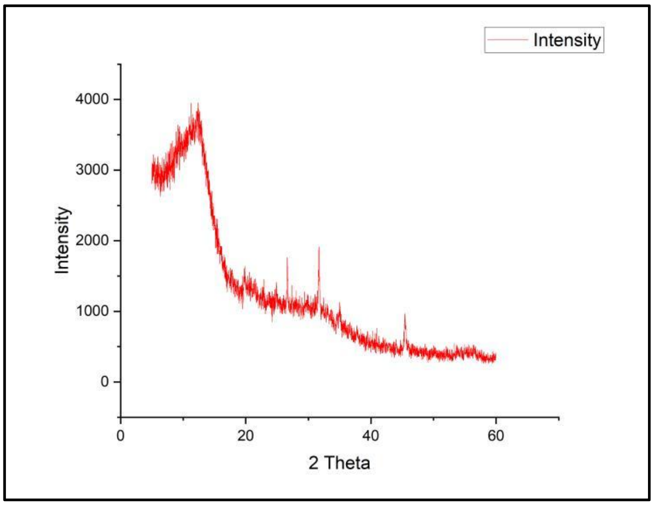
CLAY MINERALOGY

X ray diffraction patterns were obtained for 5 samples using Rigaku X ray diffractometer with copper source. Sample slides were prepared and placed in the diffractometer. Each sample analysis took 12 minutes for analysis.



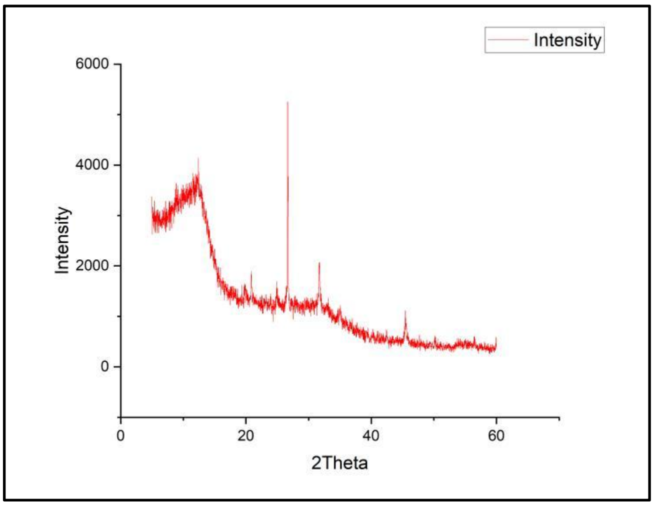
(a)

Sample_3m: P



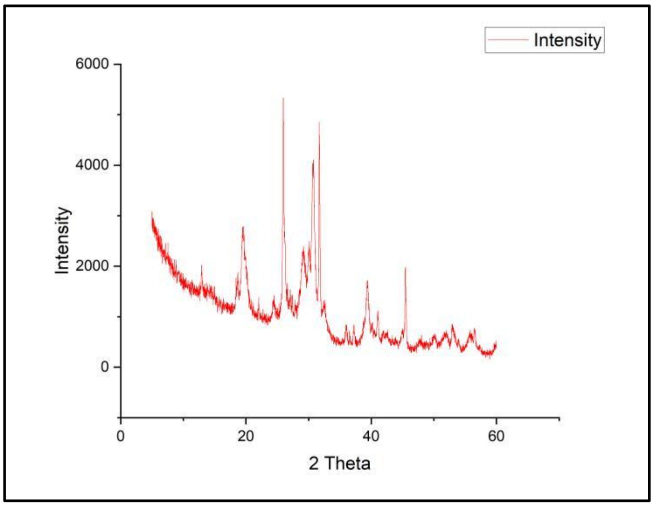
(b)

Sample_3m: Q



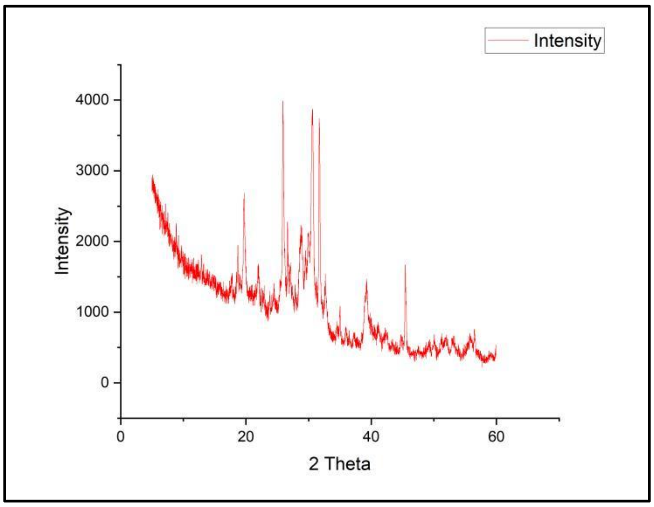
(c)

Sample_3m: R



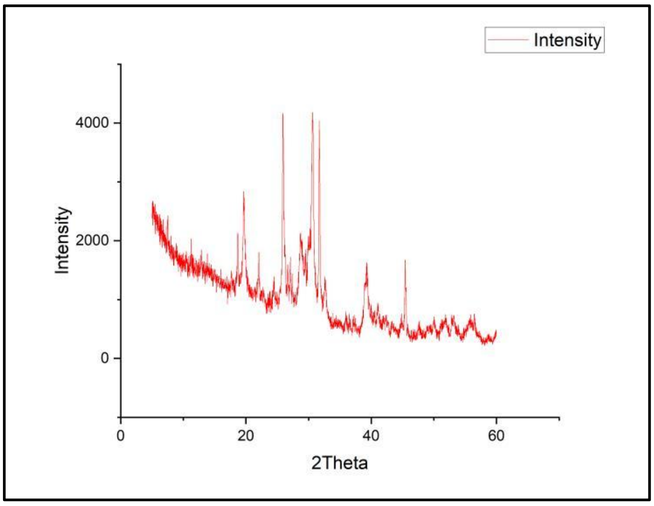
(d)

Sample_8m: P



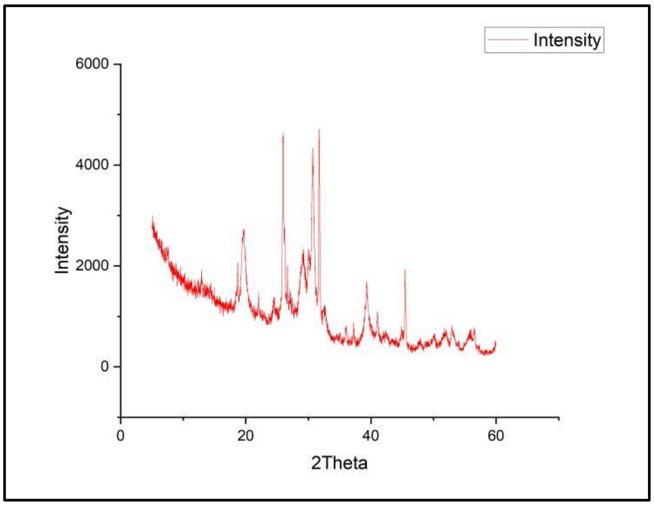
(e)

Sample_8m: Q



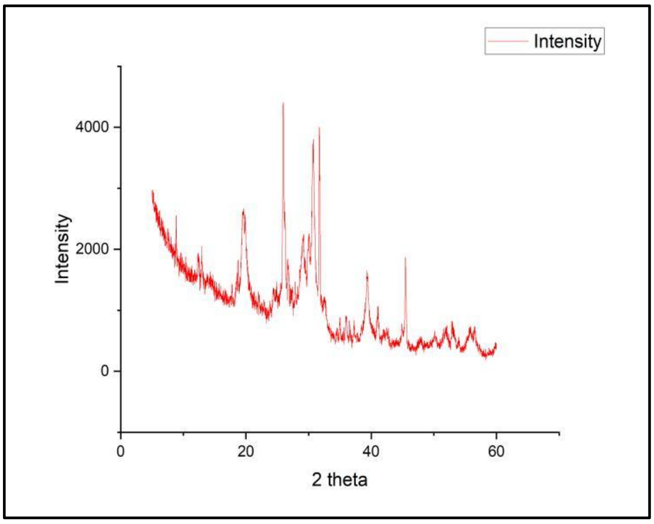
(f)

Sample_8m: R



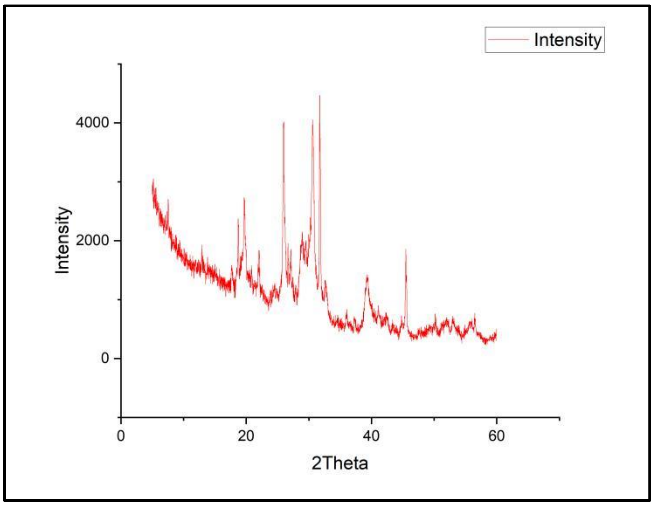
(g)

Sample_14m: P



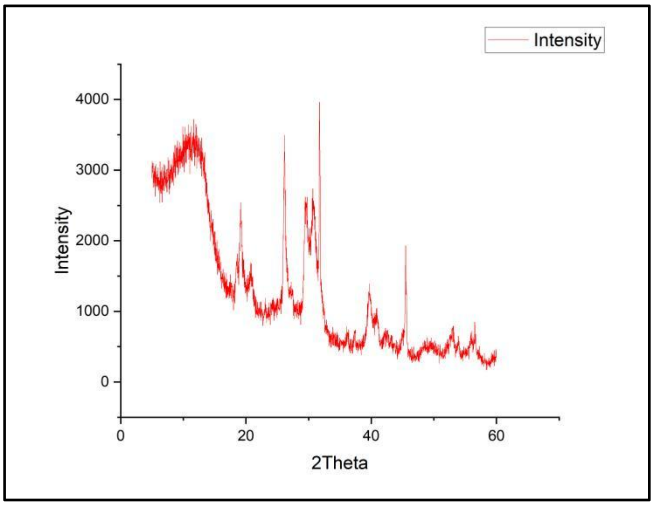
(h)

Sample_14m: Q



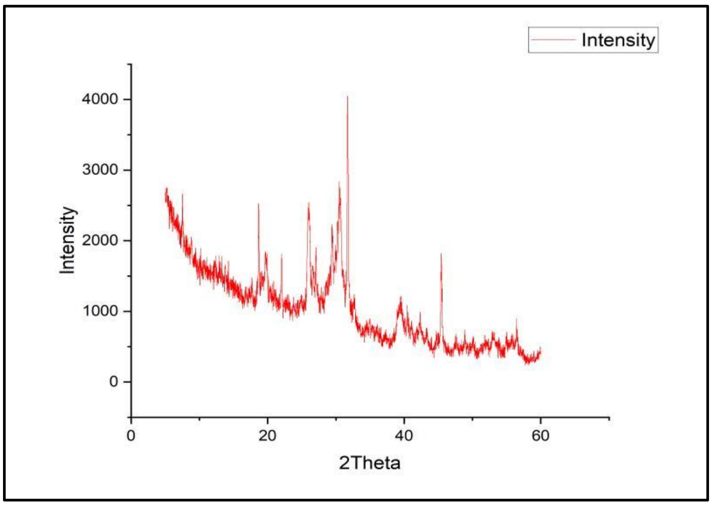
(i)

Sample_14m: R



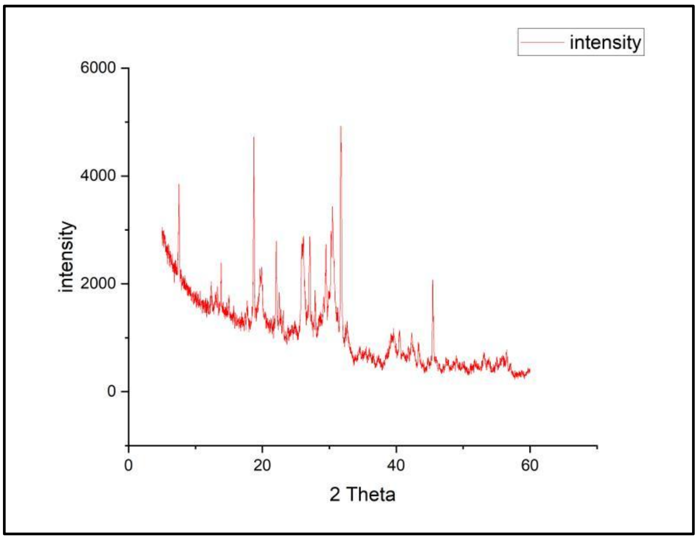
(j)

Sample_18m: P



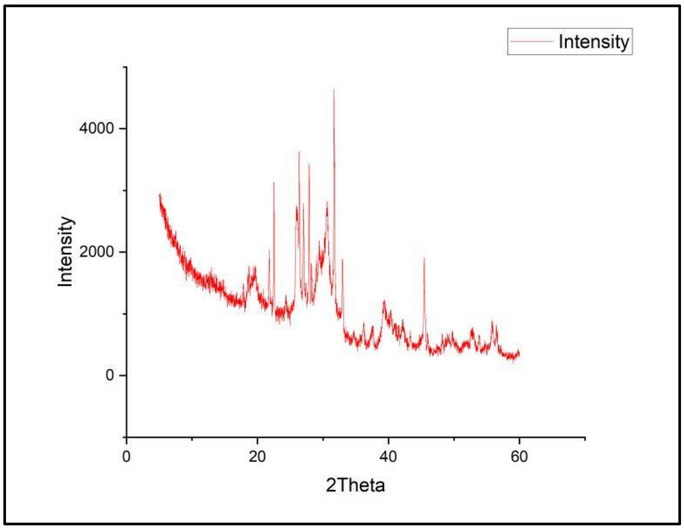
(k)

Sample_18m: Q



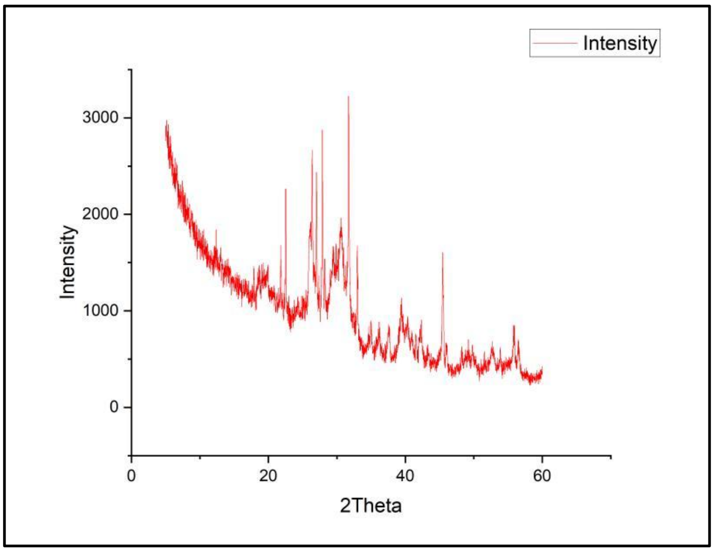
(l)

Sample_18m: R



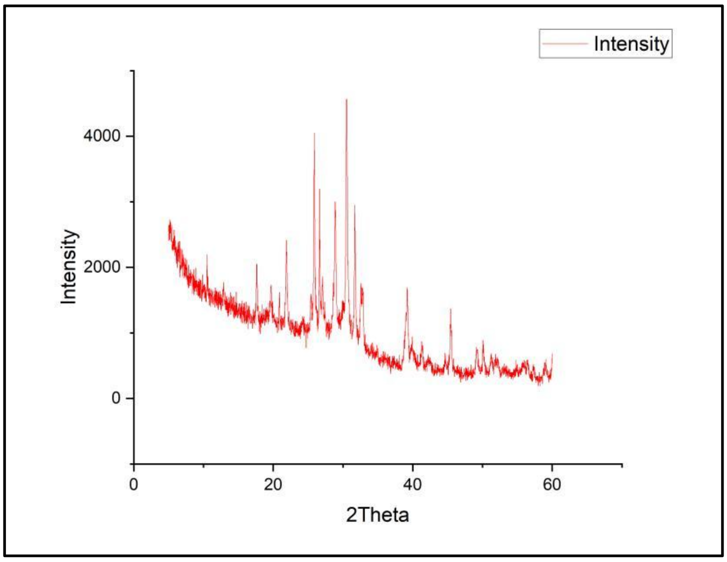
(m)

Sample_20m: P



(n)

Sample_20m: Q



(o)

Sample_20m: R

Figure 4.5: (a) to (o) XRD pattern of samples

SAMPLE ID	CLAY MINERALS
Sample_ 3m: P	Chlorite, Kaolinite, Montmorillonite, Quartz
Sample_ 3m: Q	Chlorite, Kaolinite, Montmorillonite, Quartz
Sample_ 3m: R	Chlorite, Kaolinite, Montmorillonite, Quartz
Sample_8m: P	Chlorite, Kaolinite, Montmorillonite, Quartz
Sample_8m: Q	Chlorite, Illite, Kaolinite, Montmorillonite, Quartz
Sample_8m: R	Chlorite, Illite, Kaolinite, Quartz
Sample_14m: P	Chlorite, Illite, Kaolinite, Quartz
Sample_14m: Q	Chlorite, Illite, Kaolinite, Quartz
Sample_14m: R	Chlorite, Kaolinite, Quartz
Sample_18m: P	Illite, Kaolinite, Quartz
Sample_18m: Q	Chlorite, Illite, Kaolinite, Quartz
Sample_18m: R	Chlorite, Illite, Kaolinite, Quartz
Sample_20m: P	Chlorite, Illite, Kaolinite, Montmorillonite, Quartz
Sample_20m: Q	Chlorite, Illite, Kaolinite, Montmorillonite, Quartz
Sample_20m: R	Chlorite, Illite, Kaolinite, Quartz

Table 4.4 : Clay minerals depth-wise

PETROGRAPHY

The hand specimen appears as a dark to medium green-colored rock with a glossy to dull luster. The surface of the rock reveals a foliated texture, with alternating layers of felsic and mafic minerals. The mineral chlorite, which typically occurs as small, platy grain or aggregates. These chlorite grains impart a slight sheen to the surface of the rock. In addition to chlorite, the specimen contain other minerals such as quartz, feldspar and opaque mineral.

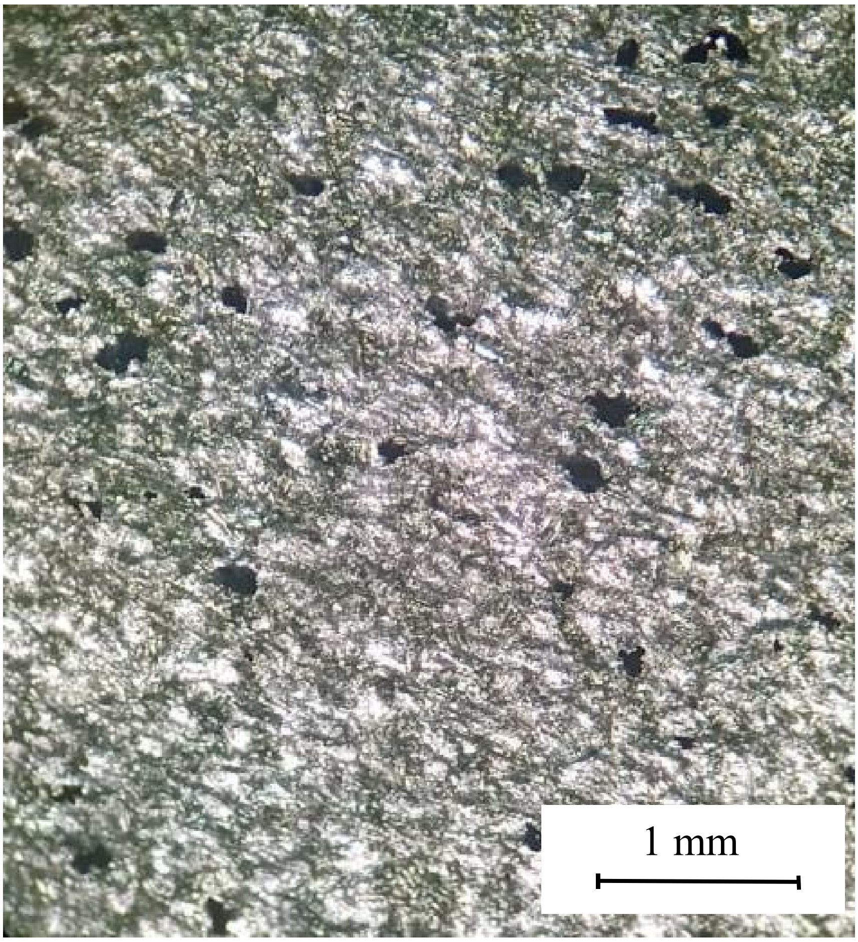


Figure 4.6: Plane polarized photomicrograph chlorite exhibiting lepidoblastic texture.

Under the microscope the rock specimen shows low relief, no clevage, colorless, interference color is first order grey, the mineral is quartz, K-feldspar has low relief no cleavag, colorless, interference colors first order grey and pitted appearance and opaque mineral. Chlorite is colourless to green/pale green, perfect cleavaget, low to moderate relief , interference color is very weak Grey/berlin blue, extinction angle is 0-10°. Alignment of the Chlorite gives rise to lepidoblastic texture to the rock. There is characteristic quartz vein which is crosscutting the foliation but the biotite grains are aligned parallel to the foliation. Quartz vein must have crystallize during the escape of fluids from the parent rock. A porphyroclast of quartz has crystallized parallel to the foliation and the foliation is wraping around the grain. The rock shows abundance of anhedral magnetite grains which are aligned along the foliation of rock. As the abundant mineral is chlorite, which is showing schistosity and also characterized by quartz which is present as porphyroclast and injected vein of quartz, the rock can be called as qaurtz-chlorite schist. The degree of schistosity indicates that the rock has undergone low grade regional metamorphism and the expected parent rock is shale or mudstone.

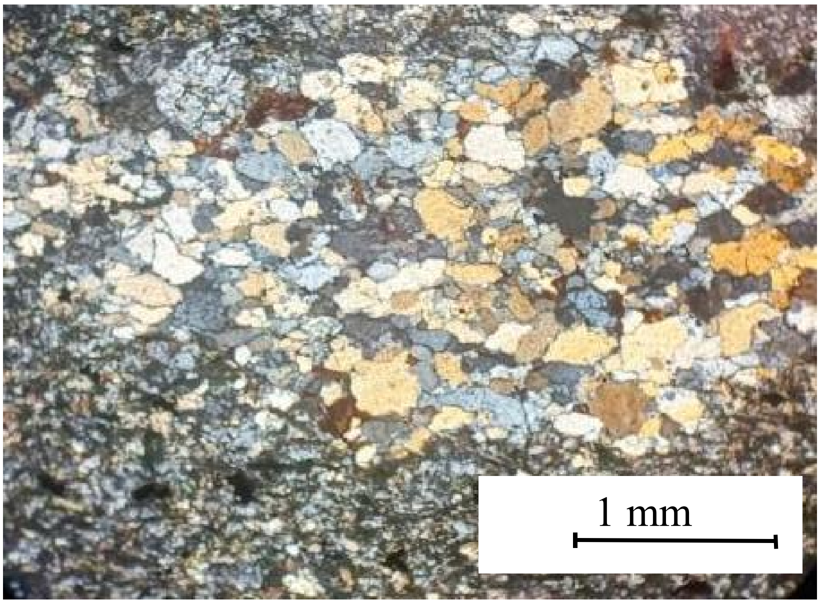
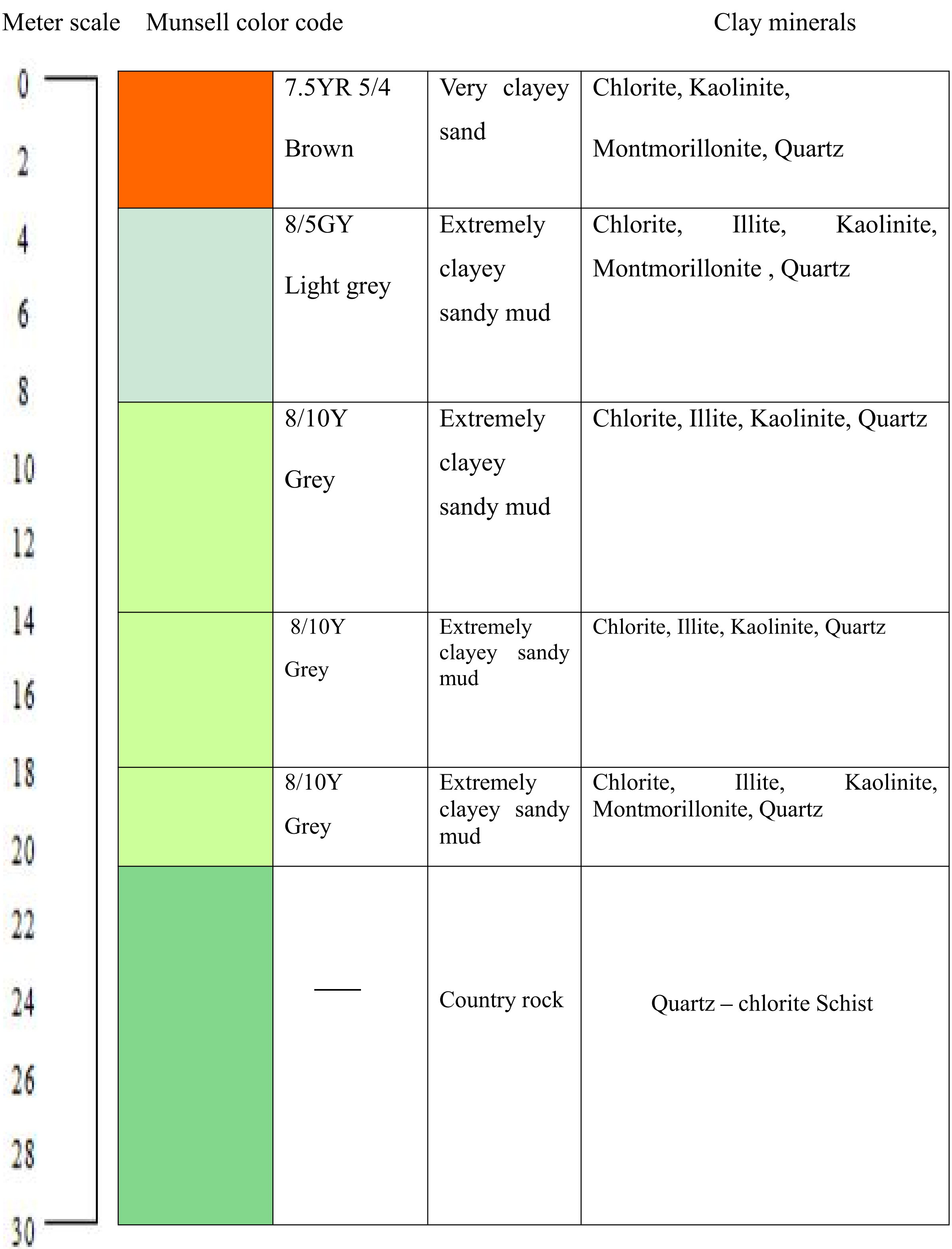


Figure 4.7: Cross polarized photomicrograph showing porphyroclast of quartz, wraped by Chlorite.

Figure 4.8: Plane polarized photomicrograph showing quartz vein crosscutting the foliation.



Meter scale	Munsell color code			Clay minerals
	🟧	7.5YR 5/4 Brown	Very clayey sand	Chlorite, Kaolinite, Montmorillonite, Quartz
	⬜	8/5GY Light grey	Extremely clayey sandy mud	Chlorite, Illite, Kaolinite, Montmorillonite , Quartz
	🟨	8/10Y Grey	Extremely clayey sandy mud	Chlorite, Illite, Kaolinite, Quartz
	🟨	8/10Y Grey	Extremely clayey sandy mud	Chlorite, Illite, Kaolinite, Quartz
	🟨	8/10Y Grey	Extremely clayey sandy mud	Chlorite, Illite, Kaolinite, Montmorillonite, Quartz
	🟩	—	Country rock	Quartz – chlorite Schist

Figure 4.9: Descriptive log of study area

INTERPRETATION

From the sediment pile five available samples from various depth that were taken for analysis at 3m, 8m, 14m, 18m, 20m and at a depth of 30m is the country rock quartz-chlorite schist belongs to Bicholim Formation. Grain size parameter revealed rough texture at 3m, whereas at 8m, 14m, 18m and 20m it shows smooth texture. Rough texture typically contain coarse grains that can be felt as irregularities or roughness when touched. These sediments may consist of larger particles such as sand, or pebbles, which create a rough surface due to their angular shapes and uneven distribution and smooth texture usually consist of finer particles that feel soft or silky to the touch. These sediments may contain silt or clay particles, which have smaller sizes and smoother surfaces. Graphic mean size is the average size of the sediments and is influenced by the source of supply, transporting medium and the energy condition of the depositing environment. Mean size indicates the central tendency or the average size of the sediment and in terms of energy, it indicates the average kinetic energy velocity of depositing agent (Sahu 1964). The sorting variation observed attributes to the difference in water turbulence and variability in the velocity of depositing current. Except for the samples 14m and 18m, the other are poorly sorted indicating that there was velocity variation that caused turbulence, which resulted in samples that were poorly sorted to moderately sorted in nature.

Skewness is used to determine the symmetry of the central part of the distribution. It reflects the symmetry or asymmetry of the frequency distribution of the sediments. Skewness results from mixing of two different normal populations in different proportions and implies the dominance of coarse over fine and vice versa. The skewness values indicate the kinetic energy of the depositing medium. 3m is positively skewed value and 8m, 14m, 18m and 20m are negatively skewed values. (Layade, et.al 2019).

The percentage of clay in the mud fraction serves as a proxy for the hydrodynamic state of the depositional environment. From section I to section IV, the hydrodynamic condition becomes more turbulent. The category is divided into four groups, A to D (after Perjup, 1988) based on how much sand contains. Triangle has been divided into 16 groups. 3m falls into B-I between 50% and 90%. With group C-I samples, between 50% and 10% of them contain sand that was laid down in low energy environmental conditions, which denotes a calm and quiet situation (rarely observed in an estuarine environment). The component plotted in this study for 8m, 14m, 18m, 20m. The plots in the ternary diagram for classification of hydrodynamic conditions of the depositional environment (after Perjup, 1998) indicates that the sediment underwent only one sequence of depositional environments.

The presence of the following clay minerals suggest kaolinite may be formed by strong leaching crystalline metamorphic, phyllitic rock and silicate alteration of laterite tropical climates. Illite may be formed by leaching of crystalline metamorphic and phyllitic rock. Montmorillonite may be brought to estuary from hinterland. It is important mineral constituents of soils, especially in tropical areas. It is created by the surface weathering of rocks with the presence of solutions containing Na, Ca, Mg and ferrous iron, series under the conditions of low pH which is characteristic of fresh and seawater. Chlorite is typically formed through the alteration, particularly those rich in magnesium, iron, and aluminum, such as biotite, hornblende, and certain types of micas. This alteration process, known as chloritization, occurs under low-grade metamorphic conditions or during hydrothermal alteration. Essentially, chlorite forms when these parent minerals react with water and other substances in the presence of heat and pressure, leading to the replacement of their original chemical components with those of chlorite. (Kumari, et al. 2021)

At 30m depth a country rock which contains chlorite as a essential mineral with other accessory minerals includes quartz, K- feldspar and opaque mineral. Based on the mineral composition of the country rock which were observed both megascopically and microscopically the rock is classified as quartz -chlorite schist. The study area indicates that the lithology of that area belongs to the Bicholim Formation.

The drill pile from study area described above permits to make certain deductions on the paleo-environmental conditions during the deposition of the sediments in the Zuari estuary. Given that the sediments contains organic matter, it can be inferred that they are a product of the maritime intrusions throughout this time period depositional phases. Dessai (2018) considers the clayey sediments over the sub-sea level laterites equivalent to the clays from Dapoli which have been dated to 3.5-2.6 ka BP (Kumaran et al, 2012). Further work on these sediments is required to place firm controls on the type of sedimentation and the paleo-enviroments of their deposition.

RESULT

In the 3P60 pile the country rock, quartz - chlorite schist was encountered at 30m depth which belongs to Bicholim Formation. Five available samples were collected from the borehole, each representing a different depth of 3m, 8m, 14m, 18m and 20m. 3m have a slightly rougher texture than the samples 8m, 14m, 18m and 20m, which have a relatively smoother texture, according to the grain size measure. The transport method, supply source, and energy level of the environment where the sediment is being deposited all have an impact on the mean size, or average size, of the sediments. The skewness values indicate the kinetic energy of the depositing medium. 3m is positively skewed (coarse) and 8m, 14m, 18m and 20m are negatively skewed (fine) (Layade, et.al 2019). To know the hydrodynamic condition the data are plotted of each sample, 3m falls into B-I symbolises very quite hydrodynamic conditions. On the other hand 8m, 14m, 18m, 20m which falls into C- I which denotes calm and quite situation. The plots in the ternary diagram for classification of hydrodynamic conditions of the depositional environment (after Perjup, 1998) indicates that the sediment underwent only one sequence of depositional environments.

From XRD analysis of clay quartz, kaolinite, illite, chlorite and montmorillonite minerals are found. The descriptive log is the result of this data. Paleo - environment conditions of the study area during the depositon of the sediments in the Zuari estuary. Given that the sediments contains organic matter, it can be inferred that they are a product of the maritime intrusions throughout this time period depositional phases. According to Dessai (2018), the clayey sediments above the sub-sea level laterites are identical to the Dapoli clays, which have been dated to between 3.5 and 2.6 ka BP (Kumaran et al, 2012).

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