LITHOGENIC FLUXES TO THE ARABIAN SEA AND BAY OF BENGAL

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BY

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CERTIFICATE

Mr. V. Ramaswamy has been working for his PhD. degree under my guidance since 1988. The PhD. thesis entitled '**Lithogenic fluxes to the Arabian Sea and Bay of Bengal**' submitted by him contains the results of his original investigations of the subject. This is to certify that the thesis has not been the basis for the award of any other research degree or diploma of this or any other university.

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V. Ramaswamy.

(V.Ramaswamy)

To my Father

LITHOGENIC FLUXES TO THE ARABIAN SEA AND BAY OF BENGAL

CONTENTS

STATEMENT

As required under the University ordinance 19.8 I state that the present thesis entitled "**Lithogenic fluxes to the Arabian Sea and Bay of Bengal**" is my original contribution and the same has not been submitted for any degree of this or any other University on any previous occasion. To the best of my knowledge the present study is the first comprehensive study of its kind from the area mentioned.

The literature concerning the thesis has been cited. Due acknowledgement have been made wherever facilities have been availed of.

V. Ramaswamy

(V.RAMASWAMY)

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General Introduction

There are two main sources of particles in the oceans. Biogenic particles which are formed due to planktonic metabolism in the upper layers of the oceans and lithogenic particles (mostly clay minerals) transported from the continents by rivers and winds.

Vertical settling of particles is one of the most important pathways for transfer of material and energy from the surface of the ocean to the sea floor. Phytoplankton synthesizes carbon, silica and other essential elements from dissolved to solid form in the photic zone with the help of the Sun's energy. Most of the biogenic material is rapidly remineralised, but the more refractory elements and lithogenic particles settle through the bathypelagic layers. As particles fall through the water column their composition and concentrations are altered by physico-chemical and biological processes of aggregation, disaggregation and dissolution.

Most conclusions on oceanic sedimentary processes have come from the study of physical and chemical properties of sediments collected from the ocean floor and speculation about the various processes controlling their deposition. Core tops are usually bioturbated and have a resolution between a few hundred to thousands of years. Studies on settling particles in the water column are few but can be important as seasonal and interannual differences in sedimentary processes can be easily distinguished.

In order to understand the settling mechanism of particles in the oceans it is necessary to differentiate between suspended particles and settling particles. A suspended particle is often made of fine discrete particle like clay flakes, cocolith or diatom tests. They have insignificant settling velocities and can be advected over long distances easily. Settling particles are coarser and may be a discrete particle like foraminifera or pteropod shell, or more often, an aggregate of fine particles known as 'marine snow' (*Alldredge and Silver, 1988*). They have settling velocities ranging between a few tens of meters up to a kilometer per day. Since their residence time in the water column is very less they are not significantly altered by chemical or biochemical reactions before arriving on the sea floor. However, there are significant exchange of material between the suspended and settling pool due to aggregation and disaggregation of particles in the water column (*Lal, 1977*).

Suspended sediment in the ocean has been traditionally sampled by filtering sea water collected by conventional bottle samplers. A major drawback of this method is that the samples are biased towards finer sized particles (*Gardner, 1977*). Coarse particles like macro-aggregates or zooplankton fecal pellets which contribute the major portion of particles settling in the water column (*McCave, 1975*) are missed out (*Gardner, 1977*). Coarser particles, though rare in the oceanic water column, are more important as there is an exponential increase in mass as well as sinking velocity with increase in size (*McCave, 1975*). These particles can be effectively sampled by filtering large volumes of water in situ (*Bishop, 1976*) or by deploying particle collectors (commonly known as sediment traps) in the water column (*Gardner, 1977; Honjo and Doherty, 1988*).

Since the early 1900's containers of various shapes have been deployed to collect settling particles. Hydrodynamic characteristics of these samplers have been studied by *Gardner (1977; 1980a,b*) and based on his studies sediment traps having the necessary hydrodynamic characteristics could be designed. Since then a number of sediment trap experiments have been conducted in the World oceans (*Wefer, 1989*). The direct measurement of particle fluxes in the oceans has followed developments in marine instrumentation and material sciences. Deployment and recovery of deep sea moorings, which position sediment traps in the water column, with a high rate of success has been accomplished only during the last two decades. The advent of time-series sediment traps (*Honjo and Doherty, 1988; Honjo, 1990*) has facilitated the study of seasonality in particle fluxes.

During the last decade a number of year round sediment trap studies have been carried out in the World oceans. In general it is seen that particle sedimentation in the open ocean is related to biological processes in the surface waters (*Deuser and Ross, 1980; Honjo 1982; Wefer 1989*). Biological processes not only produce a major portion of the particles in the open ocean but also provide 'carrier particles' that scavenges fine particles from the water column and transports it to the sea floor.

A significant fraction of settling particles is made of lithogenic particles derived mostly from land (*Honjo, 1990*). Most of the lithogenic particles in the deep sea are less than 5 micrometer in diameter and have an effective sinking velocity of a few centimeters per hour. This suggests that clays should take a few decades to hundreds of years to settle through a 4 km water column. By this time they should be thoroughly dispersed throughout the world oceans. In fact, particles finer than 40 micrometer cannot settle in the oceans as its sinking velocity is less than the speed of convection of water in the oceans (*Degens and Ittekkot, 1987; Degens, 1989*). Latitudinal distribution of clay minerals in the oceans points to the fact that clays actually settle much faster (*Griffin et al., 1968; Lisitzin, 1972*). The importance of biological and physico-chemical processes in aggregation and accelerated sinking of fine lithogenic particles has been demonstrated by *Honjo* (*1982*) and *Deuser et al.* (*1983*). Accelerated sinking of clays is mainly due to its incorporation into zooplankton fecal pellets and marine snow which have sinking velocities in-excess of a hundred meters per day.

The northern Indian Ocean has received substantial amounts of sediment from the Indus, Ganges-Brahmaputra and other rivers draining the Indian Peninsula since Eocene times (*Kolla and Kidd, 1982*). This has resulted in the formation of two of the Worlds largest submarine fans, namely the Bengal fan and the Indus fan. The bulk of the suspended sediment load of rivers draining into the Arabian Sea and Bay of Bengal is deposited in deltas, estuaries and shallow continental shelves. They are then reworked and transported to the deep sea by various physical processes of resuspension and dispersion like deep sea currents. Terrigenous matter can also be deposited directly on the deep sea floor by turbidity currents.

The northern Indian Ocean is also affected by the SW and NE monsoons. During the monsoons, fresh water and suspended sediment discharge of rivers as well as eolian fluxes are higher. This input of material from land together with monsoons winds strongly influences the circulation, stratification, productivity and sedimentation patterns of the northern Indian Ocean. Sediment trap studies can give us important information about the effect of the monsoons on settling behavior of particles in the oceans.

The main objectives of the present investigations are

1. To understand factors controlling particle fluxes in the northern Indian Ocean. In particular, factors controlling dispersion and settling of lithogenic matter in the oceans.
2. To determine the mineralogy and grain size of the lithogenic fraction in order to establish their provenance.
3. To compare the present day rain rate of particles in the water column with Holocene mass accumulation rates.

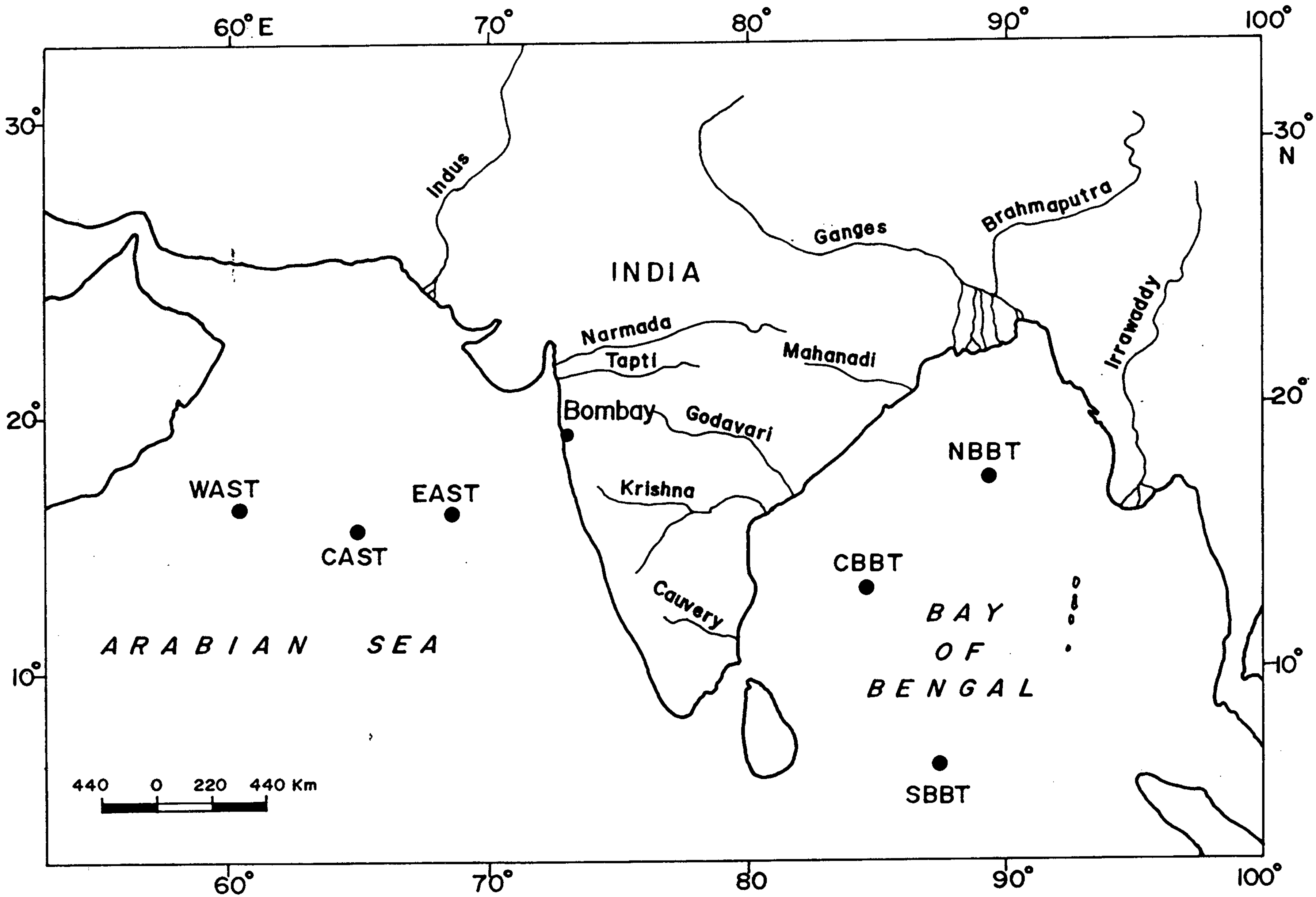


Fig. 1.1 Figure showing sediment trap locations in the northern Indian Ocean.

For the present study settling particles have been collected using twelve time- series sediment traps deployed at six locations in the northern Indian Ocean (Fig. 1.1). The samples have been collected under a joint collaborative programme between the **National Institute of Oceanography, Goa, India** and the **Institute of Biogeochemistry and Marine Chemistry, Hamburg University, Hamburg, Germany**.

The fluxes of first order components which make up the bulk of the particles like carbonates, opal, organic carbon and lithogenic matter have been determined by chemical analysis. The lithogenic matter was separated, and clay mineral and grain size analysis has been carried out on this fraction.

Factors responsible for the seasonality in fluxes of total particulate matter and lithogenic matter have been studied and related to monsoon processes. A comparison has been made between particle fluxes to the Arabian Sea and the Bay of Bengal.

Seasonality in lithogenic fluxes, increase in lithogenic flux with depth as well as its relation to other components in the Arabian Sea and Bay of Bengal have been presented and discussed in detail. Biological control on lithogenic fluxes has been demonstrated. Based on the relative abundance of various clay minerals in the sediment trap samples, material derived from the Himalayan region, Peninsular Indian rivers and eolian dust from the Arabian deserts have been distinguished. Grain size analysis of the lithogenic fraction in sediment trap and core tops from the Arabian Sea and Bay of Bengal was carried out using a laser based particle size analyzer. Multi modes in grain size with prominent peaks at 1.5 and 4.5 micrometer have been related to mineralogy of the clay minerals.

Rain rates of various first order components calculated from sediment trap data have been compared with accumulation rates in sediments. Also the remineralization rates of these components at the sediment water interface has been calculated. An attempt has been made to understand factors responsible for preservation of labile biogenic components.

Chapter 2

The Study Area

To understand formation of hemi-pelagic sediments in the oceans it is necessary to understand factors controlling production of biogenic matter in the photic zone, water column processes as well as the geology of the surrounding landmass from which freshwater and saspended matter is derived. This chapter contains a description of the physiography of the region, geology of surrounding continents and the amount and type of material discharged into the area from adjacent landmass. Also included are a brief description of the general meteorological and oceanographic parameters which control production and settling of particles in the oceans.

2.1 Physiography of the Arabian Sea and Bay of Bengal

The Arabian Sea and Bay of Bengal are two arms of the northern Indian Ocean separated by the Indian peninsula (Fig. 1.1). They are situated in more or less the same latitudinal belt in the tropics and are blocked on the north by the Asian landmass. Both the seas also come under the influence of the monsoons. However, the freshwater and suspended sediment discharge into the Bay of Bengal is much higher.

The Arabian Sea is located in the northwestern part of the Indian Ocean. It occupies an area of 7.5 million km^2 (*LaFond, 1966*) and extends from the equator up to 25^{o}N. To the north and west, it is surrounded by arid land masses of the Iran-Makaran-Thar area, Arabian Peninsula and Horn of Africa. The Bay of Bengal is located in the northeastern part of the Indian Ocean and occupies an area of 2.2 million sq.km (*LaFond, 1966; Curray and Moore, 1974*). It extends up to 22^o N and is bordered on the north by the deltaic regions of the Ganges and Brahmaputra rivers. On the east it is bounded by the Burmese peninsula and the Andaman-Nicobar ridges and to the west by the Indian Peninsula. The southern boundary of the Bay of Bengal is fixed from Dondra head at the southern tip of Sri Lanka to the northern tip of Sumatra and is open to the Indian Ocean towards the south (*Curray and Moore, 1974*).

In the Arabian Sea, the widest shelf is off Cambay where the width exceeds 350 km. Off Karachi the width is about 185 km and is incised by the Indus Canyon. The continental shelf of the western margin of India narrows gradually towards the south and off Cochin it is about 45 km wide. Off the southern tip of India it is about 120 km wide. Off Baluchistan and Iran the shelf width narrows down with a mean width of 37 km. The continental shelf off the Arabian Peninsula and Somalia is very narrow and is less than 50 km wide. Along the east coast of India the shelf is relatively narrow with an average width of 40 km and increases from south to north (*Curray and Moore, 1974*). It is narrowest off Kakinada where the shelf width is only 13 km. Off the Ganges-Brahmaputra delta it widens to over 160 km. Similarly it widens to over 120 km off the Irrawady-Salween and Mergui deltas.

The deep sea floor of the Arabian Sea consists of two distinct basins separated by the NW-SE trending Murray Ridge (*Coumes and Kolla, 1984*). The Arabian Basin is dominated by the Indus Submarine Fan. This fan covers an area of 1.1 x 10^6 km^2. and is about 1500 km and has a maximum width of 960 km (*Kolla and Coumes, 1985*). The upper Indus Fan is characterized by a relief of several hundred meters, as a result of aggradation of channel-levee complexes. The lower fan has a smooth relief with smaller channel and levees. Sedimentation is dominantly channelized-turbidity currents with overbank deposition on the upper fan and both channelized and unchannelised turbidity currents during lowstands of sea level. The Indus Submarine Canyon also called "The Swatch" incises the shelf and slope off the Indus river and serves as the main conduit for channelizing the suspended load of the Indus River (*Kolla and Coumes, 1987*).

South of the Indus Fan lies the Arabian Abyssal Plain. It has a maximum depth of 4700 m, and is bounded on the south by the Carlsberg Ridge and on the east by the Chagos-Laccadive ridge. These ridges prevent the transport of sediments from one basin to the other. The Murray-Owen Ridge consists of a linear series of seamounts, scarps and small basins. East of this lies two relatively smaller abyssal plains of the Oman Basin and Owen Basin.

The Bengal basin is a broad U shaped basin open to the south and dominated by the Bengal and Nicobar Fans. The Bengal Fan extend from the river mouths of the Ganges-Brahmaputra for more than 3000 km to south of the equator (*Curray and Moore, 1974; Emmel and Curray, 1984*). This Fan is about 1000 km wide with an estimated thickness of over 12 km in the northern part. The central part of the Fan slopes uniformly from around 2000 to 4000 m at the rate of 2 m per mile which points to the stability of the underlying crust. Most of the tectonic features in the deeper part of the Bay of Bengal have been buried by sediments of the Bengal Fan. The Ninety East Ridge is the most prominent bathymetric high (*Emmel and Curray, 1984*). It prevents sediments from the Irrawady from being deposited on the Bengal Fan. The portion of the Bengal Fan between the Ninety East Ridge and Sumatra-Java is called the Nicobar Fan (*Curray and Moore, 1974*).

There are numerous submarine trenches and canyons in the Bay of Bengal. The north-south trending Indonesian trench near the Nicobar-Sumatra mainland has a maximum depth of 4500 m (*Emmel and Curray, 1984*). The Swatch-of-no-Ground, near the Ganges-Brahmaputra river mouths has a width of 8 miles and gouges 600 to 800 m into the surrounding plains. The sides of the canyons are extremely steep but the bottom is nearly plain. Sediments are funnelled to the Bengal Fan via a delta-front trough the "Swatch of No Ground". This trough is presently connected to only one active fan channel, but has been effectively cut off from its principal sediment supply and the Ganges-Brahmaputra Delta, since the rise in sea level probably about 7000 to 10000 BP (*Curray and Moore, 1974*). There are numerous other channels extending for various distances. Most of these are thought to have been abandoned by channel switching. Several other canyons and channels also cut across the east Indian margin. In 1963 during the IIOE the Andhra, Mahadevan and Krishna canyons were discovered off the coast of Andhra Pradesh. The Cuddalore, Puducherry and Palar Canyons are located off Pondicherry .

The Bengal fan can be divided into a upper, middle and lower fan. Although there are numerous submarine channels in this fan, one major channel runs right from the Swatch of no Ground (20 ^{o}N) to the equator and it is possible that the sediment discharge of the Ganges-Brahmaputra is channeled through this major channel and deposited in the distal parts of the Bengal Fan (*Emmel and Curray, 1984*). The lower fan has a smooth topography and is composed of silty, muddy, and bioclastic sediments, with thin interbedded pelagic and hemipelagic layers (*Stow et al., 1990*).

2.2 General oceanography of the northern Indian Ocean

The northern boundary of the Indian Ocean is blocked by the Asian land mass. There are two major consequences of this. First, the northern Indian Ocean is separated from the convectional circulation of the northern hemisphere leading to weak circulation and poor renewal of waters. Secondly, the large land mass acting as a barrier to the north has led to the development of an atmospheric circulation system known as the monsoon which influences the Indian Ocean up to 10 °S (*Wyrtki, 1973*).

The Indian Ocean summer monsoon represents one of Earth's most dynamic interactions between land, ocean and atmosphere (*Duing, 1970*). The summer monsoon is driven primarily by differential heating between land and ocean which in summer leads to a seasonal formation of low atmospheric pressure over the Asian Plateau and high atmospheric pressure over the relatively cold Indian Ocean leading to initiation of the summer monsoon circulation (*Webster, 1987*). The monsoon circulation is intensified by transfer of heat from the Indian Ocean over the area of precipitation. Latent heat collected over the equatorial and subtropical Indian Ocean is transported over the equator and released into the troposphere due to precipitation. The combined effect of the sensible and latent heating results in strong southwest monsoon winds (*Hasenrath and Lamb, 1979*). During winter the oceans are warmer than land and consequently winds blow from the NE during the months of December to February.

2.2.1 Surface circulation and wind pattern

The entire northern Indian Ocean comes under the influence of the seasonally changing monsoon gyre. *Wyrtki* (*1973*) has reviewed the circulation of the Indian Ocean. The most important characteristic of this is the biannual seasonal wind and surface circulation reversals associated with the SW (June to September) and NE monsoons (November to February) (Fig. 2.1). A prominent signature heralding the onset of the summer monsoon is the cross equatorial wind component in the northern hemisphere. During May this low level flow progressively moves northward along the east African coast where it accelerates and produces the Somali Jet - the strongest persistent low-level wind in the world. During the SW monsoon, the winds are

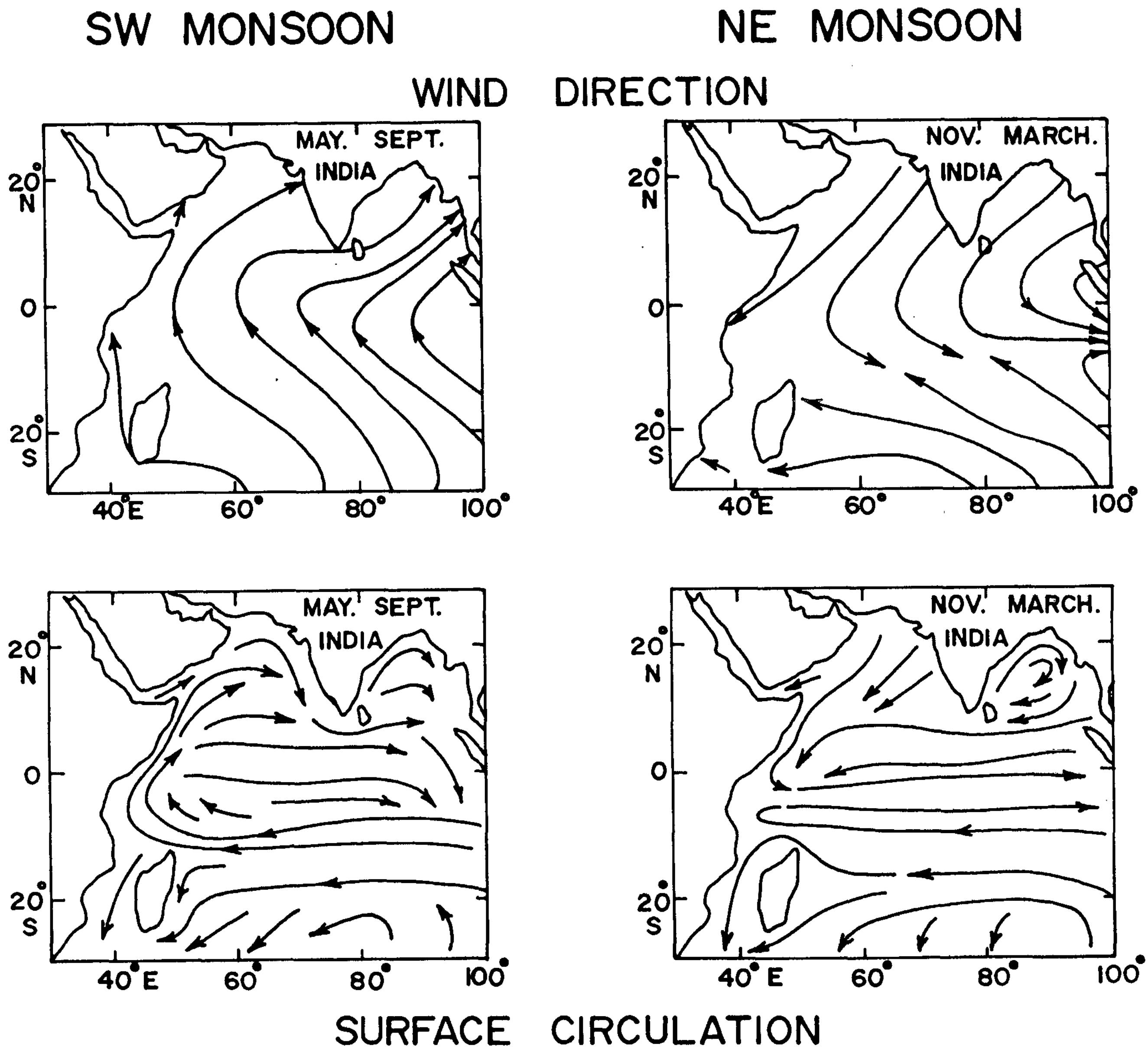


Fig. 2.1 Figure showing wind direction and surface circulation in the northern Indian Ocean during the southwest (SW) and northeast (NE) monsoon period.

strongest and blow consistently from west to southwest with speeds ranging between 10 to 13 m s^{-1} and reaching up to 25 m s^{-1} during gusty spells. During winter, winds blow from the northeast with windspeed around 8 m s^{-1}. In between the monsoons, the winds are weak and variable.

During the NE monsoon period, circulation is only moderately developed. Water movements north of the equator are from east to west which forms the NE monsoon current. The surface flow does not penetrate the thermocline and striking upwelling areas do not develop. The flow starts developing in November, reaches its greatest strength in February and subsides in April. A strong branch of this current turns north and flows along the west coast of India, carrying low salinity water from the Bay of Bengal into the eastern Arabian Sea.

With the onset of the SW monsoon the water circulation is reverse and starts flowing towards the east along the equator and north along the Somali coast and reaches its greatest strength in July. The monsoon current, Somali current and the South Equatorial current form a very strong wind driven gyre with surface currents much stronger than during the NE monsoon period.

In the Bay of Bengal, cool winter winds lower the SST at the head of the Bay and a anti-clockwise gyre is formed. It persists till April and occupies the entire Bay. During the SW monsoon period the surface waters flow towards the eastern and northeastern sides, and the waters pile up against the coast resulting in cyclonic circulation at the head of the Bay up to September. Fresh water discharge from the rivers cause low salinity water to be formed which covers the whole of the Bay of Bengal. This low salinity waters has been studied in detail by *Murthy et al* (*1990*) and they report that the fresh water from the Ganges-Brahmaputra rivers form a thin layer of about 25-40 m and bifurcate into two lobes, one flowing along the east coast of India and the other lobe flowing in the central part of the Bay of Bengal. The eastern lobe is further strengthened by the fresh water influx from the rivers draining into the Bay of Bengal along the east coast of India like the Mahanadi and the Godavary.

The annual freshwater input to the Bay of Bengal is shown in Table 2.1. The freshwater from rivers dilutes the upper 25 m of the water column in the Bay of Bengal by 5 parts per thousand. Further dilution is by rainfall, the excess of rainfall over evaporation being approximately 3500 km^3 (*Qasim, 1977*). The immense freshwater discharge leads to the development of an almost estuarine type of circulation in the northern part of the Bay (*Murthy et al., 1990*). Low salinity waters of the Ganges-Brahmaputra over-rides the high salinity waters of the Bay of Bengal. The upper mixed layer depth is very shallow and is between 30 and 50 m in the northern and central part of the Bay. Strong stratification and a shallow mixed layer keeps the low saline waters derived from the rivers on the surface and is thus able to transport it over a longer distance thus affecting a wider area. The depth of the mixed layer decreases from more than 75 m in the southern Bay of Bengal to less than 25 m in the northern Bay of Bengal (*Murthy et al., 1990*).

2.2.2 Upwelling

The vigorous atmospheric and oceanic circulation during the SW monsoon causes intense upwelling in several places. Large parts of the Somali current are recirculated in an intense eddy, the center of which is 300 km offshore (*Wyrtki, 1973*). The upwelling is most intense between 5^{o} and 11^{o} N latitude and subsurface waters with temperatures below 20^{o} C reaches the sea surface (*Warren et al., 1966*). Phosphate and nitrate concentrations reach more than 10 microgram at l^{-1}.

Strong winds also blow parallel to the coast of Arabia during this period and cause upwelling. This upwelling is different from the Somali upwelling in that no strong currents develop parallel to the coast. The upwelling here is greatest 300 km away from the coast and can be described as open ocean upwelling (*Brock and McClain, 1991*).

During the SW monsoon, upwelling also takes place along the west coast of India (*Sharma, 1978*). The 20^{o} isotherm rises to less than 50 m depth in July and August. Upwelling begins in the southern part and propagates northwards until about end of May. The cause of this upwelling is not yet well understood. Moreover, the winds blow in the wrong direction to produce upwelling during the SW monsoon. According to *Sharma* (*1978*) upwelling in this region may be due to geostrophic adjustment.

Only weak upwelling has been recorded in the Bay of Bengal and that too during the premonsoon period (*Qasim, 1977*). Strong monsoon winds displace the surficial waters, but it is replaced by fresh low saline waters rather than nutrient rich deep waters (*Rao, 1977; Rao and Sastry, 1981; Murthy et al., 1990*). Further, the strong monsoon currents flowing parallel to the coast during the monsoons keep the upwelled waters close to the coast. During the NE monsoon period the wind direction is not favorable for upwelling along the east coast of India. However, weak upwelling may occur along the Burmese coast.

From satellite photographs a western boundary current has been identified along the east coast of India up to 15^{o}N during February, March and April (*Legeckis, 1989*). It then veers off to the right and flows into the central part of the Bay. No data is available to confirm the existence of this current during the monsoon due to intense cloud cover. Weak upwelling during the premonsoon periods which occur along the east coast of India may be probably due to this boundary current.

2.2.3 Primary productivity patterns

As yet there is no comprehensive study of productivity in the Arabian Sea or Bay of Bengal and previous studies have always been separated in time and space (*Qasim, 1977; Babenard and Krey, 1974*). There are no published results on seasonal and interannual variations in primary productivity. However, the following broad conclusions can be made.

The Arabian Sea is an area of high productivity, averaging twice that of the world oceans (*Ryther and Menzel, 1965*). The regions of upwelling and associated eddies are areas of high productivity as they replace warm surface waters depleted of nutrients with cold nutrient rich waters from the sub-surface regions across the barrier of the thermocline. Highest primary productivity values have been reported mainly along the upwelling regions of Somalia, along the Arabian Coast and off the SW coast of India (*Sastry & D'souza, 1972*). The non-upwelling areas of the Arabian Sea also show high productivity values as the subsurface waters just below the euphotic zone contain very high nutrient concentrations. Any vertical perturbation like wind induced upwelling or divergence at current boundaries, internal waves or simple wind mixing can introduce nutrient rich subsurface waters to the photic zone and stimulate productivity. One of the characteristics of phytoplankton production in the Arabian Sea is its patchy distribution. Dense plankton blooms several kilometers long can lie adjacent to unproductive waters. Summaries of primary productivity in the northern Arabian Sea suggests annual rates between 200 to 400 gC m^{-2} y^{-1} and daily rates exceeding 2 gC m^{-2} d^{-1} reaching up to 6 gC m^{-2} d^{-1} (*Babenard and Krey, 1974; Qasim, 1982*). In the areas of upwelling and along the north western coast of India, productivity values exceed 200 gC m^{-2} y^{-1}. In the oligotrophic areas of the central Arabian Sea productivity values are only 30 gC m^{-2} y^{-1}. Most of these observations were made before the modern clean techniques were introduced and therefore these values may be underestimates of true primary productivity values.

Primary productivity patterns were studied in the Arabian Sea with the help of the Coastal Zone Color Scanner (CZCS) images (*Banse and McClain, 1986*). Eighty per cent of the total annual productivity in this area may be during the months of July, August and September. Although the area of the Arabian Sea is too small to account for more than a few percent of total world oceans primary production, it may contain a large fraction of new and export production.

During the premonsoon season, surface waters of the Arabian Sea are strongly stratified and exhausted of nutrients. Massive blooms of blue-green nitrogen fixing algae *trichodesmium* sp. proliferate (*Devassy et al., 1977*). Bacterial degradation of these blooms release nutrients which can support other algal species as well as provide food for heterotrophs.

Productivity studies in the Bay of Bengal are still more sparse. Here the levels of primary productivity, though higher that of the Indian Ocean, is still much lower than the Arabian Sea (*Qasim, 1977*). *Rao* (*1977*) examined the thermohaline structure in the upper layers and concluded that productivity in the Bay of Bengal is low because of the stability developed due to strong halocline in the upper layers which inhibit the process of nutrient replenishment by vertical transfer.

Zooplankton abundance varies strongly with maximum values reported near coastal and upwelling areas. The zooplankton biomass of the Arabian Sea is very high (*Qasim, 1977*). Compared to the other oceans the species diversity of mesopelagic fishes and zooplankton in the Arabian Sea is very less. The Bay of Bengal has moderate values compared to the rest of the Indian Ocean but lesser than that of the Arabian Sea. A strong oxygen minima with oxygen content less than 0.2 ml/l probably plays a major role in controlling horizontal and vertical distribution and migration of zooplankton and mesopelagic fishes in this area.

2.3. Recent Sediments and sedimentary processes in the northern Indian Ocean

Since lithogenic matter in the oceans are derived from the adjacent land, the geology of the Asian and African continents has a strong bearing on the type of minerals present in the northern Indian Ocean.

2.3.1. General geology of the surrounding landmass.

The Indian subcontinent can be divided into three divisions the Peninsular shield, the Indo-Gangetic Plains and the Himalayan fold belt (*Krishnan, 1968*). The Indian peninsular shield is a stable part of the crust and formed part of the ancient Gondwanaland. It is composed of metamorphic schist and gneisses. Towards the end of the Mesozoic era there was a passive eruption of volcanic basalts in the western part of the peninsula which have formed the Deccan Traps. The western and eastern ghats are the most prominent hill ranges along the western and eastern coast of India. Sedimentary rocks have a limited geographical extent and are mostly found along the coastal areas as detached outcrops.

The Indo-Gangetic plains are formed mainly due to denudation of the Himalayas during the Quaternary era. They are composed of an enormous thickness of sand and clay layers and extend from Sind on the western side across northern India to the Brahmaputra valley in the east. The soft sedimentary rocks in these plains are easily eroded and supply substantial amounts of suspended load to the numerous major rivers which runs across it.

The Himalayas are the highest mountain chains present on the surface of the Earth. They have formed due to the collision of the Indian and Eurasian plates during the Cenozoic Era. The Himalayan fold belt consists of a granitic core and is covered with sedimentary deposits laid down in the former Sea of Tethys. An extension of this fold belt runs towards the east into the Arakan Yoma mountains of Burma and towards the Zagros mountains of Iran to the west. Denudation of the Himalayas have supplied enormous volumes of sediment to the Indus and Bengal Fan via the Indus and Ganges-Brahmaputra river systems.

The northern part of Africa is composed of Paleozoic rocks with recent volcanic rocks associated with the African Rift System. Volcanic rocks are also found towards the east of Bay of Bengal associated with the Andaman Back arc systems.

The main input of lithogenic material to the northern Indian Ocean is fluvial sediments derived from the Indian Subcontinent. Significant eolian fluxes, mostly in the western Arabian Sea, are from the Arabian Peninsula while the rivers draining the Burmese Peninsula deliver their suspended load into the Andaman Sea.

2.3.2. Freshwater and Suspended sediment discharge into the northern Indian Ocean:

Fig 2.2 shows the seasonal discharge pattern of the principal rivers draining into the northern Indian Ocean. Most of the discharge is during the SW monsoon period as the catchment areas of many of the rivers receive maximum rainfall during this period. The rivers draining the Himalayan region

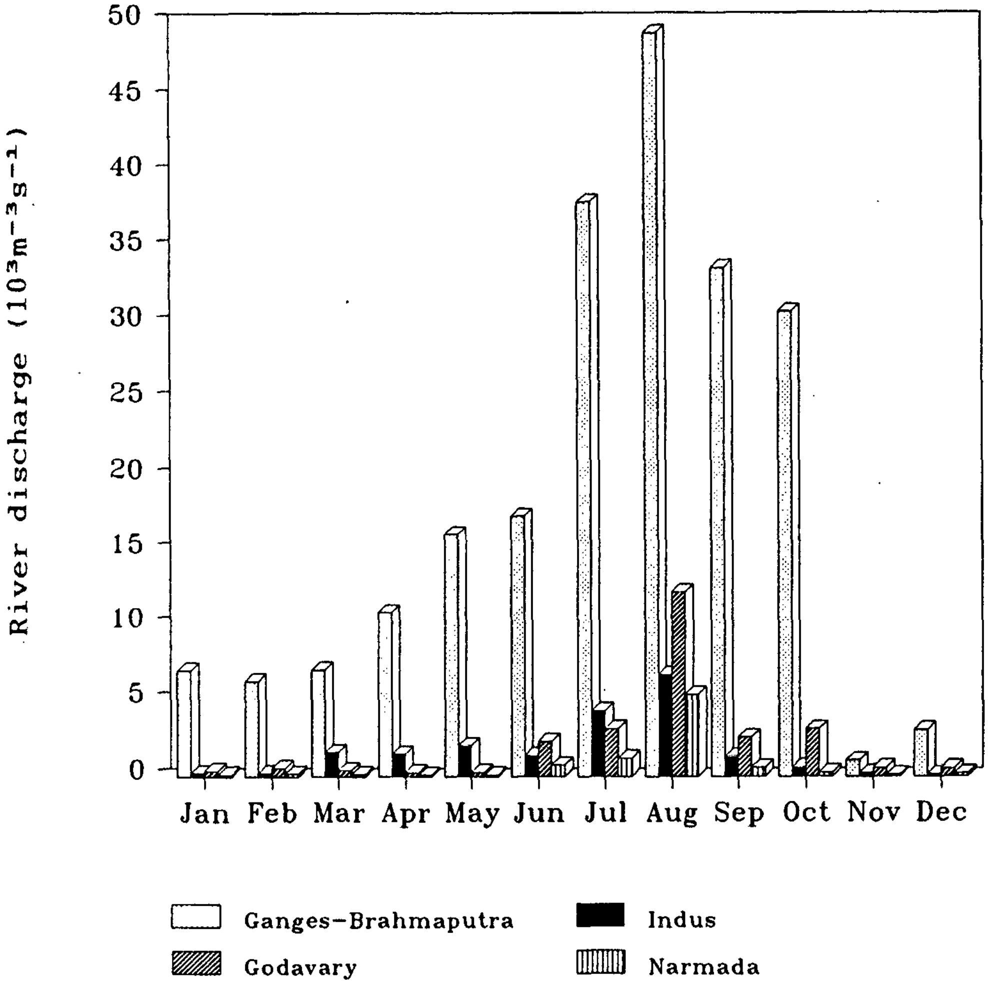


Fig. 2.2 Discharge pattern of major rivers draining into the Arabian Sea and Bay of Bengal. Data from UNESCO reports (197) and Rao (1975)

Table 2.1. Freshwater discharge of major rivers into the Arabian Sea and Bay of Bengal. Data from *UNESCO reports (1971)*, *Subramanian (1985)* and *Rao (1979)*.

River	Freshwater discharge m^{-3} s^{-1}
Ganges	14556
Brahmaputra	16186
Indus	6589
Godavari	3330
Mahanadi	2113
Krishna	2146
Cauvery	664
Narmada	1291
Tapti	570
Minor rivers	3000

Table 2.2. Suspended sediment discharge into the Arabian Sea and Bay of Bengal. Data from *Mlliman and Meade (1983)* and *Subramanian et al, (1985)*. The suspended sediment discharge values of *Subramanian (1985)* are much lower compared to *Milliman and Meade (1983)* but the authors do not explain the reasons for this. One of the reasons may be diversion of river water for agricultural purposes.

River	**Suspended Sediment discharge 10-6 tonnes per year**	
	Milliman and Meade 1983	**Subramaniam et al.,1985**
Ganges-Brahmaputra	1670	1171
Indus	45	
Irrawady	265	
Godavary	96	16.8
Mahanadi	62	2.07
Krishna		10.56
Cauvery		0.71
Narmada and Tapti	60	7.88

show and increase in discharge during the months of April and May due to melting of snow in the Himalayas (*Rao, 1979*).

At present about 100 million tonnes of fluvial material is discharged into the Arabian Sea every year. Of this about 60 million tonnes is from the Narmada and Tapti rivers. The Indus formerly discharged more than 440 million tonnes of suspended sediment annually, but with the construction of numerous dams and barrages the sediment discharge has dropped by 80% and the present annual suspended sediment discharge is less than 50 million tonnes (*Milliman et al., 1984*).

The annual suspended sediment discharge into the Bay of Bengal by rivers is more than 2000 million tonnes (Table 2.2). This forms roughly 14% of the world total of 18.3 x 10^{15} g (*Holeman, 1968; Milliman and Meade, 1983*). The Ganges-Brahmaputra alone contributes more than 1600 million tonnes (*Milliman and Meade, 1983*) of fluvial sediments each year. Suspended sediment discharge values for peninsular rivers reported by *Subramanian* (*1985*) (Table 2.2) are about a magnitude lower than that of *Milliman and Meade* (*1983*). The authors do not give any reasons for this. *Subramainans* (*1985*) data is more recent. Therefore, it is possible that the construction of many dams together with diversion of river waters for agriculture in the deltaic regions of these rivers in recent years could have significantly decreased the freshwater and suspended sediment discharge of these rivers in recent years.

2.3.3 Clay mineralogy of the river sediments

The Brahmaputra and Indus rivers drain the Himalayas while the other Indian rivers derive their sediments from a variety of rocks, especially those of the Deccan Traps. The mineralogy of the suspended sediments of rivers

Table 2.3. Clay mineral percentages in bed load of river mouths of major rivers of India. Data from Naidu et al., 1985

River	Illite	Chlorite	Kaolinite	Smectite
Ganges	75	24	00	00
Mahanadi	34	05	24	37
Godavari	10	04	09	73
Krishna	00	07	06	86
Cauvery	21	07	25	45
Narbada	11	6	7	76

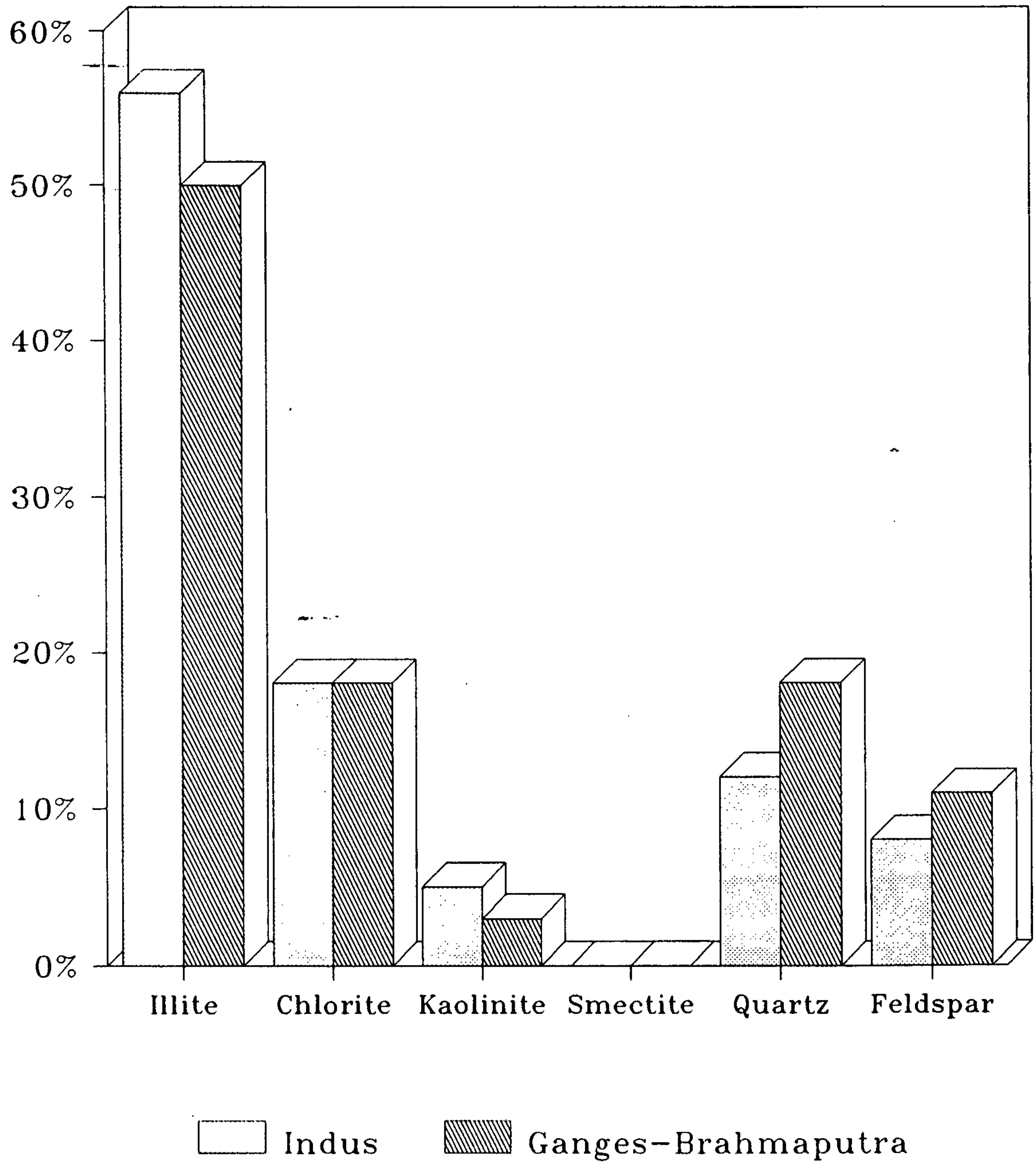


Fig. 2.3 Clay mineral percentages in suspended sediments of the Ganges-Brahmaputra and Indus rivers (From Konta, 1985)

therefore show a wide variation in composition and percentage of different minerals (Table 2.3)

Naidu et al. (*1985*) have studied the clay mineralogy of the < 2 micrometer fraction of the bed load of Indian rivers (Table 2.3) and show that illite and chlorite comprise almost 100% of the clay minerals in the rivers draining the Himalayas while smectité and kaolinite is not detectable. *Konta* (*1985*) has reported that illite, quartz and chlorite are dominant in the suspended load of the Ganges-Brahmaputra rivers and during peak discharge period illite and chlorite percentages are more than 80% (Fig. 2.3). According to *Konta* (*1985*) smectite and kaolinite comprise less than 3% of the total minerals present in the Ganges-Brahmaputra and Indus rivers. In contrast, the peninsular Indian rivers have very low illite percentages and high smectite percentages (Table 2.3). The main rivers of the Indian peninsula, the Godavari Narmada and Krishna have their catchment basins in the Deccan Trap area and the bedload of these rivers contain over 70 per cent smectite.

2.3.4. Eolian sediment input

Due to the SW and NE monsoon winds considerable amounts of eolian dust is supplied to the Arabian Sea from the arid regions surrounding it. The highest frequency of haze in the world oceans is reported in the Arabian Sea (*McDonald, 1938*). *Goldberg and Griffin* (*1970*) have calculated that about 80 million tonnes of such sediments were being deposited in this area each year. Although much of the sediments are being accumulated in the western Arabian Sea they have presented data that shows eolian dust are also being deposited in the deeper waters of the SW coast of India. Recently, *Sirocko* (*1989*) has estimated that between 115 to over 200 million tonnes of eolian sediments are being deposited in hemi-pelagic sediments of the entire Arabian Sea.

Goldberg and Griffin (*1970*) estimated that the annual eolian dust input to the Bay of Bengal is about 20 million tonnes which is less than 1 per cent of the sedimentary input from rivers.

2.3.5 Surficial sediments of the Arabian Sea and Bay of Bengal

Coarse-grained low carbonate sediments with high abundance of quartz and feldspars are found in the northern areas of the Arabian Sea and reflect terrigenous influx, primarily from the Indus River. Carbonate content of the sediments increase gradually towards the south reflecting the waning influence of terrigenous sediment supply southwards (*Kolla et al., 1976c*). Towards the southern portion of the Arabian Sea carbonate percentages are high on ridges and become less than 10% at 4800 m reflecting bathymetric control due to dissolution of calcium carbonate below the calcium compensation depth. High opal concentrations are observed in the western Arabian Sea below the Somali-Arabian upwelling areas.

The sediments of the Bengal Fan is largely derived from denudation of the Himalayas. Turbidites are generated by slumping of unstable delta fronts during low sea-level stands and channelled through the Swatch. Apart from this dark grey organic carbon rich muds are derived from the more anoxic part of the upper slope of the east coast of India and Sri Lanka. Biogenic turbidites in the lower fan is derived from areas of significant pelagic sedimentations like topographic highs and sea mounts. There are four distinct factors which have influenced sedimentation in this area in the past; Himalayan uplift in the Miocene and Pleistocene, sea level fluctuations, local tectonic effects of intraplate deformation and probable lobe/ channel switching.

Calcium carbonate in the Bay of Bengal is primarily of biogenic origin and its distribution reflects dilution by terrigenous material. High carbonates values are observed on the Ninety East Ridge and south of India and reflect shallow bathymetry and low clastic deposition. Low carbonate values in the SW and SE portion of the Bay of Bengal in spite of low clastic deposition reflects dissolution due to increasing water depth.

High amounts of organic carbon occur in the sediments of the Arabian Sea particularly in the westerly, northerly and easterly areas close to land. They reflect primarily the high biological productivity of the overlying waters. However, it is along the Indian continental margins between 200 to 1500 m depth that the highest organic carbon values (5 to 20%) are reported (*Paropkari et al., 1992*). High organic carbon along the continental slope regions in this area is due to impingement of low-oxygenated waters on the slope and high sedimentation rates. The reason for the higher degree of preservation of organic carbon along the Indian margin (15 to 80 times higher than off the Arabian Peninsula) is not well understood (*Paropkari et al., 1992*).

2.3.6 Clay mineralogy of the surficial sediments

Nair et al. (*1982a,b*) have studied the sediment distribution on the western continental shelf of India. They postulated the existence of a tidal barrier off Gujarat which prevents clay sized minerals from the Indus from being deposited on the continental shelf of India. Smectites (over 60%) derived from the Deccan Traps are the dominant clay minerals in the inner shelf. Marked compositional differences in the mineralogy of the inner (smectite rich) and outer (illite rich) shelf clays show that inner shelf clay sources are contemporary rivers draining the west coast of India while outer shelf sediments are derived predominantly from the Indus river (*Ramaswamy, 1985; Ramaswamy and Nair, 1989*).

Goldberg and Griffin (*1970*) and *Kolla et al.,* (*1976a; 1981*) have described the clay minerals found in the deeper parts of the Arabian Sea. Illite is the dominant mineral over most of the area. Smectites are dominant near the Indian margin while palygorsite, derived from the Somali-Arab deserts, are predominant in the western Arabian Sea.

Several workers (*Gorbunova,, 1966; Goldberg and Griffin, 1970; Venkatarathnam and Biscaye, 1973; Kolla and Biscaye, 1977; Kolla et al., 1976; 1981; Kolla and Rao, 1990*) have discussed the significance of clay minerals quartz and carbonate distribution in the deep Bay of Bengal. Based on distribution of clay minerals three provinces can be distinguished in the surface sediments of the Bay of Bengal (*Venkatrathnam and Biscaye, 1973*). The Deccan province which lies in the western and SW part of the Bay of Bengal is characterized by >60% smectites and <20% illite, less than 10% each of kaolinite and chlorite. The smectite rich sediments are derived from the Deccan Trap soils through the Godavari and Krishna rivers. The Ganges-Brahmaputra province has the same extent as the Bengal Fan proper, and has >30% illite, <30% smectite and 10 to 15% each of chlorite and kaolinite, kaolinite being somewhat lower than chlorite. The Indonesian province is found on the Ninety East Ridge and southernmost areas of the Bay of Bengal. They are characterized by high smectites (>53%) derived from the alteration of either in situ basalt volcanics and/or volcanic ash blown from the Indonesian Islands. The dispersal of sediments within and between the provinces has been affected primarily by surface oceanic circulation and by bottom and turbidity currents. Based on clay and heavy mineral analysis *Kolla and Rao* (*1990*) have suggested that sediments of the western Bengal

Fan have been derived from peninsular Indian rivers while the rest of the fan are derived from Himalayan rivers.

2.3.7 Sedimentation rates in the northern Indian Ocean.

Till recently there was not much data on sedimentation rates in the Arabian Sea. *Zobel* (*1973*) in her study of the sediments of the India and Pakistan continental margins gave a rough estimate of 1-2 cm/1000 years in the deep basin and up to 40 cm/1000 years on the upper continental slope. Sedimentation rates are lower off southern India than in the northern portion of the continental margin of India. Recently, *Sirocko* (*1989*) has determined sedimentation rates on 39 turbidite free hemi-pelagic cores north of 10° S. According to him during the Holocene bulk accumulation rates are highest off the Oman coast (> 20 cm/1000 years) and decrease to less than 3 cm/1000 years off the southwest coast of India.

There are not many reliable reports of sedimentation rates in the Bay of Bengal. According to *Sarin et al.* (*1979*) sedimentation rates decrease from 3 cm ky^{-1} in the northern Bay of Bengal (17° Lat.) to less than 0.5 cm ky^{-1} near the equator.

Chapter 3

Sedimentation Traps, Sample Collection and Analysis

3.1 Introduction

Particulate matter in the oceans can be placed under two categories; suspended particles (which do not settle) and settling particles. Suspended particles can be collected in any desired quantity by filtering sea water, collected either by conventional bottle samplers or by pumps. These methods cannot be used to collect settling particles as they are very rare. Filtration of extremely large volumes (thousands of liters) of sea water can collect large settling particles in quantity but is not very practical as the ship has to remain at station for the entire duration of sampling.

The most direct method available for sampling settling particles is the particle interceptor trap, commonly known as sediment traps. A sediment trap is a container deployed on a mooring in the water column that intercepts and collects particles falling into it from the overlying water. The container can later be retrieved to collect the sample.

A sediment trap is designed to open its horizontal aperture of a known area, and to intercept and collect a falling mass of particles, for a known duration. The mass flux of particles can be represented as $F = ma^{-2}t^{-1}$ where

F= Mass flux

m= mass of particles collected

a= area of collector

t= duration of deployment.

The essential assumption is that all particles settle by gravity and the interception at the horizontal aperture of the trap involves no biases to the rate of settling of particles through it.

There are many types of sediment traps having different shapes, sizes, configuration and operating mechanisms. While various asymmetric shapes have been used it is generally agreed that traps with round cross section (cones and cylinders) are the preferred geometry (*Gardner, 1977*). Of these, cylinders collect the most unbiased samples under various flow conditions and there is not much loss of material due to adhesion to the walls, as in the sloping walls of a cone. When large samples are desired, cones or funnels are preferred as cylinders are unwieldy to deploy or recover. Also, it is easier to design a multiple sampler from a cone than a cylinder.

Before the last decade, the deep sea was considered to be one of the most stable environments of the world. The earlier samplers were designed to collect a single sample over relatively long periods (up to six months). Recent results have indicated significant temporal variations (*Deuser, 1986; Honjo, 1990*) in particulate fluxes indicating a need to collect continuous time-series samples at intervals of about two weeks.

Multiple samplers are less reliable and more expensive than single samplers. Reliability of samplers is very crucial as samplers are often left unattended for up to one year. Recent advances in microprocessor technology has enabled fabrication of sediment traps with a fairly good reliability. Multiple samplers are preferred as normally more than ten time-series samples can be obtained at a time. To collect the same samples using single samplers would require more than 10 cruises the cost of which can be prohibitively expensive.

Sediment traps can be deployed on taut bottom moored arrays or on a surface floating array. Floating traps are better in that sampling is not biased as the trap drifts along with the surface currents and therefore there is no flow of water relative to the mouth of the traps. However, a major disadvantage of using these traps is that they can only be deployed at shallow depths, as at depths of more than a few hundred meters shear is unavoidable.

3.2. Limitations of sediment trap methododlogy

Interpretation of sediment trap samples is not easy due to several important sources of potential error (*Bloesch and Burns, 1980; Blomquist and Hakanson, 1981*). Though sediment traps are admittedly an imperfect sampling tool, they remain the only direct and practical method available to measure the in situ flux of particulate matter. The data collected must be used carefully and within their limitations.

3.2.1. Hydrodynamic effects

The most difficult source of error may be due to the interaction of the trap structure with the flow of water and particles. Several studies have been conducted to understand the bias to be expected due to geometry of the container and baffle covering the mouth of the trap, tilt of the container and varying current regimes (*Gardner, 1980a,b; Bloesch and Burns, 1980*). These studies have indicated that if proper care is not taken, sediment traps may over or undertrap particles or may be biased to a particular size range. These errors are more in high energy environments. Sediment traps can be used with reasonable accuracy in bathypelagic and abyssal areas. The sediment traps used for the present study have been deployed in deep waters at least 700 m below the sea surface.

3.2.2 Swimmers

A potential source of error is the effect of living organisms entering the trap and interacting with the trap (*Lee et a., 1991*). Actively swimming zooplankton can consume or defecate into the trap adding or subtracting from the real flux. They may also die on contact with the poison filled in the collecting cups. This source of error is more on cone shaped traps as particle can be scavenged by organisms as they rest on the walls of the trap. This problem is especially acute in the upper 500m of the oceans where migrating zooplankton are found. Many methods have been tried to overcome this problem none have been very successful. The most satisfactory procedure is to painstakingly remove all swimmers from the sample under a binocular microscope. The samples used for the present study were taken from samples deployed well below the range of migrating zooplankton.

3.2.3 Poisons and preservatives

A very important problem especially in long term deployments is the degradation of the samples collected (*Knauer, 1991*). Use of poisons like mercuric chloride and sodium azide prevents bacterial degradation but not chemical degradation. Preservatives like formaldehyde or glutraldehyde limit both biological and chemical degradation. Preservatives however do not prevent dissolution of many soluble chemicals, but may complicate sample processing by interfering with some of the chemical analyses of interest. In many deployments only poisons are used. After extensive trials it was found that mercuric chloide in low concentrations is the best preservative.

3.3 Field methods

The samples used for this study have been jointly collected under a joint collaborative programme between the **National Institute of Oceanography, Goa, India** and the **Institute of Biogeochemistry and Marine Chemistry, Hamburg University, Hamburg, Germany**.

For collection of settling particles **PARFLUX MARK VI** traps were used (Fig. 3.1). Details of construction of this sediment trap are given in *Honjo and Doherty* (*1988*). Basically, the sediment trap is made of a PVC or fiberglass funnel mounted on an aluminium frame with series of polyethylene collection

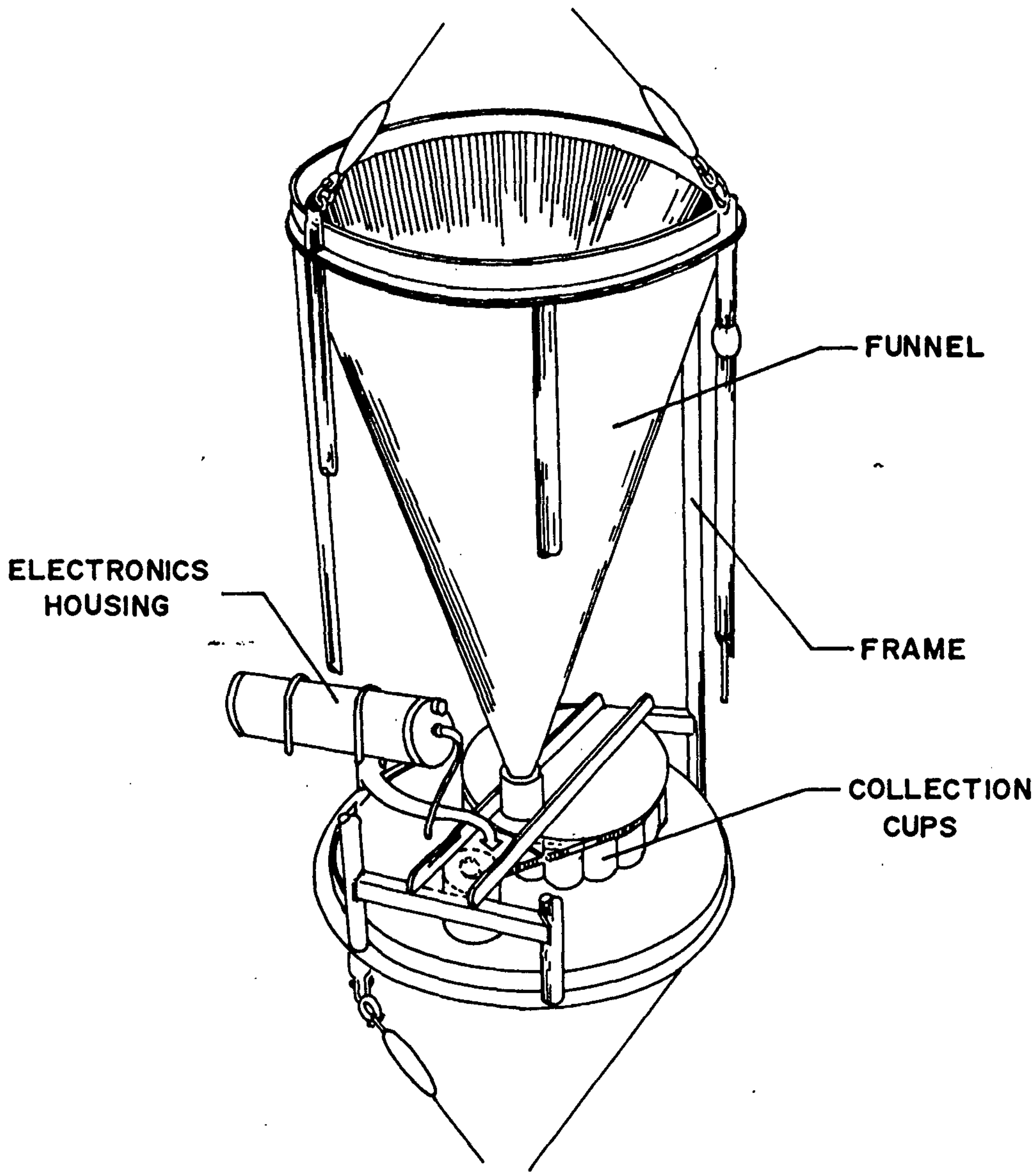
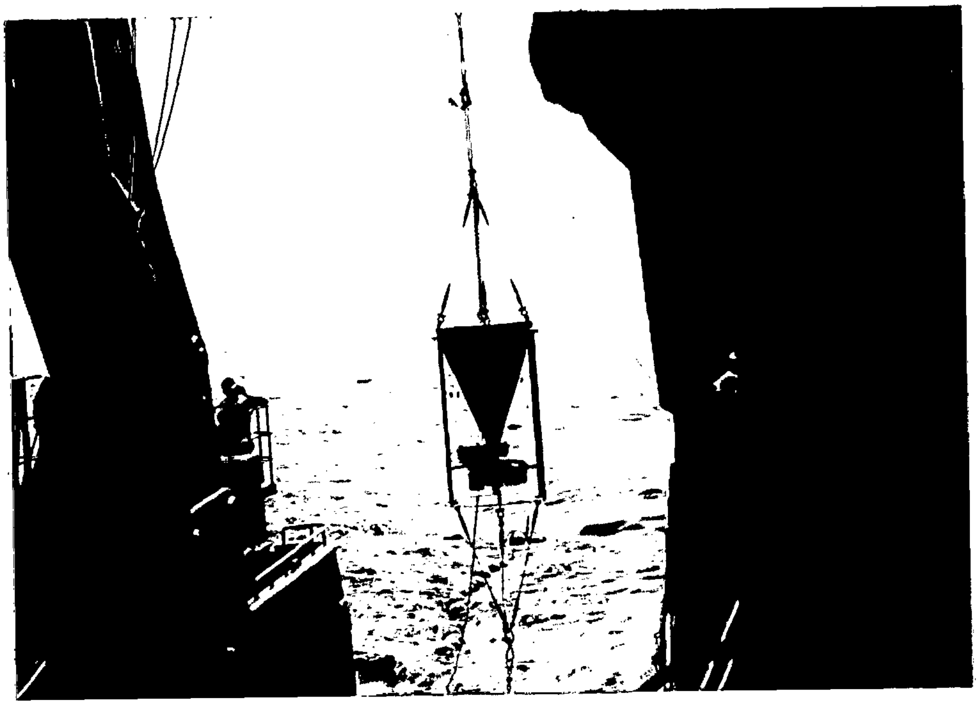
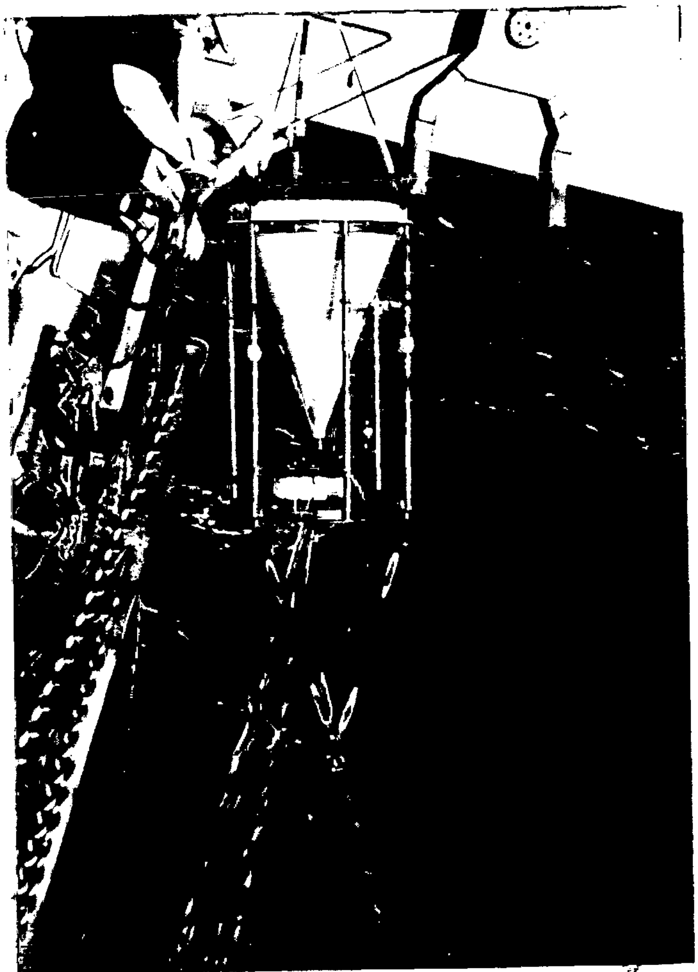


Fig 3.1 Details of the PARFLUX MARK VI particle interceptor trap (sediment trap). Baffles covering the funnel is not shown.



(a)



(b)

Fig 3.2 Photographs showing deployment (a) and recovery (b) of sediment trap moorings in the northern Indian Ocean. Glass spheres (with yellow hard-hats), used for providing buoyancy, is seen in the first photograph. Plastic bottles in which samples are collected and stored is seen at the bottom of the funnel cups at its base (Fig 3.1, 3.2). Thirteen sample collection cups are screwed onto a delrin geared disc at the base of the funnel. Each collection cup comes under the opening at the base of the funnel at a pre-determined time. The delrin gear is driven by a stepper motor controlled by a electronic timer housed in a pressure casing and mounted on the frame. Prior to deployment the data for controlling the movement of the collection cup is fed through a microprocessor. Also, prior to deployment the cups are filled with sea water containing elevated concentrations of sodium chloride (70 ppt) and low concentrations of mercuric chloride (3 ppt) to inhibit bacterial decay of the samples.

The traps are attached to the mooring with PVC coated steel bridles as shown in Fig 3.2 and 3.3. The mooring locations as well as depths at which the traps were deployed is shown in Table 3.1. Usually the traps were deployed at 1000 m from the sea surface and 1000 m above the sea floor. The reason for this is that most of the migrating zooplankton and small fishes live in the top 500 m and by positioning the trap well below this limit the problem of zooplankton entering into the traps and feeding on the collected material or dying in the trap contributing to the flux is avoided. The deeper traps are deployed 1000 m above the sea floor to avoid collection of resuspended sediments.

The buoyancy for the mooring is provided by glass flotation spheres (Fig. 3.3). The mooring is held in place by means of an anchor made of railway wheels having a combined weight of over 1 tonne. After the samples have been collected the anchor is released by means of an acoustic release and when the mooring surfaces it is tracked with the help of a radio transmitter and flashing lamp beacon.

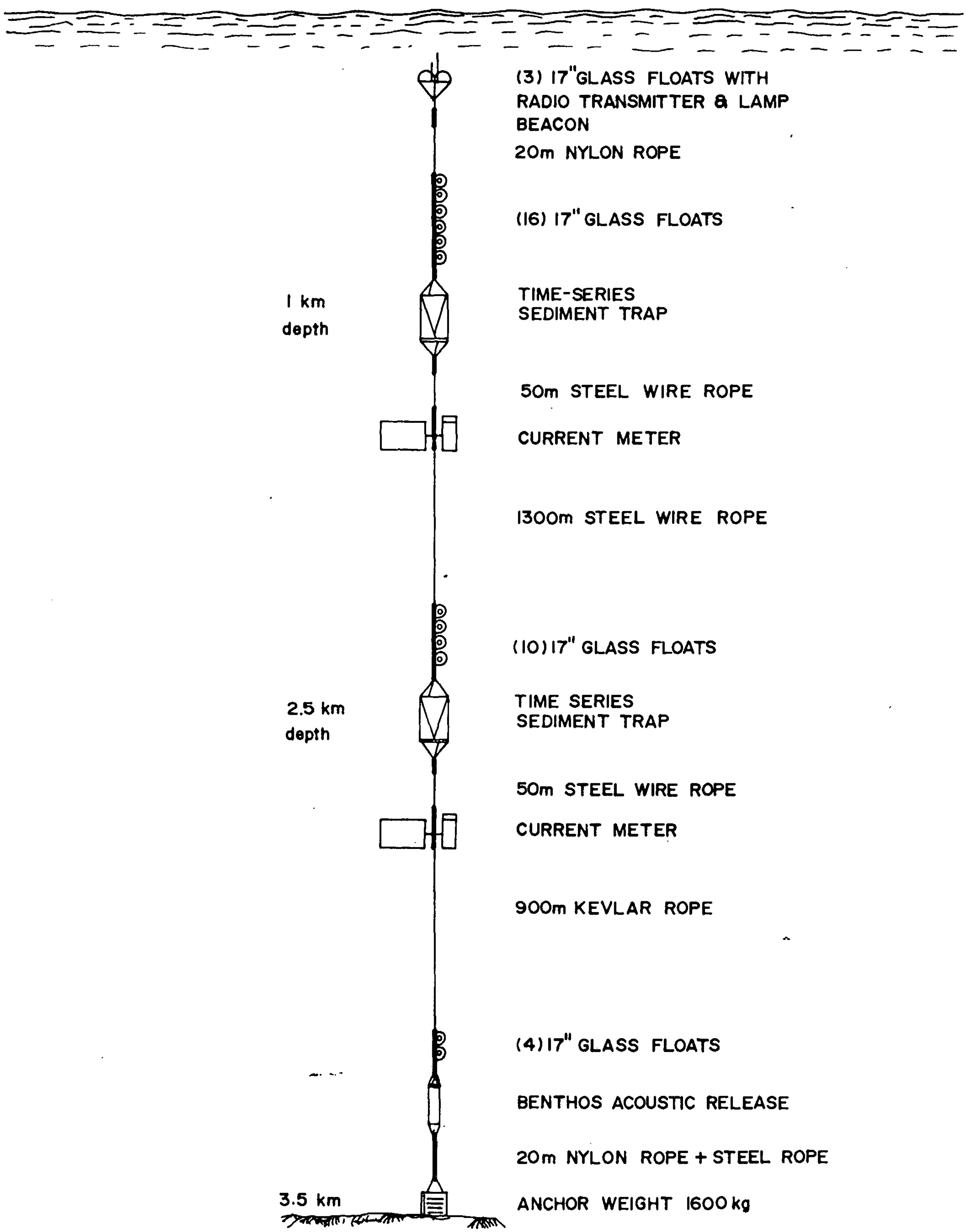


Fig. 3.3 Details of mooring used for deployment of sediment traps.

Table 3.1. Mooring location and sampling intervals in the Arabian Sea and Bay of Bengal.

Mooring	Position	Water depth	Deployment No.	Sampling Period From	Sampling Period To	Sampling interval days	Trap Depth in meters Shallow	Middle	Deep
EAST	Lat- $15^{0}30'$	3776	01	10.05.86	26.10.86	13.0	1395		2787
	Long $68^{0}45'$		02	20.11.86	02.05.87	12.6	1205		2764
			06	27.03.90	23.10.90	21.0	1692		2928
CAST	Lat- $14^{0}30'$	3920	01	10.05.86	26.10.86	13.0	0732		2914
	Long $64^{0}05'$		02	20.11.86	02.05.87	12.6	0834		2894
WAST	Lat- $16^{0}0'$	4016	01	10.05.86	26.10.86	13.0	1023		3024
	Long $60^{0}0'$		02	20.11.86	02.05.87	12.6	1051		3021
			03	12.05.87	21.10.87	12.5	1080		3033
NBBT	Lat $17^{0}6'$	2263	01	28.10.87	28.02.88	09.5	0809		1727
	Long $89^{0}5'$		02	18.3.88	06.10.88	14.5	0754		1790
			03	02.11.88	19.10.89	26.0	0968	1498	2029
CBBT	Lat $13^{0}9'$	3259	01	28.10.87	28.02.88	09.5	0906		2282
	Long $84^{0}1'$		02	18.3.88	06.10.88	14.5	0950		2227
			03	02.11.88	19.10.89	26.0	0955		2291
SBBT	Lat $04^{0}6'$	4017	01	28.10.87	28.02.88	09.5	1040		3006
	Long $87^{0}9'$		02	18.3.88	06.10.88	14.5	1036		3012

A total of six moorings with two sediment traps on each mooring were deployed in the Arabian Sea since May 1986 and in the Bay of Bengal since November 1987 (Fig. 1.1). Details about the water depth, sediment trap depth, duration of sampling etc are given in Table 3.1. The Arabian Sea moorings were deployed by RV Sonne in May 1986 and recovered and redeployed by ORV Sagar Kanya in November 1986 and May 1987. The Bay of Bengal traps were first deployed in the Bay of Bengal in November 1987 and recovered and redeployed by RV Sonne and ORV Sagar Kanya in May 1988, November 1988 and November 1989.

Due to unavoidable circumstances, participants from the National Institute of Oceanography could not participate in some of the cruises of R.V.Sonne. Therefore, samples for the SW monsoon period of the first year of deployment (1988) in the Bay of Bengal were not available for analysis. Detailed mineralogy and grain size of the lithogenic fraction has been described for the period between May 1986 to May 1987 for the Arabian Sea and between November 1988 to October 1989 for the Bay of Bengal.

Of the three moorings deployed in the Bay of Bengal during November 1988 only the northern and central traps could be recovered. The northern mooring had three time-series sediment traps while the central Bay of Bengal mooring had two traps.

As soon as the traps were recovered the sample collection cûps were removed and refrigerated. The samples cups were taken out and a rough estimate of the volume of material collected was measured. The pH of the supernatant waters was measured before decantation. The supernatant water was then filtered using a nuclepore filter having a pore diameter of 0.4 micrometer. The sample was poured out onto a clean plastic tray and examined under a binocular microscope. The sample was then sieved through a 1 mm plastic sieve. Both the greater and less than 1 mm fractions are then split into 4 aliquots using a rotary splitter (*Honjo et al., 1988*). A one quarter aliquot of the sample was further split into 4 parts of which two 1/16 parts were used for this study. These two parts were filtered on board using a preweighed nuclepore polycarbonate filter having a pore size of 0.4 micrometer.

3.4 Laboratory methods

The samples collected were split and shared between the two participating institutes. The split samples were nearly identical (splitting error <5%). In some of the cups the sample collected was very less and in such cases the entire sample was taken by one of the institutes. The samples were processed independently in the respective laboratories and data generated were cross checked and shared between the two Institutes. Data on total fluxes generated by the two institutes were averaged out to minimize splitting error and to avoid reporting different values. All the analyses reported in this thesis have also been been carried out by me and averaged out with similar results from our collegues in Germany. Results on lithogenic fluxes, mineralogy and grain size have been done exclusively by me.

The samples collected by the sediment traps after initial processing were analysed for total and component fluxes in a closed, air-conditioned, dust free room.

The following analyses were carried out:

3.4.1 Total flux

The filters were dried using an infrared lamp and repeatedly weighed till a stable weight was obtained. Weighing was done on a Mettler filter weighing balance which had a least count of 5 microgram. A Polonium 210 anti-static device was kept in the balance to remove static electricity on the filters which would otherwise have added considerably to the recorded weight. The total flux was calculated by using the formula $F = m\, a^{-2}\, t^{-1}$.

3.4.2 Carbonate

Carbonate was determined by gravimetric method. The weighed sample was transferred to a polyethylene container and treated with 35% v/v dilute acetic acid. After completion of the reaction the sample was filtered through a nuclepore filter and dried under an infrared lamp and reweighed. The weight loss was taken as equivalent to carbonate content of the sample.

3.4.3 Organic matter and organic carbon

Organic matter was measured by treating the sample with 30% hydrogen peroxide (H_2O_2). The temperature of the reaction was kept at around 40 °C by heating the container with an infrared lamp. After completion of the reaction (when the sample stopped bubbling) the sample was again filtered through a nuclepore filter and reweighed. The loss of weight was taken as organic matter. Organic carbon was calculated by dividing organic matter by 1.8.

For the Arabian Sea, organic carbon values were supplied by colleagues from Hamburg University, who made the measurements on replicate samples using a Carlo-Erba elemental analyzer (*Haake et al., 1993*). The values obtained by both the method is very similar and difference is less than 5% (Fig. 3.4)

3.4.4 Biogenic silica

The method used for determination of biogenic silica is a modification of the one used by *Eggimann* (*1981*). The sample was transferred to clean polyethylene containers and treated with 10 ml of sodium carbonate and kept in a water-bath at 90° C for 2 hours. Microscope examination of the sample showed that after 2 hours most of the diatoms and radiolaria had dissolved

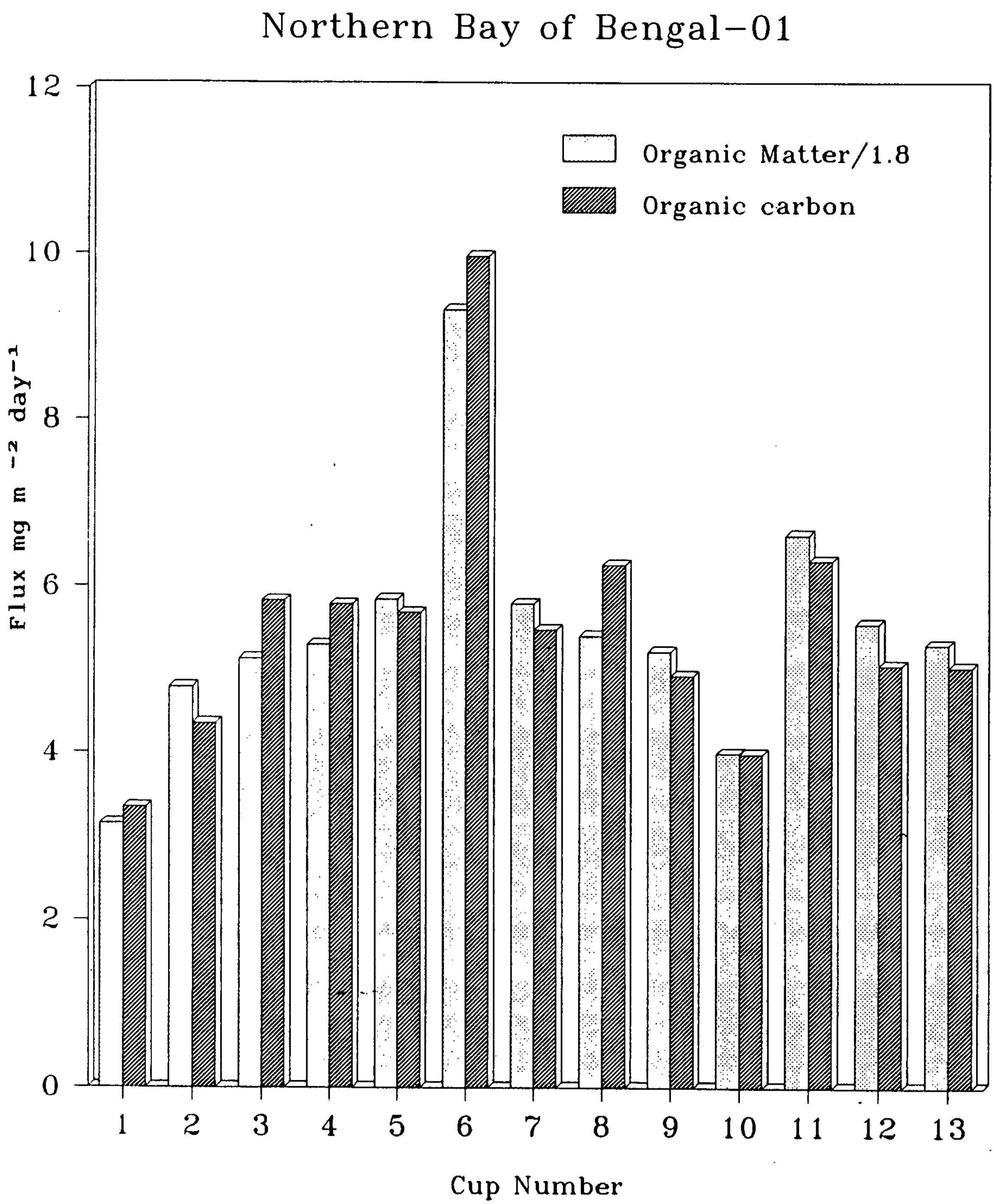


Fig 3.4. Relation between organic carbon values obtained by an elemental analyser (Carlo Erba) and by dividing organic matter percentages by 1.8.

except for a few resistant species of radiolarians. The sample was again filtered through a nuclepore filter and the leachate obtained was made up to 100 ml. Silicon in the leachate was measured using atomic absorption spectrophotometry after diluting to the required level. AAS technique was preferred to spectrophotometry as it is faster and less prone to contamination.

3.4.5 Lithogenics

The fraction remaining on the filter after leaching of the biogenic silica was weighed and taken as the lithogenic fraction.

The methods used for determination of the first order components, as described above, are straightforward and results obtained are accurate and repeatable. Determination of mineralogy and grain size of the lithogenic fraction is not so accurate as the results obtained are dependent on the sample preparation procedures as well as type of instrument used. Details of the methods used for determination of mineralogy and grain size along with their limitation are discussed separately in chapters 6 and 7.

Chapter 4

Total and Component Particle Fluxes in the Arabian Sea and Bay of Bengal

4.1 Introduction

A major fraction of particles in the ocean is of biogenic origin. Phytoplankton synthesizes carbon, silica and other essential elements from dissolved to solid form in the photic zone with the help of the Sun's energy. The amount of particles synthesized depends upon the primary productivity of that area. Primary productivity in the tropics is generally limited by availability of nutrients, usually nitrates (Fig. 4.1). Processes which introduces nutrients into the photic zone, like upwelling, river discharge etc., play a major role in particle production and biogeochemical cycling of elements in the ocean. The northern Indian Ocean experiences extremes in atmospheric forcing which lead to the greatest seasonal variability observed in any ocean basin. It is therefore of interest to study the effect of monsoons on particle production and fluxes in this area.

Most of the particles produced in the photic zone are rapidly remineralised but a small but significant fraction sinks out of the photic zone as detritus or fecal pellets. This includes the more refractory elements, skeletal material and also lithogenic particles which are not easily remineralised in the upper part of the water column. Labile particles which are enclosed within fecal pellets and are out of contact with the corrosive effects of sea water can also sink out of the photic zone.

There is much uncertainty regarding the amount of material being transported out of the photic zone by this process. A knowledge of this is necessary as settling particles control the biogeochemical cycling of several key elements which influences life on earth. For example, settling particles is

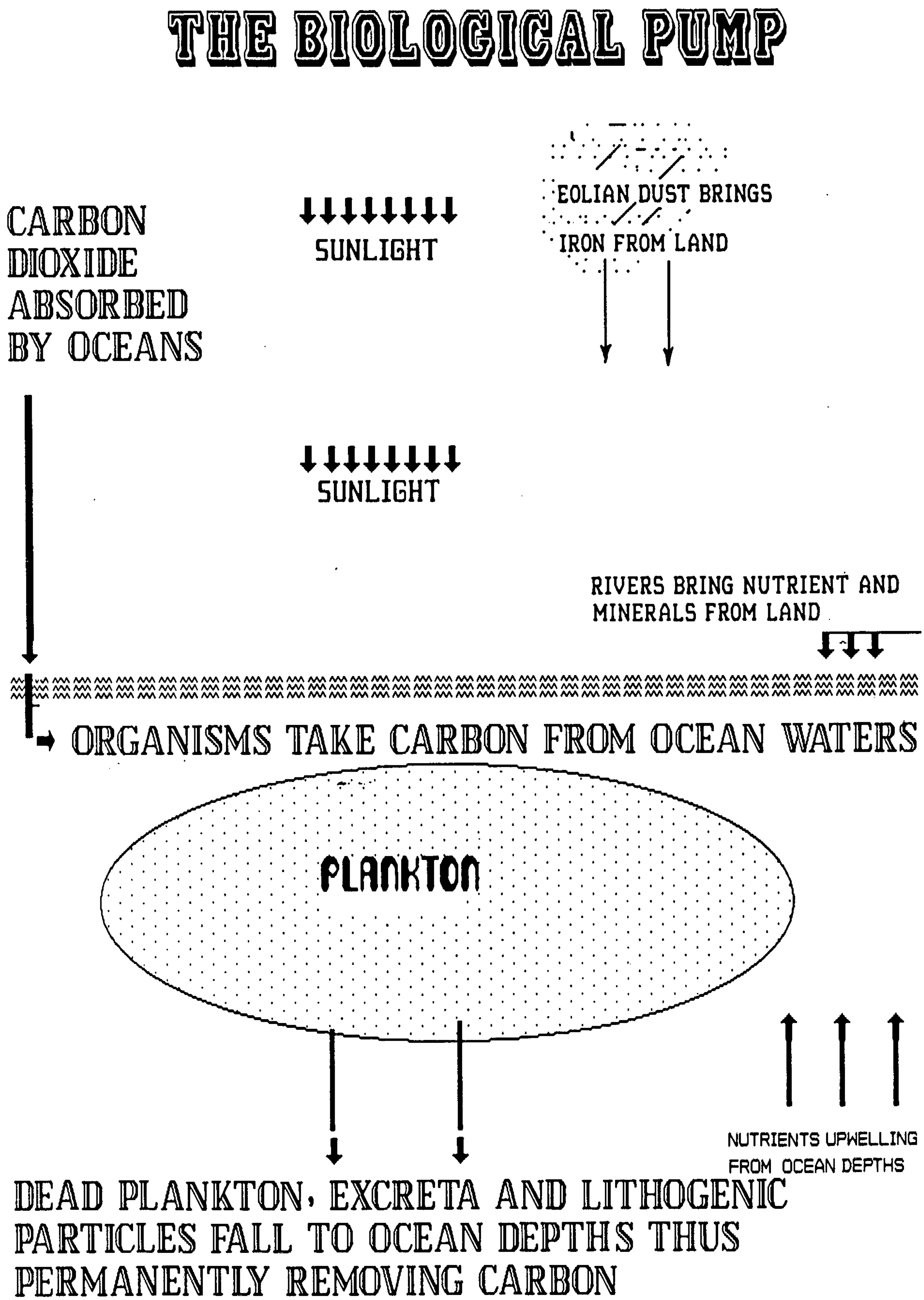


Fig. 4.1 Figure showing the working of the "Biological Pump". Phytoplankton synthesize particulate organic matter in the photic zone taking dissolved carbon from the ocean waters. A part of the organic matter sinks to the sea floor, therby removing carbon dioxide from the atmosphere. Generally, the limiting factor for the bilogical pump is availability of nutrients.

one of the key pathways by which carbon is removed from the atmosphere and sequestered within the oceans (Fig 4.1). The amount of carbon transported by the 'Biological pump', is estimated to be between 5 to 20 giga tonnes (*JGOFS Report, 1990*). To narrow down this uncertainty it is necessary to determine the flux of particles in coastal, eutrophic and oligotrophic areas of the worlds oceans (*Ramaswamy and Nair, 1992*). For this pupose year round sediment trap experiments have been carried out at several sites in the World's oceans (*Wefer, 1989*).

In the Sargasso Sea, variation in flux pattern are coupled to the annual cycle of primary productivity in the surface waters, peaking in early spring and showing lowest values in Autumn (*Deuser and Ross, 1980; Deuser, 1986*).

In the Black Sea, particle flux pattern are related to diatom, dinoflagellate and coccolith blooms (*Honjo et al., 1987*). In addition, detritus from rivers and shelf sediments advected to the trap sites at depth were also major components.

In the Panama Basin, particulate flux directly reflect variations in surface primary productivity. Peak fluxes here were related to a coccolith (*Umbelliospaera Sibogae sp*) bloom (*Honjo, 1982*).

In the equatorial Pacific, particle flux is at least a factor of two lower during periods of intense El Nino Southern Oscillation (ENSO) events (*Dymond and Collier, 1988*). During periods of greater nutrients availability, diatom productivity dominated over coccolith productivity.

In the Norwegian-Greenland area, highest fluxes are not in spring but during the fall between August and October (*Honjo et al., 1985*). It appears that zooplankton are responsible for retention of organic matter in the surface layers during the spring phytoplankton bloom.

In the Bransfield Straits, 90 per cent of the annual flux occur during the Austral summer. Peak fluxes here were related to vigorous feeding and defecation by krill swarms (*Von Bodungen et al., 1987*).

In the Weddel Sea, very low fluxes are reported with significant fluxes occurring only after the retreat of the ice sheet.

In the north Atlantic Ocean peak fluxes of particles occur during the spring bloom. The waters are charged with nutrients during winter but due to light limitation and turbulent conditions primary productivity is low. With adequate light conditions a plankton bloom is initiated during early spring which migrates slowly towards the North Pole (*Honjo and Manganini, 1993*)

Thus we see that there is a close relationship between particulate flux patterns and surface hydrography and biological processes in the oceans. The northern Indian Ocean with its seasonally reversing monsoons, stable intermonsoon periods, strong productivity gradient and periodic inputs of freshwater and mineral aerosols offers an excellent opportunity to study the effects of physical forcing together with freshwater and dust inputs on particle fluxes to the deep sea.

4.2 Material and methods

Six sediment traps were deployed at three locations in the eastern, central and western Arabian Sea in May 1986 (Fig.1.1). Each mooring had two traps one at approximately 1000 m and the other at 2500 m (exact depths are given in Table 3.1). The shallow traps in the eastern and western Arabian Sea malfunctioned. Hence, total and component flux data from all the deep traps and total flux data from the shallow trap in the central Arabian Sea is presented in detail in this chapter.

In the Bay of Bengal six sediment traps were deployed in the northern, central and southern parts of the Bay in November 1987. Scientist from India could not participate in two cruise of RV Sonne during the first year of deployment (1987-88) and hence I could not collect any samples for this period. During the second year (November 1988 to November 1989) the southern Bay of Bengal traps could not be recovered. Hence in this thesis detailed data on total fluxes, component fluxes as well as mineralogy and grain size is presented from three traps in the northern Bay of Bengal and two traps in the southern Bay of Bengal.

The data on total and component fluxes for the first year of deployment (November 1987 to November 1988) in the Bay of Bengal were shared between the two laboratories but this data is not presented in detail in this thesis. This data is used for mainly comparing the fluxes between the Arabian Sea and Bay of Bengal as high resolution data (11 days interval) is available from six traps at all the three locations.

The details of sampling depth, sampling interval etc. is given in Table 3.1. Methods used for determination of total and component fluxes are described in detail in chapter 3.

4.3 Results and discussions:

4.3.1 Total and component particle fluxes in the Arabian Sea

Total fluxes: The total annual fluxes to the deep eastern, central and western Arabian Sea traps during the first year of deployment are 23.5, 26.4 and 33.9 g m^{-2} day^{-1} of which between 53 and 70% is during the southwest monsoon period (Table 4.1.). All three sites showed strong seasonality with high fluxes during the monsoons (Fig 4.2). Individual fluxes varied by over two order of magnitudes from less than 5 mg m^{-2} y^{-1} to over 700 mg m^{-2} y^{-1} (Table 4.2 to 4.4). The highest difference between maximum and minimum fluxes are in the eastern and western trap which are characterized by their proximity to

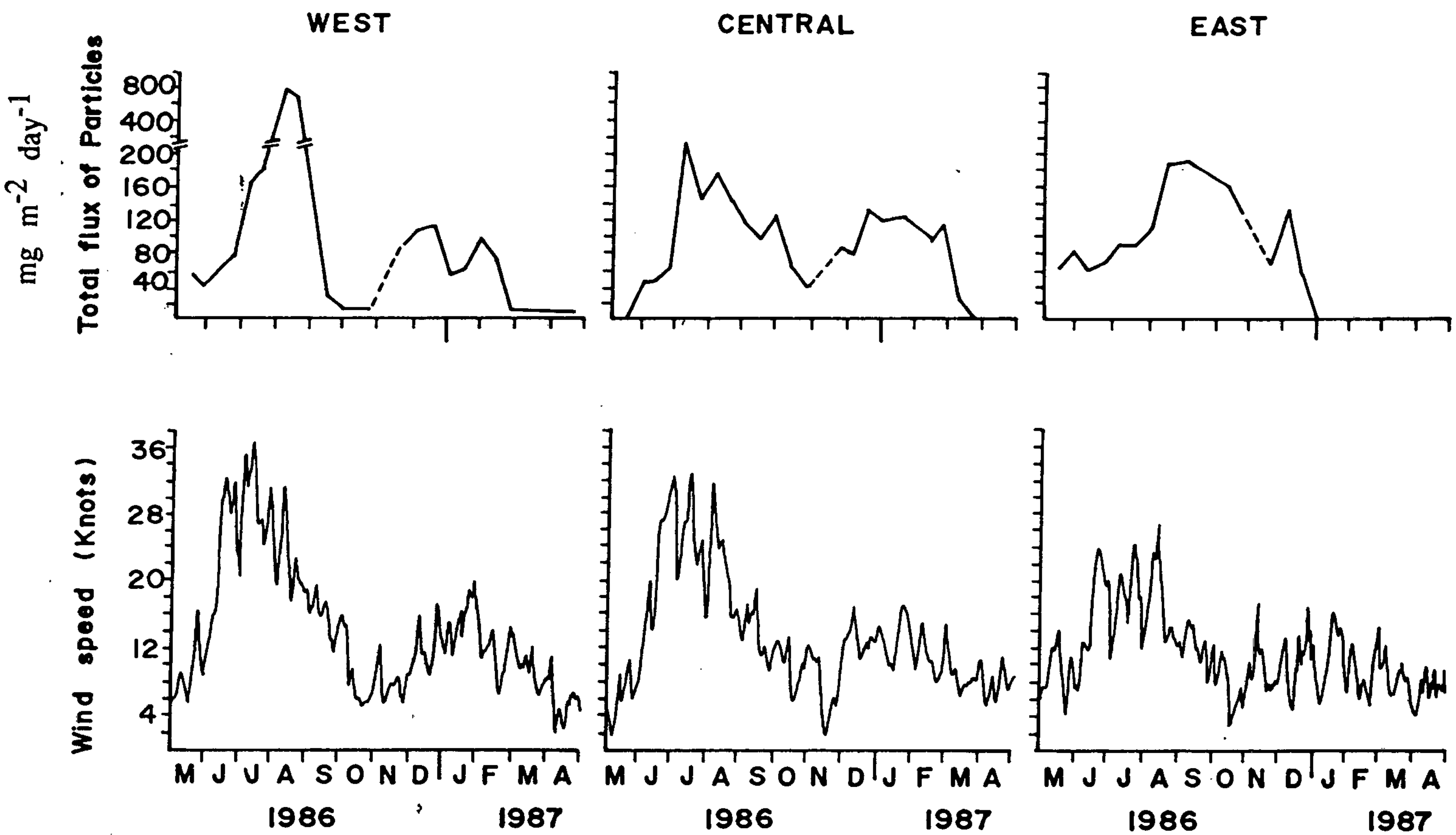


Fig 4.2 Particle flux and wind speed patterns in the eastern, central and western Arabian Sea. Both particle flux as well as wind speeds are high during the SW monsoon.

Table 4.1 Annual and seasonal fluxes of particulate matter and first order components to the Western (WAST), central (CAST) and eastern (EAST) Arabian Sea. All fluxes are in g m^{-2}

	Total Flux	Carbonate Flux	Opal Flux	C.org. Flux	Lith. Flux	Carb/ Opal	C.org./ C.Carb
WAST							
NE-SW	2.16	1.17	0.12	0.16	0.11	9.75	1.14
SW	23.56	12.10	6.08	1.18	2.10	1.99	0.81
SW-NE	2.91	2.13	0.30	0.14	0.28	7.10	0.55
NE	5.29	3.60	0.78	0.32	0.15	4.62	0.74
Annual	33.92	19.00	7.28	1.80	2.64	2.61	0.79
CAST							
NE-SW	0.85	0.49	0.06	0.06	0.07	8.17	1.02
SW	14.02	9.31	1.51	0.74	1.84	6.17	0.66
SW-NE	3.22	2.15	0.35	0.20	0.35	6.14	0.78
NE	8.36	5.55	1.16	0.53	0.79	4.78	0.80
Annual	26.45	17.50	3.08	1.53	3.05	5.68	0.73
EAST							
NE-SW	1.85	0.80	0.30	0.18	0.42	2.67	1.88
SW	14.68	7.27	1.81	1.00	3.60	4.02	1.15
SW-NE	6.25	3.40	0.79	0.33	1.35	4.30	0.81
NE	0.80	0.41	0.23	0.05	0.03	1.78	1.02
Annual	23.58	11.88	3.13	1.56	5.40	3.80	1.09

Table 4.2 Total and component fluxes to the eastern Arabian Sea.
All fluxes are in mg m-2 day-1

EAST 01 Deep

Period	<1 mm	>1 mm	Total	Carbonate	Opal	Lithogenic	C.org
10 May-23 May	57.47	0.94	58.41	21.0	12.2	15.8	7.6
23 May-05 Jun	82.73	4.02	86.75	40.8	10.7	16.7	7.1
05 Jun-18 Jun	58.58	2.82	61.4	21.5	10.6	16.7	6.0
18 Jun-01 Jul	67.78	1.06	68.84	32.5	10.3	10.7	6.7
01 Jul-14 Jul	91.06	1.27	92.33	47.0	10.3	19.9	10.1
14 Jul-27 Jul	88.86	4.02	92.88	45.7	10.5	16.1	5.7
27 Jul-09 Aug	108.95	6.88	115.83	62.6	11.5	24.8	7.7
09 Aug-22 Aug	185.2	13.4	198.6	96.5	22.7	45.7	12.0
22 Aug-04 Sep	186.4	8.4	194.8	97.5	21.0	46.5	12.7
04 Sep-17 Sep	177.06	4.37	181.43	83.2	22.0	47.2	13.2
17 Sep-30 Sep	165.3	2.81	168.11	72.9	20.4	49.1	10.8
30 Sep-13 Oct	155.6	1.43	157.03	68.1	20.1	45.1	10.7
13 Oct-26 Oct	123.49	1.79	125.28	55.8	15.9	36.5	8.6

EAST 02 Deep

Period	<1 mm	>1 mm	Total	Carbonate	Opal	Lithogenic	C.org
19 Nov-01 Dec	69.59	20.88	90.47	42.8	7.3	12.8	3.7
01 Dec-14 Dec	138.74	10.56	149.3	99.0	18.4	10.0	5.8
14 Dec-26 Dec	60.66		60.66	32.4	18.0	2.4	4.3
26 Dec-08 Jan	1.63		1.63	0.0	0.0	0.0	0.3
08 Jan-21 Jan	0.7		0.7	0.0	0.0	0.0	0.1
21 Jan-02 Feb	0.1		0.1	0.0	0.0	0.0	0.0
02 Feb-15 Feb	0.11		0.11	0.0	0.0	0.0	0.0
15 Feb-27 Feb	0.18		0.18	0.0	0.0	0.0	0.0
27 Feb-12 Mar	0.39		0.39	0.0	0.0	0.0	0.0
12 Mar-25 Mar	0.55		0.55	0.0	0.0	0.0	0.0
25 Mar-06 Apr	0.14		0.14	0.0	0.0	0.0	0.0
06 Apr-19 Apr	0.96		0.96	0.0	0.0	0.0	0.1
19 Apr-01 May	0.08		0.08	0.0	0.0	0.0	0.0

Table 4.3 Total and component fluxes to the central Arabian Sea.
All fluxes are in mg m-2 day-1

CAST 01 Deep

Period	<1 mm	>1 mm	Total	Carbonate	Opal	Lithogenic	C.org
10 May-23 May	1.22	0.37	1.59				
23 May-05 Jun	42.02	3.05	45.07	25.7	4.3	5.3	3.4
05 Jun-18 Jun	43.79	4.73	48.52	27.7	4.8	5.6	4.0
18 Jun-01 Jul	58.27	3.33	61.6	3.4	4.0	13.4	3.3
01 Jul-14 Jul	209.17	9.21	218.38	146.6	23.5	20.7	11.3
14 Jul-27 Jul	137.87	9.43	147.3	97.3	12.8	15.9	6.9
27 Jul-09 Aug	174.12	11.87	185.99	127.5	14.8	22.4	9.5
09 Aug-22 Aug	137.52	5.62	143.14	87.0	15.3	19.8	8.0
22 Aug-04 Sep	107.56	2.1	109.66	68.7	14.3	12.5	7.0
04 Sep-17 Sep	90.66	2.29	92.95	55.1	10.9	14.4	6.0
17 Sep-30 Sep	119.13	4.21	123.34	73.2	15.8	17.2	7.9
30 Sep-13 Oct	61.73	4.63	66.36	39.9	7.7	7.6	4.6
13 Oct-26 Oct	36.15	2.66	38.81	21.0	4.8	4.3	2.5

CAST 02 Deep

Period	<1 mm	>1 mm	Total	Carbonate	Opal	Lithogenic	C.org
19 Nov-01 Dec	82.23	8.56	90.79	57.2	7.7	8.8	5.9
01 Dec-14 Dec	71.41	2.32	73.73	49.7	69.3	6.8	4.3
14 Dec-26 Dec	31.23	20.16	51.39	20.3	5.1	1.7	2.4
26 Dec-08 Jan	113.59	4.25	117.84	78.9	13.2	9.0	6.2
08 Jan-21 Jan	118.07	15.22	133.29	83.2	13.1	8.2	7.8
21 Jan-02 Feb	104.34	9.48	113.82	76.1	10.0	12.4	7.9
02 Feb-15 Feb	84.8	14.68	99.48	52.6	12.0	8.3	6.7
15 Feb-27 Feb	111.45	66.38	177.83	64.9	22.3	17.3	9.0
27 Feb-12 Mar	58.36	20.63	78.99	12.0	0.0	0.0	1.9
12 Mar-25 Mar		0.44	0.44	0.0	0.0	0.0	0.0
25 Mar-06 Apr		0.43	0.43	0.0	0.0	0.0	0.0
06 Apr-19 Apr		0.34	0.34	0.0	0.0	0.0	0.0
19 Apr-01 May		0.73	0.73	0.0	0.0	0.0	0.0

Table 4.3 continued Total fluxes to the central Arabian Sea, shallow traps. All fluxes are in mg m^{-2} day^{-1}. Analysis of component fluxes could not be carried out due to nonavailability of samples.

CAST 01 Shallow

Period	<1 mm	>1 mm	Total
10 May-23 May	53.24	2.64	55.88
23 May-05 Jun	47.36	7.59	54.95
05 Jun-18 Jun	33.53	4.73	38.26
18 Jun-01 Jul	180.28	30.88	211.16
01 Jul-14 Jul	166.38	2.66	169.04
14 Jul-27 Jul	272.54	40.49	313.03
27 Jul-09 Aug	217.49	24.38	241.87
09 Aug-22 Aug	89.23	11.89	101.12
22 Aug-04 Sep	116.02	9.99	126.01
04 Sep-17 Sep	144.31	18.61	162.92
17 Sep-30 Sep	83.68	8.54	92.22
30 Sep-13 Oct			
13 Oct-26 Oct	88.43	6.14	94.57

CAST 02 Shallow

Period	<1 mm	>1 mm	Total
19 Nov-01 Dec	83.41	16.84	100.25
01 Dec-14 Dec	68.14	5.85	73.99
14 Dec-26 Dec	101.98	35.47	137.45
26 Dec-08 Jan	91.03	12.88	103.91
08 Jan-21 Jan	122.47	57.44	179.91
21 Jan-02 Feb	115.78	14.29	130.07
02 Feb-15 Feb	136.74	31.21	167.95
15 Feb-27 Feb	184.05	21.08	205.13
27 Feb-12 Mar	0.71		0.71
12 Mar-25 Mar	148.31	42.99	191.3
25 Mar-06 Apr	1.09		1.09
06 Apr-19 Apr	0.07		0.07
19 Apr-01 May	1.46		1.46

Table 4.4 Total and component fluxes to the western Arabian Sea. All fluxes are in mg m-2 day-1

WAST 01 Deep

Period	<1 mm	>1 mm	Total	Carbonate	Opal	Lithogenic	C.org
10 May-23 May	58.24	5.58	63.82	30.8	9.4	8.3	4.3
23 May-05 Jun	37.31	4.88	42.19	25.2	0.0	0.0	1.8
05 Jun-18 Jun	60.46	2.61	63.07	36.2	7.1	9.2	4.5
18 Jun-01 Jul	83.53	3.04	86.57	45.5	8.1	19.3	5.2
01 Jul-14 Jul	162.56	5.3	167.86	101.7	19.4	22.5	8.8
14 Jul-27 Jul	186.45	9.0	195.45	125.4	31.8	11.7	8.8
27 Jul-09 Aug	600.82	158.45	759.27	304.2	154.7	51.5	36.2
09 Aug-22 Aug	525.4	108.91	634.31	233.2	187.0	33.8	28.7
22 Aug-04 Sep	149.76	9.21	158.97	64.4	59.1	10.0	7.9
04 Sep-17 Sep	28.06	1.43	29.49	10.2	0.0	0.0	1.4
17 Sep-30 Sep	15.11	2.11	17.22	10.1	0.1	3.7	0.5
30 Sep-13 Oct	15.11	2.11	17.22	10.1	0.1	3.7	0.5
13 Oct-26 Oct	15.11	2.11	17.22	10.1	0.1	3.7	0.5

WAST 02 Deep

Period	<1 mm	>1 mm	Total	Carbonate	Opal	Lithogenic	C.org
19 Nov-01 Dec	90.28	6.28	96.56	67.0	11.2	6.1	5.8
01 Dec-14 Dec	109.68	7.33	117.01	81.1	12.6	8.8	6.7
14 Dec-26 Dec	114.04	7.34	121.38	81.6	14.7	4.8	6.2
26 Dec-08 Jan	56.82	5.14	61.96	51.5	1.8	0.0	2.5
08 Jan-21 Jan	63.24	12.26	75.5	48.2	6.5	2.6	3.8
21 Jan-02 Feb	99.56	23.51	123.07	60.9	19.7	4.4	7.2
02 Feb-15 Feb	71.79	39.07	110.86	36.5	19.3	0.0	7.5
15 Feb-27 Feb	14.59	4.79	19.38	7.1	0.0	0.0	1.7
27 Feb-12 Mar	14.59	4.79	19.38	7.1	0.0	0.0	1.7
12 Mar-25 Mar	14.59	4.79	19.38	7.1	0.0	0.0	1.7
25 Mar-06 Apr	14.59	4.79	19.38	7.1	0.0	0.0	1.7
06 Apr-19 Apr	14.59	4.79	19.38	7.1	0.0	0.0	1.7
19 Apr-01 May	14.59	4.79	19.38	7.1	0.0	0.0	1.7

coastal upwelling areas. It is seen that the bulk of the particles are composed of fresh biogenic material derived from the surface layers of the ocean (Table 4.1).

In the central Arabian Sea particle flux pattern in the shallow traps were similar to the deep traps. The peaks in the shallow trap during the SW monsoon period were offset by one cup interval (Fig 4.3). This shows that particles produced at the surface sink rapidly to the sea floor. Since the depth difference between the shallow and deep trap is 2182 meters and the particles takes 13 days to settle through this distance the sinking speed of particles can be computed approximately as 165 meters per day. That means that particles sinking out of the photic zone should reach the deep ocean floor in 3 to 4 weeks.

Total flux pattern at all three sites showed strong similarity with highest flux during the southwest monsoon and very low flux during the intermonsoon period. The bimodal pattern in particle fluxes (Fig. 4.2) show that the Arabian Sea alternates between oligotrophic conditions during the inter-monsoon periods to highly eutrophic conditions during the monsoons. Chlorophyll concentrations and primary productivity in the surface waters also show a similar pattern (*Banse, 1987, Yoder et al, 1991*). This similarity in flux patterns indicates that a common factor governs particle fluxes in the entire Arabian Sea.

There is close similarity between wind speeds and particle flux patterns at all the three sites in the Arabian Sea (Fig 4.2). Wind speeds over the three sites were obtained from ship observations within a radius of 2 degrees of the trap location as reported in the Indian Daily Weather Report and smoothed with a five point moving average. It is known that wind stress causes upwelling of nutrient rich deep waters which stimulate primary productivity in the photic

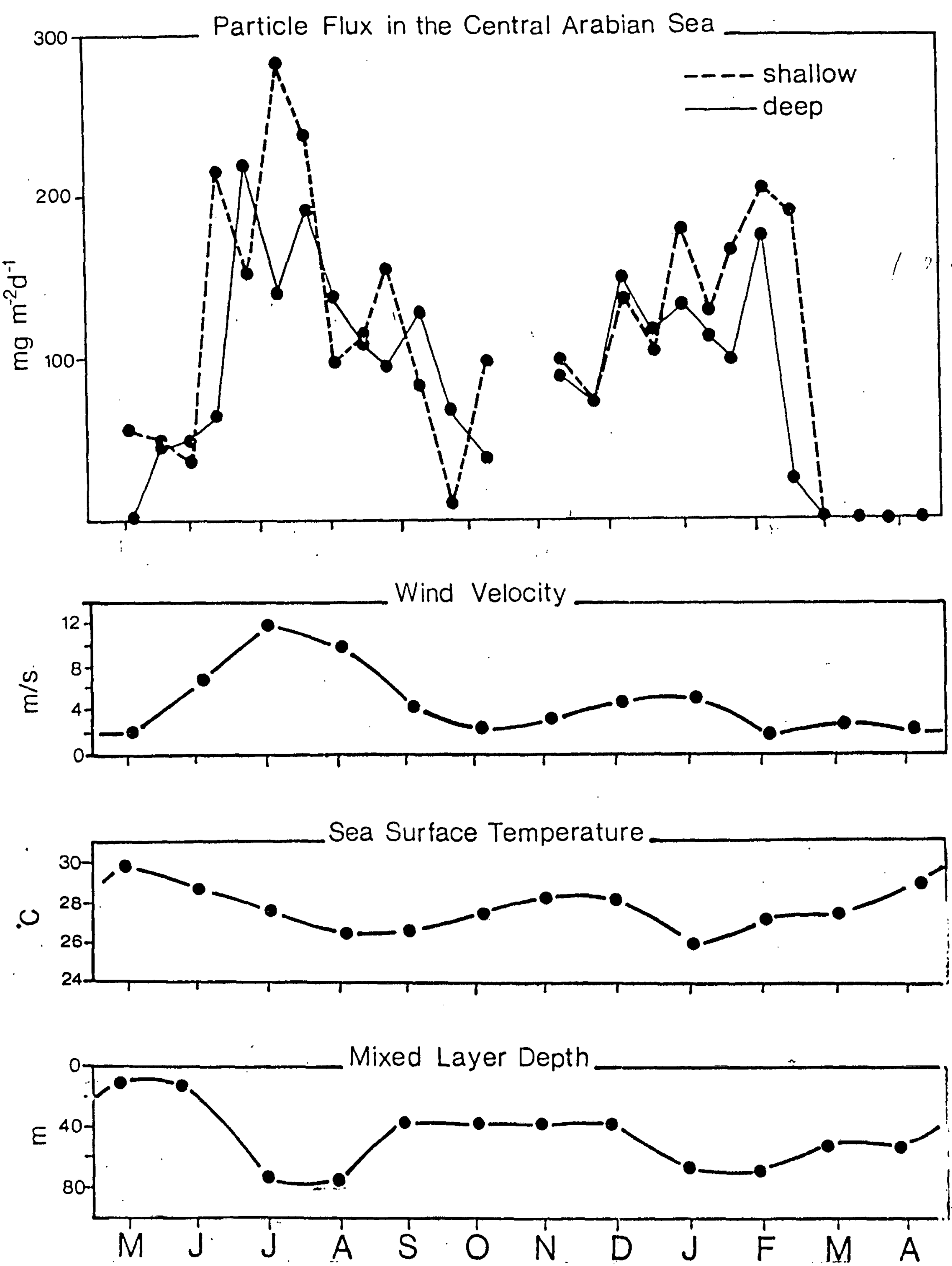


Fig. 4.3 Particle flux pattern in the shallow and deep traps in the central Arabian Sea between May 1986 and April 1987. Also shown are historical data on wind velocity, sea surface temperature and mixed layer depth in the open Arabian Sea.

zone. Upwelling in the Oman upwelling center in the western Arabian Sea has been well documented for many years. The monsoon winds produce a low-level jet, called the Finlater Jet, which causes open ocean upwelling near the Oman coast. Nutrient rich upwelled waters from the Oman upwelling area can be carried offshore and eastwards by vigorous surface currents and advected over the traps (*Elliot and Savidge, 1990; Swallow, 1984*) to contribute part of the observed high particle fluxes in the western Arabian Sea. It is also possible that particles produced in the nearby upwelling centers are horizontally advected at depth.

In the eastern Arabian Sea, highest fluxes occur during August and September. The dissipating phase of the summer monsoon was noticed off the coast of Goa during September 1986 (*Madhupratap et al., 1990*) and part of the observed high fluxes here may have been derived from the coastal upwelling areas.

A different process operates in the central part of the Arabian Sea. Historical data on oceanographic parameters over the Arabian Sea show that during high wind speed periods the sea surface temperature (SST) is reduced and the mixed layer deepens from about 30 to 40 m during the premonsoon months to over 100 m during the southwest monsoon period (Fig. 4.3) (*Hasenrath and Lamb, 1979, Bauer et al, 1991*). The surface waters are enriched in nutrients primarily due to mixing with nutrient rich sub-surface waters as well as turbulent entrainment from the nutracline. As a consequence, chlorophyll concentrations in the western Arabian Sea; which are less than 0.2 microgram l^{-1} during May increase to over 2 microgram l^{-1} between August and October (*Yentsch et al., 1992*). A deep chlorophyll maximum around 80 to 100 m depth is formed due to a combination of Ekman pumping and advection of dense fluids from the north (*Bauer et al,* *1991*). Shallow mixed layers are found throughout the Arabian Sea between September to November which again deepen by December (*Bauer et al, 1991*) leading to high particle fluxes during this period. During the NE monsoon, winter blooms of phytoplankton are found in the northern and central parts of the Arabian Sea (*Banse and McClain, 1986*).

High primary production during both the SW and NE monsoons is reflected in high particle fluxes to the deep sea (Fig. 4.3 and 5.16). Wind induced nutrient pumping into the euphotic zone is especially effective in the Arabian Sea as the sub-surface waters here have high nutrient concentrations (*Ryther and Menzel, 1965*).

During the winter monsoon, low particle fluxes are observed in the eastern Arabian Sea while the western and central sites have moderate to high fluxes (Fig. 4.2). Low particle fluxes in the eastern trap could be due to stronger stratification of the upper layers of the water column as a result of influx of low saline waters from the Bay of Bengal or eastern tropical Indian Ocean. Stronger stratification prevents deepening of the mixed layer and leads to lowering of primary productivity due to nutrient exhaustion.

Effects on eolian dust on particle fluxes: The monsoon winds introduce 100 to 200 million tonnes of eolian dust into the Arabian Sea every year (Sirocko, 1989). Along with eolian dust significant amount of iron is also supplied to the open ocean. Iron is an important trace nutrient and is thought to limit productivity in high productive areas like the Southern Oceans and upwelling centers where excess nitrogen is available (*Martin and Fitzwater, 1988; Martin et al, 1990*). The monsoon winds also bring in more than 100 million tonnes of eolian dust into the Arabian Sea every year (*Sirocko and Sarnthein, 1989*) which increases the availability of iron in the photic zone. The simultaneous introduction of nutrients and soluble trace metals into the surface waters of the Arabian Sea may produce a more intense bloom than would otherwise occur. The response of the biological pump to the increased wind speeds is extremely fast as the increase in primary productivity and particle fluxes begins almost simultaneously with the onset of the monsoons (Fig. 4.2, 4.3).

Component fluxes: Table 4.1 to 4.4 show the component fluxes to the deeper traps in the Arabian Sea. Biogenic components form between 77 to 92 per cent of the total annual fluxes the remainder being composed of terrigenous material derived from land. The major components in the sediment trap material are carbonates, opal, lithogenic material and organic matter.

Carbonate fluxes and carbonate opal ratios: Carbonates are the dominant components in the sediment traps at all three sites in the Arabian Sea with annual fluxes ranging between 12 to 19 g m^{-2} y^{-1}. Very high carbonate fluxes, over 300 mg m^{-2} y^{-1}, are seen during the SW period in the western Arabian Sea. Microscope observations show that in this area carbonates are mainly contributed by foraminifers and coccoliths with minor contributions by pteropods. During the southwest monsoon up to 15,000 planktonic foraminifers per square meter per day were captured by the sediment traps (*Curry et al, 1992*).

Carbonate opal ratios indicate the prevalent planktonic community structure which in turn is dependent on nutrient availability. The high values in the deep traps of the Arabian Sea (Table 4.1) point to the high carbonate productivity of the area. Here, carbonate opal ratios are high during the NE monsoon and intermonsoon period but low during the SW monsoon period (Fig. 4.5). It is interesting to note that in the western Arabian Sea, during the earlier part of the SW monsoon period when total particle fluxes are moderately high (between 50 to 200 mg m^{-2} y^{-1}) carbonate/ opal ratios increase, (Fig. 4.6) but during the later part of the monsoon when particle fluxes exceed 200 mg m^{-2} y^{-1} there is a fall in the ratio. Foraminifera fluxes are high during the later part of the monsoon (*Curry et al., 1992*) showing that the high carbonate fluxes in the early part of the monsoon are due to coccoliths. Probably, the coccolith production is high during the early SW monsoon period and the coccolith bloom is replaced by a diatom bloom during the later part of the monsoon when silica concentrations in the surface waters increase (*Haake et al., 1993*. Carbonate percentages and coccolith numbers in the sediments below upwelling centers in the western Arabian Sea are extremely high (*Thiede, 1974*). In fact the upwelling sediments of the Arabian Sea have more carbonate and less opal than any other upwelling zone.

Organic matter: The seasonality of organic carbon fluxes is very similar to total particulate matter fluxes. Organic carbon flux range from less than 1 mg m^{-2} day^{-1} during the inter-monsoon period to over 28 mg m^{-2} day^{-1} during the SW monsoon. Organic carbon percentages range between 4 and 19% of the total fluxes in the shallow traps and between 4 and 17% in the deep traps. Highest fluxes are in the western Arabian Sea and decrease towards the east. The annual fluxes are between 1.53 and 1.8 g C m^{-2} y^{-1}. Organic carbon fluxes decrease with depth and between 30 and 40 per cent of the organic matter is decomposed between the shallow and deep traps (*Haake et al, 1992*). Most of the C.org fluxes are pelagic with minor terrestrial contribution (*Reemtsma et al., 1990; Haake et al., 1992*).

Opal: Up to 40% opal is found in the western Arabian Sea during the SW monsoon period which is more than twice that of the central and eastern traps (*Nair et al, 1989*). In the western Arabian Sea opal contributed more than 21% of the total annual fluxes and during the SW monsoon period their contribution exceeded 40% (*Ramaswamy et al., 1991*). The main contributor of opal are diatoms, radiolarians and silicoflagellates. Microscopic observations show that the *rhizosolenia* a major upwelling species is the dominant diatom in the western Arabian Sea traps (*Nair et al., 1989*) The main diatom and radiolarian bloom is present towards the later part of the monsoon during the end of July and in August (*Haake et al., 1993*).

Lithogenics: Highest lithogenic fluxes are in the eastern Arabian Sea during the southwest and post monsoon period. There is a east west gradient in lithogenic fluxes with annual fluxes decreasing from 5.4 g m^{-2} in the eastern Arabian Sea to 2.64 g m^{-2} in the western Arabian Sea (Table 4.1). The main source of lithogenic material are the rivers, Indus, Narmada and Tapti and eolian input from the Somali-Arab deserts (*Goldberg and Griffin, 1970, Weser, 1974, Kolla et al, 1981, Sirocko, 1989*). Factors controlling lithogenic fluxes are discussed in detail in chapter 5 and The mineralogy and grain size of the lithogenic fraction is discussed in detail in chapters 6 and 7.

4.3.2 Total and component particle fluxes in the Bay of Bengal

Total fluxes: During the period November 1988 to November 1989 the total annual fluxes to the deep traps of the Bay of Bengal range between 42 and 54 g (Table 4.5 to 4.8) and fluxes in both the northern and central Bay of Bengal traps were similar in quantity as well as composition even though the northern trap is located about 450 km closer to the mouths of the Ganges-Brahmaputra river.

In both the areas, strong seasonality in particle fluxes were observed, with the difference between maximum and minimum fluxes varying by an order of magnitude (Fig 4.4 and figures 5.5 to 5.9). The southwest monsoon peak in particle fluxes is very prominent and fluxes during this period account for 40 to 50% of the annual fluxes (Table 4.5). During the rest of the year fluxes

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Table 4.5 Annual and seasonal fluxes of particulate matter and first order components to the northern (NBBT) and Central (CBBT) Bay of Bengal for the period November 1988 to November 1989. All fluxes are in g m-2

	Total Flux	Carbonate	Opal	OM	Lith	Carb/ Opal	C.org/ C.Carb
NBBT 03 Shallow							
SW	11.9	4.8	1.5	1.7	3.5	3.2	1.7
SW-NE	4.6	1.5	0.8	0.7	1.5	1.9	2.3
NE	13.2	5.7	2.8	2	1.9	2.0	1.7
NE-SW	10.7	3.7	1.5	2.5	2.5	2.5	3.3
Annual	44.1	16.8	7	7.6	10.2	2.4	2.2
NBBT 03 Middle							
SW	17.3	6.5	1.8	3	5.1	3.6	2.2
SW-NE	4.9	1.6	0.9	0.6	1.5	1.8	1.8
NE	13.9	5.9	2.7	1.8	2.3	2.2	1.5
NE-SW	9.4	3	1.4	1.6	2.7	2.1	2.6
Annual	48.6	18.1	7.3	7.5	12.6	2.5	2.0
NBBT 03 Deep							
SW	16.5	5.8	2.3	2.3	5.9	2.5	1.9
SW-NE	5.3	1.5	1.1	0.6	2	1.4	1.9
NE	14.2	5.8	3.5	1.7	2.8	1.7	1.4
NE-SW	9.5	3.5	1.7	1.2	2.8	2.1	1.7
Annual	48.7	17.8	9.1	6.2	14.4	2.0	1.7
CBBT 03 Shallow							
SW	17.1	6.1	3.2	2.6	4.9	1.9	2.1
SW-NE	6.9	2.5	1.1	0.9	2.2	2.3	1.7
NE	10.5	4.7	2.3	1.5	1.8	2.0	1.5
NE-SW	5.5	2.3	1.1	1	1	2.1	2.1
Annual	41.8	16.4	8.1	6.2	10.1	2.0	1.8
CBBT 03 Deep							
SW	24	8.2	3.8	2.9	8.6	2.2	1.7
SW-NE	3.1	1.3	0.4	0.3	1	3.3	1.1
NE	14.3	5.4	3.1	1.7	3.8	1.7	1.5
NE-SW	6.2	2.8	1.1	0.8	1.4	2.5	1.4
Annual	53.8	20.1	9.6	6.4	16.6	2.1	1.5

OM = organic matter
Lith. = Lithogenic
C.org. = Organic carbon
C.Carb. = Carbonate carbon

Table 4.6 Total and component fluxes to the northern Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

North Bay of Bengal- 01 Shallow

Period	<------------FLUX in mg m-2 day-1-----> <1mm	Carbonate	Opal	Lithogenic	C.org.
28 Oct-06 Nov 1987	93.45	26.72	18.21	26.55	6.11
06 Nov-16 Nov	101.25	29.92	20.94	32.20	4.95
16 Nov-25 Nov	119.24	31.85	28.22	41.46	7.35
25 Nov-05 Dec	115.94	26.85	35.92	35.81	8.57
05 Dec-14 Dec	136.41	52.95	18.21	37.47	10.76
14 Dec-24 Dec	142.60	63.62	20.93	26.87	8.51
24 Dec-02 Jan 1988	77.26	29.51	8.19	16.11	5.02
02 Jan-12 Jan	123.86	47.83	20.93	20.42	10.89
12 Jan-21 Jan	123.70	37.56	18.21	42.38	8.40
21 Jan-31 Jan	125.75	27.60	36.42	44.60	7.63
31 Jan-09 Feb	164.67	54.93	47.31	38.22	13.04
09 Feb-19 Feb	92.61	24.97	22.14	29.73	6.85
19 Feb-28 Feb 1988	52.94	10.88	13.09	16.33	2.57

North Bay of Bengal- 03 Shallow

Period	<------------FLUX in mg m-2 day-1--------> <1mm	Carbonate	Opal	Lithogenic	C.org
02 Nov-29 Nov 1 1988	82.57	24.98	15.32	27.91	6.0
29 Nov-26 Dec 2	109.77	47.36	27.70	10.68	8.1
26 Dec-22 Jan 3 1989	148.37	74.18	31.29	10.06	12.8
22 Jan-18 Feb 4	109.33	48.02	21.83	14.78	10.0
18 Feb-17 Mar 5	121.80	40.69	21.83	33.21	10.5
17 Mar-13 Apr 6	190.88	53.97	23.79	47.03	32.1
13 Apr-10 May 7	188.57	63.96	28.68	47.71	20.3
10 May-06 Jun 8	143.65	58.09	20.86	30.72	15.4
06 Jun-03 Jul 9	151.98	59.68	12.06	51.77	12.6
03 Jul-30 Jul10	98.52	39.41	12.06	27.73	8.3
30 Jul-26 Aug11	121.63	49.59	17.92	33.11	8.7
26 Aug-22 Sep12	69.70	26.15	12.38	15.82	5.9
22 Sep-19 Oct13 1989	88.99	29.31	12.50	28.65	7.5

Table 4.6 continued. Total and component fluxes to the northern Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

North Bay of Bengal 03 Middle

Period	<1mm	Carbonate	Opal	Lithogenic	C.org
	<------------FLUX in mg m-2 day-1-------->				
02 Nov-29 Nov 1 1988	103.68	33.30	17.98	34.75	5.8
29 Nov-26 Dec 2	152.84	48.97	38.32	27.01	11.5
26 Dec-22 Jan 3 1989	139.87	75.66	23.30	13.59	9.5
22 Jan-18 Feb 4	97.01	47.52	15.76	14.06	6.9
18 Feb-17 Mar 5	124.77	46.22	21.15	32.08	9.2
17 Mar-13 Apr 6	152.85	42.76	22.41	53.04	14.1
13 Apr-10 May 7	169.23	53.39	25.35	47.95	16.9
10 May-06 Jun 8	142.61	54.25	21.73	34.31	12.7
06 Jun-03 Jul 9	279.20	112.48	28.98	78.06	35.1
03 Jul-30 Jul10	93.21	39.05	9.65	25.84	7.5
30 Jul-26 Aug11	181.92	64.77	25.69	55.56	13.6
26 Aug-22 Sep12	84.76	26.19	15.64	29.65	6.0
22 Sep-19 Oct13 1989	77.72	27.41	15.19	22.38	5.6

Table 4.6 continued. Total and component fluxes to the northern Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

North Bay of Bengal-01 Deep

	FLUX in mg m-2 day-1				
Period	<1mm	Carbonate	Opal	Lithogenic	C.org.
28 Oct-06 Nov 1987	74.85	14.43	11.58	38.18	3.34
06 Nov-16 Nov	88.76	24.61	16.11	30.03	4.34
16 Nov-25 Nov	124.46	44.34	21.14	46.90	5.82
25 Nov-05 Dec	109.87	28.04	24.16	43.18	5.78
05 Dec-14 Dec	106.47	32.31	20.64	35.78	5.67
14 Dec-24 Dec	165.84	64.27	28.19	49.67	9.97
24 Dec-02 Jan 1988	103.37	31.89	19.12	35.64	5.47
02 Jan-12 Jan	109.57	42.45	16.61	31.87	6.26
12 Jan-21 Jan	81.75	24.16	15.10	24.52	4.90
21 Jan-31 Jan	73.32	17.85	14.60	27.86	3.97
31 Jan-09 Feb	120.99	39.99	27.79	39.73	6.30
09 Feb-19 Feb	90.43	20.66	23.15	32.70	5.03
19 Feb-28 Feb 1988	110.76	23.85	19.63	56.40	5.01

North Bay of Bengal-03 Deep

	FLUX in mg m-2 day-1				
Period	<1mm	Carbonate	Opal	Lithogenic	C.org
02 Nov-29 Nov 1 1988	101.59	21.46	23.90	40.64	6.6
29 Nov-26 Dec 2	133.01	46.22	37.31	29.04	8.8
26 Dec-22 Jan 3 1989	169.02	76.35	42.89	24.16	11.2
22 Jan-18 Feb 4	127.32	58.27	26.81	22.99	8.5
18 Feb-17 Mar 5	96.69	33.98	21.67	26.76	6.3
17 Mar-13 Apr 6	142.57	46.78	26.58	47.67	10.2
13 Apr-10 May 7	168.50	65.22	29.58	48.59	12.0
10 May-06 Jun 8	157.76	62.46	26.69	43.23	11.9
06 Jun-03 Jul 9	192.61	66.87	22.24	75.45	13.5
03 Jul-30 Jul10	151.12	58.13	18.46	53.48	10.0
30 Jul-26 Aug11	165.36	56.48	25.35	50.23	16.4
26 Aug-22 Sep12	103.68	32.49	18.22	38.41	6.6
22 Sep-19 Oct13 1989	95.96	34.32	16.46	32.93	5.8

Table 4.7 Total and component fluxes to the central Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

Central Bay of Bengal- 01 Shallow

Period		<------------FLUX in mg m-2 day-1----->				
		<1mm	Carbonate	Opal	Lithogenic	C.org.
28 Oct-06 Nov	1987	19.51	11.19	1.54	3.54	0.00
06 Nov-16 Nov		131.99	56.98	12.35	42.05	0.00
16 Nov-25 Nov		115.27	52.47	15.00	34.80	7.88
25 Nov-05 Dec		100.12	43.47	15.00	29.48	7.87
05 Dec-14 Dec		92.61	44.10	21.17	14.96	6.58
14 Dec-24 Dec		84.55	37.87	20.74	13.83	6.12
24 Dec-02 Jan	1988	86.79	40.10	20.31	15.18	6.49
02 Jan-12 Jan		117.10	70.84	23.24	11.54	6.87
12 Jan-21 Jan		23.02				0.00
21 Jan-31 Jan		6.61				0.00
31 Jan-09 Feb		297.02	220.38	34.75	18.42	14.11
09 Feb-19 Feb		105.23	63.88	17.64	10.97	8.86
19 Feb-28 Feb	1988	48.09	20.05	13.59	7.38	3.79

Central Bay of Bengal- 03 Shallow

Period		<------------FLUX in mg m-2 day-1--------->				
		<1mm	Carbonate	Opal	Lithogenic	C.org
02 Nov-29 Nov 1	1988	133.14	53.79	21.80	35.46	10.1
29 Nov-26 Dec 2		76.35	34.16	14.90	14.54	6.4
26 Dec-22 Jan 3	1989	144.51	49.14	39.14	29.87	12.0
22 Jan-18 Feb 4		98.56	62.90	14.33	10.11	5.6
18 Feb-17 Mar 5		69.98	27.43	16.46	12.18	6.1
17 Mar-13 Apr 6		72.56	33.70	14.99	10.91	6.1
13 Apr-10 May 7		92.05	37.90	21.51	12.42	10.6
10 May-06 Jun 8		106.76	44.41	16.83	24.66	10.3
06 Jun-03 Jul 9		230.96	75.64	35.32	79.77	20.0
03 Jul-30 Jul10		168.34	53.63	41.43	44.31	13.8
30 Jul-26 Aug11		156.22	69.33	29.44	31.45	12.8
26 Aug-22 Sep12		78.92	27.59	13.75	24.55	6.1
22 Sep-19 Oct13	1989	121.29	39.22	19.55	44.67	8.6

Table 4.7 continued. Total and component fluxes to the central Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

Central Bay of Bengal- 01 Deep

Period	<------------FLUX in mg m-2 day-1-----> <1mm	Carbonate	Opal	Lithogenic	C.org.
28 Oct-06 Nov 1987	106.40	45.99	15.44	36.55	5.34
06 Nov-16 Nov	153.18	77.23	18.52	45.02	7.80
16 Nov-25 Nov	98.22	37.54	16.45	33.92	5.38
25 Nov-05 Dec	108.77	42.90	18.97	35.47	6.11
05 Dec-14 Dec	135.68	52.27	26.90	39.17	11.22
14 Dec-24 Dec	114.10	45.91	23.82	33.53	7.21
24 Dec-02 Jan 1988	221.68	113.90	49.37	39.04	12.02
02 Jan-12 Jan	34.18	149.46	116.95	39.46	21.65
12 Jan-21 Jan	112.08	46.32	25.50	27.14	6.52
21 Jan-31 Jan	150.88	83.65	28.48	25.79	7.71
31 Jan-09 Feb	136.61	64.67	30.31	24.15	8.85
09 Feb-19 Feb	149.45	63.33	32.57	33.81	10.88
19 Feb-28 Feb 1988	136.02	65.32	27.74	31.03	7.25

Central Bay of Bengal- 03 Deep

Period	<------------FLUX in mg m-2 day-1--------> <1mm	Carbonate	Opal	Lithogenic	C.org
02 Nov-29 Nov 1 1988	115.10	48.51	15.83	38.39	5.8
29 Nov-26 Dec 2	120.23	37.62	24.08	41.50	7.9
26 Dec-22 Jan 3 1989	161.91	65.09	39.20	37.00	9.4
22 Jan-18 Feb 4	147.32	63.86	28.76	33.72	9.9
18 Feb-17 Mar 5	100.36	32.67	22.37	28.16	7.6
17 Mar-13 Apr 6	91.87	40.58	18.11	20.90	5.6
13 Apr-10 May 7	84.07	41.00	16.19	15.95	5.3
10 May-06 Jun 8	129.08	55.39	21.09	33.00	9.8
06 Jun-03 Jul 9	188.43	73.19	26.63	58.32	14.9
03 Jul-30 Jul10	203.72	64.13	37.93	71.90	13.8
30 Jul-26 Aug11	362.38	125.75	53.27	134.16	22.6
26 Aug-22 Sep12	133.98	39.58	24.08	53.78	8.0
22 Sep-19 Oct13 1989					

Table 4.8 Total and component fluxes to the southern Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

Southern Bay of Bengal- 01 Shallow

Period		<1mm	Carbonate	Opal	Lithogenic	C.org.
		<------------FLUX in mg m-2 day-1----->				
28 Oct-06 Nov	1987	62.48	26.40	6.41	12.00	2.78
06 Nov-16 Nov		50.44	32.35	9.32	2.97	2.66
16 Nov-25 Nov		49.92	22.85	2.33	5.79	2.44
25 Nov-05 Dec		40.75	13.08	3.50	9.18	1.91
05 Dec-14 Dec		16.88	9.60	0.87	1.85	0.66
14 Dec-24 Dec		47.24	21.02	4.08	7.33	0.00
24 Dec-02 Jan	1988	54.53	19.78	6.41	11.04	3.24
02 Jan-12 Jan		40.15	14.71	3.50	7.75	1.75
12 Jan-21 Jan		39.26	21.12	4.08	3.90	2.91
21 Jan-31 Jan		54.98	26.45	5.24	8.11	
31 Jan-09 Feb		74.57	40.65	6.41	16.50	2.89
09 Feb-19 Feb		170.16	89.88	44.54	14.40	12.22
19 Feb-28 Feb	1988	19.69	nd	nd	nd	nd

Southern Bay of Bengal 01 Deep

Period		<1mm	Carbonate	Opal	Lithogenic	C.org.
		<------------FLUX in mg m-2 day-1----->				
28 Oct-06 Nov	1987	81.31	34.83	23.66	11.80	3.89
06 Nov-16 Nov		104.51	59.82	20.64	13.84	3.72
16 Nov-25 Nov		96.91	48.82	23.66	11.25	4.69
25 Nov-05 Dec		77.62	35.01	21.65	12.08	4.04
05 Dec-14 Dec		66.40	31.43	15.10	10.27	2.96
14 Dec-24 Dec		71.97	33.76	17.62	11.15	3.12
24 Dec-02 Jan	1988	64.39	29.40	15.10	10.47	2.74
02 Jan-12 Jan		59.30	25.87	12.08	10.68	2.21
12 Jan-21 Jan		81.76	31.76	13.59	24.81	2.90
21 Jan-31 Jan		79.96	41.69	14.09	12.78	nd
31 Jan-09 Feb		97.17	48.32	21.78	17.50	4.26
09 Feb-19 Feb		139.31	63.20	42.75	19.50	6.85
19 Feb-28 Feb	1988	143.81	62.26	44.59	19.82	10.35

were moderate or low. A maxima in particle fluxes during the SW and NE monsoon period were observed in all the traps. This seasonality was more pronounced in the central traps as over 40 percent of the fluxes at this location were during the SW monsoon period. The observed flux pattern coincides with the river discharge patterns of Himalayan as well as Peninsular Indian rivers (Fig. 4.4).

River discharge can affect the productivity of coastal seas and in the case of large rivers their influence should extend for a considerable distance out into the open sea. If river water has a higher nutrient content than sea water it will enhance the productivity. On the other hand if most of the nutrient is retained within the estuaries the productivity of the seas is decreased. Further, suspended solids present in the river plumes reduces the light intensity within the water column and at the same time remove the existing nutrients in the sea by adsorption processes.

The particle flux maxima in the Bay of Bengal coincided with the discharge maxima of the Ganges-Brahmaputra rivers. In the northern Bay of Bengal total flux during the SW monsoon showed two peaks. The first at the beginning of June related to meltwater influx and the second related to increased riverine input. Seasonally varying inputs of nutrients and lithogenic material together with changes in salinity of the surficial waters due to variations in river discharge seem to control particle fluxes in the Bay of Bengal. During the SW monsoon period, nutrient inputs from adjacent rivers together with turbulent mixing at the base of the halocline may be responsible for increase in the primary productivity of the Bay of Bengal. The nutrient concentration in rivers is enhanced due to agricultural land use in the Indian Sub-continent as well as deforestation in the Himalayas (*Ittekkot and Zhang, 1991*). Furthermore, settling rates of particles in the oceans will be

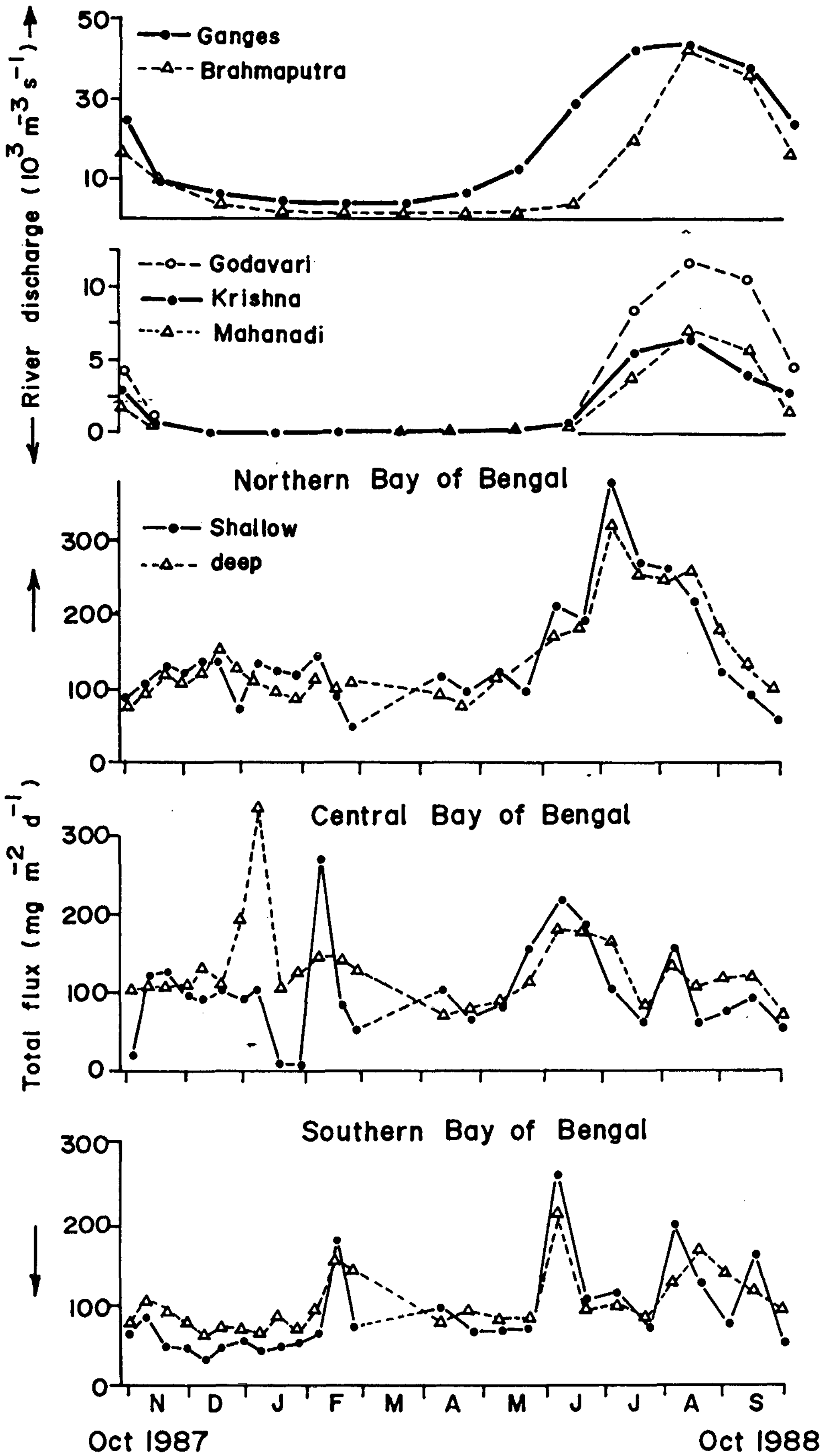


Fig. 4.4 Particle fluxes in the northern, central and southern Bay of Bengal during 1987-88 and freshwater river discharge pattern of major rivers in the region.

accelerated due to incorporation of lithogenic matter in organic aggregates (*Honjo, 1982; Deuser et al., 1983; Ramaswamy et al., 1991*).

Previous studies have shown that off the Amazon river, elevated concentrations of natural pigments were maintained for up to one year in the river plume while drifting for thousands of kilometers in the Atlantic (*Muller-Krager et al., 1989*). Also river plumes of the Amazon has been shown to increase the particle flux in traps located far away from the river mouth in the tropical Atlantic (*Deuser, 1989*). Elevated phytoplankton levels in the river plumes can be maintained for such long periods probably by either, recycling of nutrients within the plume (*Muller-Krager et al., 1989*) or by slow release of nutrients from land derived dissolved organic matter (*Deuser, 1989*).

Lateral transport of upwelled waters near the coast can provide nutrients and particles to the open ocean, as in the case of the Arabian Sea (*Nair et al., 1989*). However, in the Bay of Bengal only weak upwelling has been recorded and that too during the premonsoon period (*Qasim, 1977*). Strong monsoon winds do displace the surficial waters, but it is replaced by fresh low saline waters rather than nutrient rich deep waters (*Rao, 1977; Rao and Sastry, 1981; Murthy et al., 1990*). Further, the strong monsoon currents flowing parallel to the coast during the monsoons keep the upwelled waters close to the coast and does not affect our traps located more than 400 km from the coast. During the NE monsoon period the wind direction is not favorable for upwelling along the east coast of India. However, weak upwelling may occur along the Burmese coast.

During the SW monsoon, the winds are strongest and blow consistently from west to southwest with speeds ranging between 10 to 13 m s^{-1}. During winter, winds blow from the northeast with windspeed around 8 m s^{-1}. High wind stress does not play as important a role in deepening the mixed layer here because of stronger stratification of the surface waters. In fact the mixed layer becomes shallower during the SW monsoon with a low saline layer occupying the upper 25 m due to intense river discharge (*Rao, 1977*). However, wind stress together with winter cooling may play an important role in introduction of nutrients during winter when river discharge is low.

Apart from monsoon winds, tropical cyclones can also increase productivity by introducing nutrients from the subsurface layers. The cyclones also cause heavy rains on the adjacent landmass which leads to increased river discharge and introduction of more nutrients into the sea. The Bay of Bengal experiences a number of cyclones especially during the SW-NE intermonsoon period (*Indian Daily Weather Report, 1989, Pune*). During the period of study there were no tropical cyclones. However, two deep depressions formed during April and May in the northern Bay of Bengal (*IDWR and IWWR, 1989*). The latter depression lay centered about 2 ^{o} N of the northern mooring location. The high fluxes observed in the northern Bay of Bengal during May (Table 4.6 and Fig 5.5 to 5.7) is most probably the effect of these two depressions.

From satellite photographs a western boundary current has been identified along the east coast of India up to 15 ^{o}N during February, March and April (*Legeckis, 1989*). It then veers off to the right and flows into the central part of the Bay. No data is available to confirm the existence of this current during the monsoon due to intense cloud cover. Weak upwelling during the premonsoon periods which occur along the east coast of India is probably due to this boundary current. Material trapped during the premonsoon season may have been in part derived form the high productive regions along the continental margins and delivered to the trap sites by this boundary current.

Component fluxes: Lithogenic matter forms the main component of the particulate matter in the northern Bay of Bengal, while carbonate is the main component in other locations. Component fluxes in general follow total flux patterns with high fluxes during both the monsoons, except for lithogenic fluxes in the northern Bay of Bengal, which are comparatively low during the NE monsoon period (Table 4.6 to 4.8) (*Ramaswamy and Parthiban, 1990*). Variation in first order components here can be related primarily to variations in freshwater and suspended solids discharge by rivers

Carbonates: During 1988-89, carbonate percentage ranged between 25 to 64% with slightly higher fluxes in the central part of the Bay. Carbonate fluxes decreased slightly with depth, probably due to partial dissolution of the finer sized coccoliths as the depth of the lysocline here is about 2600 m (*Cullen, 1981*). Moreover, high C.org/C.carb ratios (Table 4.5) shows that excess organic carbon is available for dissolution of calcium carbonate (*Dymond, 1993*).

Carbonate percentages show a marked seasonal change in fluxes especially in the northern Bay of Bengal. Here peak fluxes of carbonates (>75 mg m^{-2} y^{-1}) are in January, during the NE monsoon period, which later decrease to less than 40 mg in March. During the premonsoon and beginning of the SW monsoon period carbonate fluxes are high, but drop drastically during the later part of the monsoon. These changes can be related to sharp changes in surface salinity of the Bay of Bengal waters which drops from 35 parts per thousand during the NE monsoon to less than 27 parts per thousand during the SW monsoon period (*Murthy et al., 1990*). Coccoliths can tolerate large variations in salinity between 16 to 45%. (*Haq, 1978*) and a new population may be established within a few days. Planktonic foraminifers on the other hand have limited tolerance to salinity changes and show change in species, size, ornamentation and chamber size for even small changes in salinity (as low as 1 to 2 %.) (*Boltovsky and Wright, 1976*). In the central Bay of Bengal, salinity changes are not so sharp, the minimum salinity recorded at this location being around 33 parts per thousand. Hence, carbonate and foraminifer flux do not decrease during this period. On the contrary, peak fluxes of carbonates are during July and August due to a general increase in primary productivity (*Qasim, 1977: Bhattathiri, 1980*).

Opal: Opal fluxes in the Bay of Bengal are high and annual fluxes in the northern Bay of Bengal are comparable to upwelling regions in the western Arabian Sea. Opal percentages generally range between 15 and 20 per cent. Microscope observation show that the main source of biogenous silica in the Bay of Bengal traps are diatoms with minor contributions by radiolarians and silicoflagellates. Compared to the Arabian Sea, radiolarian contribution is markedly less. Decrease in opal southwards in the Bay of Bengal reflects the plankton community structure in the surface waters. It is noticed that where nutrient concentrations are high opal producers dominate (*Dymond and Lyle, 1993*).

Organic matter: Organic matter percentages range between 10 to 15 % and is higher than generally reported from other sediment trap experiments in the open ocean at these depths. Organic carbon fluxes are much higher during the SW monsoon period than at other times as the primary productivity during this period is high. Another reason for high C.org fluxes is the high concentration of lithogenic particles which acts as a 'carrier particle' and accelerates the transport of organic matter to the sea floor.

Lithogenics: Higher lithogenic flux are associated with higher suspended solid discharge by rivers during the SW monsoons and their contribution to total flux averages around 30%. During the first year of deployment lithogenic percentages exceeded 50% in the deep northern Bay of Bengal traps (*Ittekkot et al., 1991*). Annual lithogenic fluxes recorded in the Bay of Bengal in the northern Bay of Bengal exceed 27 g m $^{-2}$. River discharge obviously accounts for a major part of the lithogenic flux as the traps are located less than 500 km away from the coast.

4.3.3 Comparison of particulate matter fluxes in the Arabian Sea and Bay of Bengal.

A comparison of total particulate matter fluxes in the Arabian Sea and Bay of Bengal show that annual fluxes in the Bay is higher by a factor of about 1.5 (Table 4.1 and 4.5). For comparison purpose data for the Bay of Bengal is taken from Ittekkot et al. (1991) as high resolution data is available for all three locations. The main component in the Arabian Sea is carbonates whereas lithogenic matter dominates in the Bay of Bengal. Higher fluxes in the Bay of Bengal is partly accounted for by high lithogenic fluxes which is supplied in considerable quantities by rivers. If we subtract the terrigenous fluxes and look only at the biogenic contribution it is seen that the western Arabian Sea and the central and southern Bay of Bengal have similar fluxes of about 28 g m^{-2} y^{-1} (Table 4.1 and 4.5). The other areas have lower biogenic fluxes ranging between 18 and 24 g m^{-2} y^{-1}. It is surprising to see that although primary productivity as well as biomass and chlorophyll content is much higher in the western Arabian Sea due to its proximity to the Oman upwelling center, total biogenic fluxes are similar to that of the oligotrophic waters of the southern Bay of Bengal located more than 600 km from any major river or upwelling center (*Babenerd and Krey, 1974; Qasim, 1977*). This shows that fluxes of biogenic components out of the photic zone is dependant not only on the production of these particles in the photic zone but also equally important on processes controlling their settling rate. In the Bay

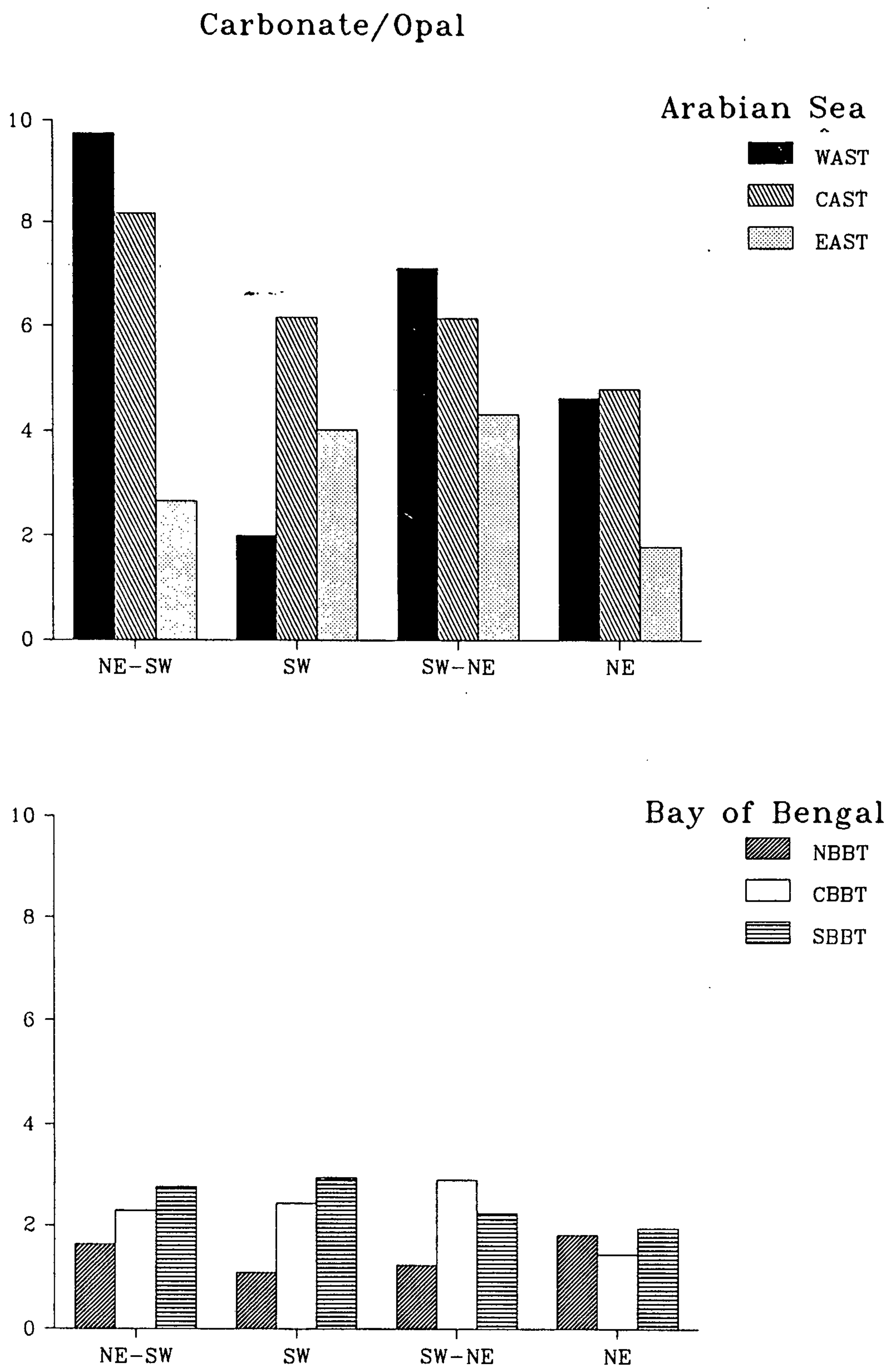


Fig. 4.5. Carbonate/Opal ratios of settling particles in the northern Indian Ocean. The Bay of Bengal data is taken from Ittekkot et al., 1991. Data is given for the SW (June - September) and NE (December-February) monsoons and for the periods in between.

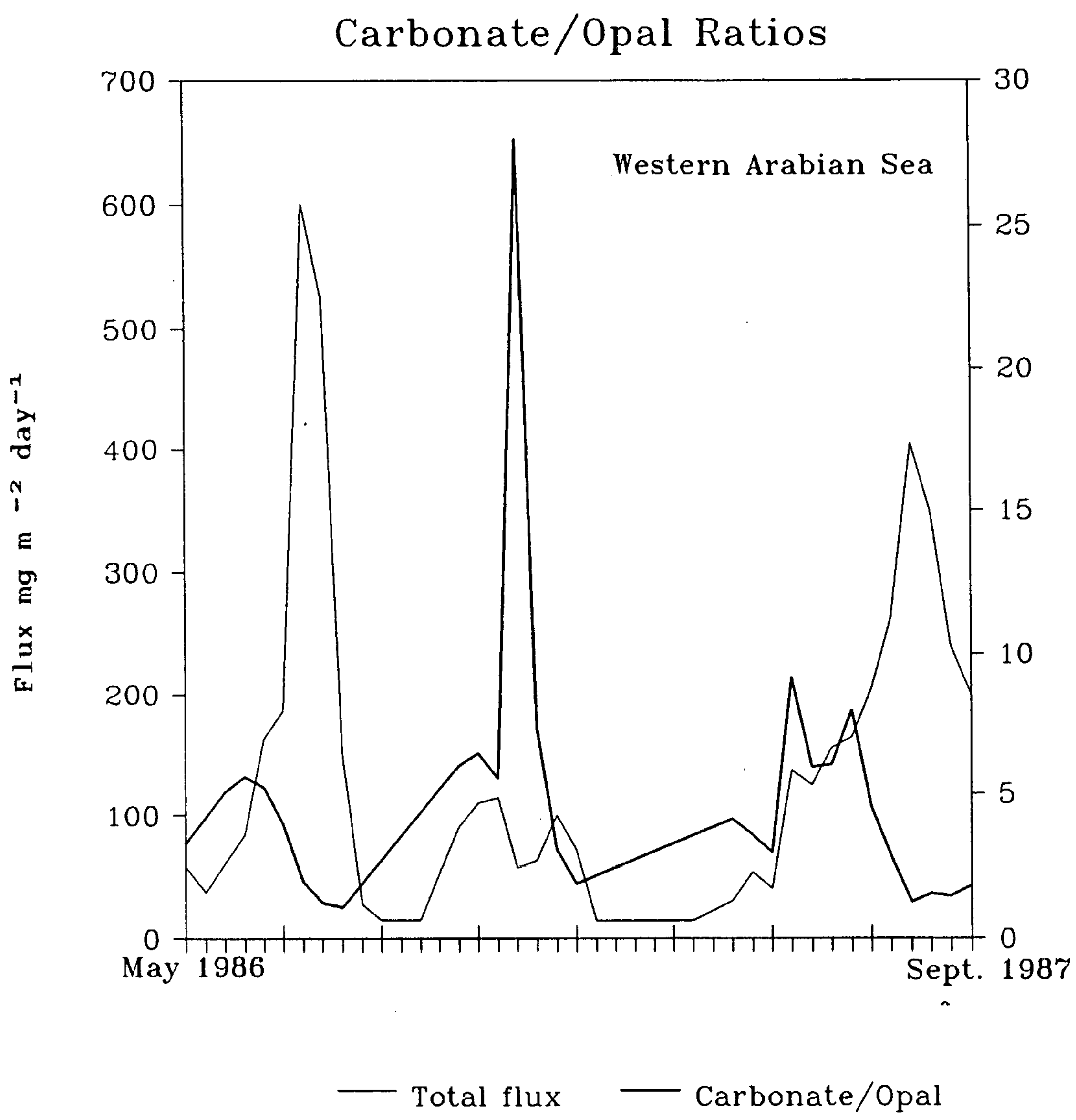


Fig 4.6. Carbonate/Opal ratios of settling particles in the deep western Arabian Sea trap for the period May 1986 to April 1987. Carbonate/Opal ratios is shown by a bold line. Total particle flux during the above period is shown by thin line.

of Bengal, lithogenic matter seems to play a major role in settling of biogenic particles.

The higher organic carbon fluxes to the Bay of Bengal in spite of lower productivity calls for an explanation. C/N ratios in the Bay of Bengal range between 7 and 8 and show that most of the organic matter is of marine origin (*Ittekkot et al, 1991, Reemtsma et al, 1992*). Apart from primary productivity, processes which transfer organic matter to the deep sea probably play a greater role in removal of organic carbon from the photic zone. Organic aggregates have low settling velocities and are rapidly remineralised in the upper layers of the ocean. Incorporation of lithogenic matter into organic aggregates increases their stability and density (*Ittekkot et al, 1992*). High density particles have higher settling rates (*Krank and Milligan, 1988*) and therefore have a better chance of being removed from the photic zone and exported to the deep sea. Higher organic carbon fluxes may be partly due to higher lithogenic fluxes in the Bay of Bengal.

Higher organic carbon fluxes may also be partly due to differences in food web structure. Compared to the Arabian Sea, zooplankton abundances are less in the Bay of Bengal (*Rao, 1973*), therefore recycling of organic matter in the upper mixed layer may be less efficient.

Berger and Kier (*1984*) have suggested that rain ratio of organic carbon/ carbonate carbon increases with increasing export production. They have shown that this shift in the rain ratio has great implication for CO_2 budget of the surface and deep ocean. Increase in organic carbon/ carbonate carbon ratios (Fig. 4.7) during high flux periods is seen in both the regions of the northern Indian Ocean. These ratios are considerably higher in the Bay of Bengal than in the Arabian Sea (fig. 4.7; Table 4.1 and 4.5). Whether this means that the Bay of Bengal is a bigger sink for atmospheric CO_2 remains

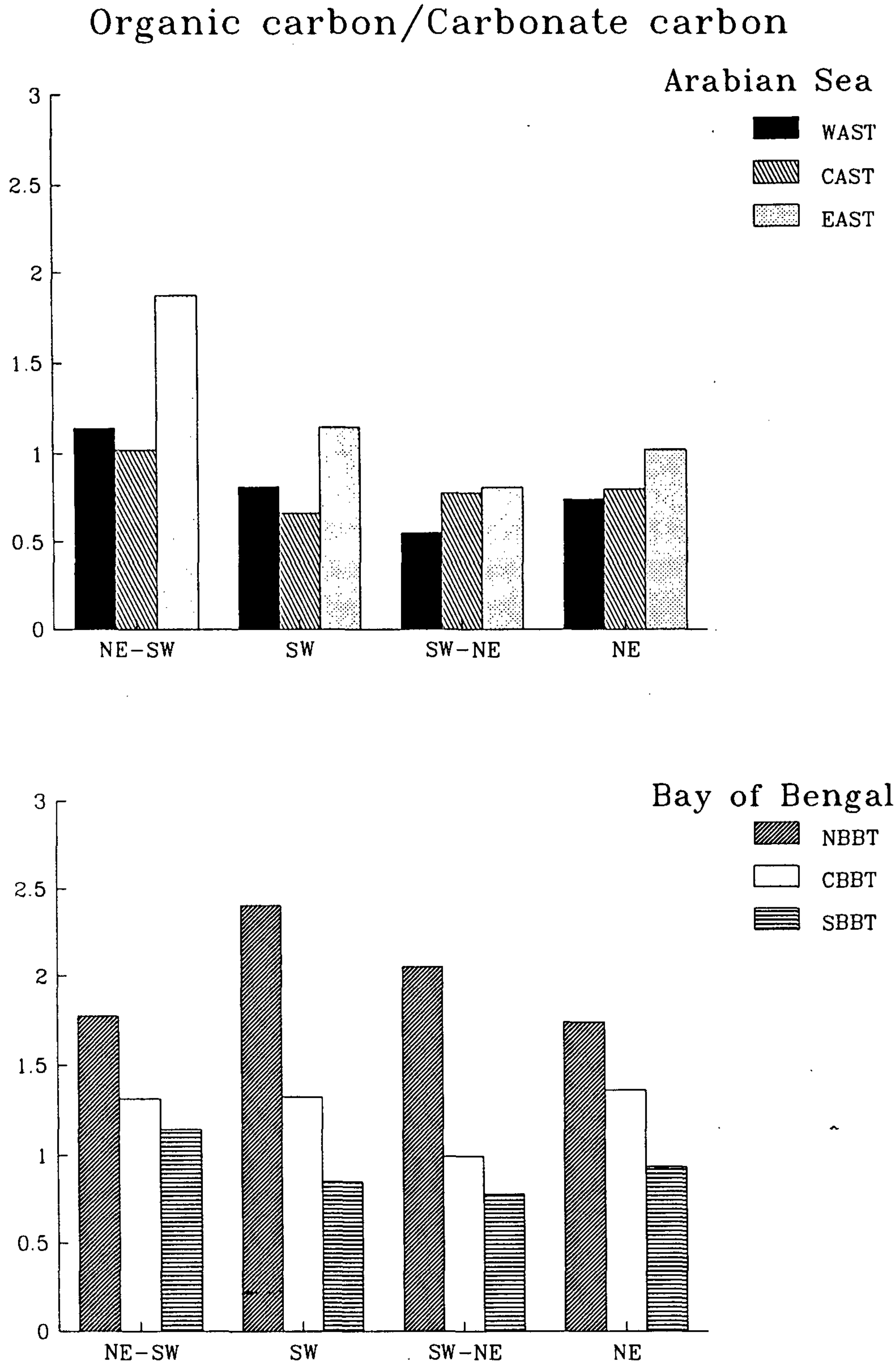


Fig 4.7. Organic carbon/Carbonate carbon ratios of particulate matter in the deep traps in the northern Indian Ocean. The Bay of Bengal data is from Ittekkot *et al.*, 1991. Data is given for the SW (June-September) and NE (December-February) monsoons and for the periods in between.

to be seen. High organic carbon/ carbonate carbon ratios in the Bay of Bengal is also due to higher organic matter settling along with high lithogenic fluxes and less efficient recycling of organic matter in the photic zone.

Opal concentrations in the sediments of the Arabian Sea are much higher than the Bay of Bengal (*Kolla and Kidd, 1982*). But contrary to expectations, annual opal fluxes are higher in the Bay of Bengal. Opal fluxes in the northern Bay of Bengal are slightly more than even the western Arabian Sea (Table 4.1 and 4.5). A major difference in opal flux pattern between the two regions is that in the Arabian Sea opal fluxes are episodic with 84 per cent of the opal flux during the SW monsoon period whereas in the Bay of Bengal moderate fluxes are maintained throughout the year. The number of colonial radiolarians, which have thicker tests compared to diatoms, are also more in the Arabian Sea (unpublished cruise reports of ORV Sagar Kanya and RV Sonne).

In summary, it is seen that seasonal variation in particle fluxes in the northen Indian ocean is controlled by the monsoon wind in the oceans and rain in adjacent continents. Monsoon winds stir up the water column and introduce nutrient rich waters into the photic zone due to upwelling or mixed layer deepening. this results in increased production and settling of biogenic matter. The monsoon winds also bring in eolian dust which is an inportant source of sediment in deep ocean areas. Higher rainfall in adjacent continents during the monsoons leads to increased discharge of suspended solids and nutrients by rivers into the oceans. This increases the production of biogenic material in the oceans leading to higher flux of particles to the deep ocean floor.

Chapter 5

Lithogenic Fluxes in the Arabian Sea and Bay of Bengal

5.1 Introduction

Lithogenic matter forms an important component of particulate matter in the deep sea. Most of it is derived from adjacent continents and introduced into the ocean as eolian dust or via rivers as fluvial deposits. Other sources of non-biogenic refractory particles in the ocean are from volcanogenic and cosmogenic sources, authigenic production at the sea floor, ice rafting, mass wasting turbidity currents etc. The various sedimentary processes responsible for transporting terrigenous particles from their point of introduction in the ocean to their final depositional sites is not very clearly understood. A study of these processes is important as most of the geological record is preserved in fine grained sediments of the deep sea.

Recent estimates of the total amount of lithogenic matter transported from land to ocean is about 21000 to 23000 million tonnes (Honjo, 1993). Of these about 18500 to 20000 million tonnes is supplied by rivers. Coastal erosion is estimated to supply about 500 million tonnes while ice rafting supplies another 1500 million tonnes of sediment and rocks to the oceans every year. The total estimate of eolian dust to the ocean is about 1600 million tonnes every year of which about 100 to 200 million tonnes is deposited in the Arabian Sea (*Sirocko, 1989*).

A major fraction of the suspended load of rivers is deposited on continental shelves and only a small portion of the fine grained sediments reach the deep sea directly. Most of the eolian dust carried by low level winds are precipitated or deposited in adjacent oceans. A significant portion of the dust may be introduced into the upper part of the troposphere and can then be carried and deposited all over the globe. For example, the major source of lithogenic particles in the north Pacific is the Gobi desert and arid regions of China (Honjo, 1993). *Lisitzin* (*1972*) estimated that the annual accumulation rate of red clays in the World's ocean is about 1700 million tonnes which is nearly the same as eolian contribution to the oceans. This has lead *Honjo* (*1990, 1993*) and others to suggest that the major contributor of lithogenic particles to the open ocean is eolian dust.

Similar to eolian dust, an unknown but significant portion of volcanic ejecta may be deposited all over the globe. Unusually red sunsets due to volcanogenic particles in the atmosphere are noticed all over the globe following massive volcanic eruptions like the Karakatau eruption in 1883 and the El Chichon eruption in 1982.

Near mid- ocean ridges, authigenic clays derived from alteration of sea-floor basalts, are the dominant clay minerals. Compared to terrigenous material, cosmogenic fluxes are very low and important only in areas of extremely low sedimentation, such as red clay sediments found in the central part of the ocean basins.

Grain size studies of deep sea sediments show that the lithogenic fraction is composed of particles usually less than 4 micrometer in diameter. Clay sizes particles should theoretically have a long Stokesian residence time in the water column, in the order of a few hundred years. Fine particles introduced into the oceans should therefore be dispersed over a very wide area. Thus, clays introduced by the Ganges-Brahmaputra rivers should be found in every part of the world's ocean. However, it is seen that fine refractory particles settle quickly to the sea floor without significant dispersion (*Lisitzin, 1972*).

Previous sediment trap studies have shown that lithogenic matter form between 2 to 50 % of the total particulate matter in sediment traps (*Honjo, 1982; Honjo et al., 1982a,b, 1983, 1988; Wefer, 1989*). Annual lithogenic fluxes vary from less than 1 mg m^{-2} y^{-1} to over 8 g m^{-2} y^{-1}. Higher fluxes are found in areas close to continental margins. The smallest lithogenic flux ever recorded is in the Weddell Sea in the Antarctic below the ice-sheets where annual fluxes are approximately 1 mg m^{-2} y^{-1}. In the Arctic Ocean near Spitzbergen, annual fluxes were 14 g m^{-2} y^{-1} due to episodic flushing out of lithogenic matter from nearby fiords (*Honjo, 1993*).

Most sediment trap experiments show an increase in lithogenic fluxes with depth. This suggests large scale lateral transport of refractory particles in the deeper layers of the oceans by advection. This trend is most noticeable in areas close to continental margins where currents are most active. However, even in mid-oceanic areas fluxes of lithogenic matter have been shown to increase with depth (*Honjo et al., 1982; Deuser, 1983*).

The catchment area of the major rivers draining into the northern Indian Ocean lies in the Himalayas, rapid erosion of which has contributed nearly 2000 million tonnes of suspended sediment to the northern Indian Ocean every year (*Holeman, 1978; Milliman and Meade, 1983*). The monsoon winds also transport over 180 million tonnes of eolian dust into the area (*Goldberg and Griffin, 1970; Sirocko, 1989*).

In this chapter, lithogenic fluxes, their seasonality and processes controlling their settling rates in the northern Indian Ocean is discussed in detail.

5.2 Materials and methods

The samples used for study of lithogenic flux represent one full year, between May, 1986 and May 1987 in the case of the Arabian Sea and between November 1988 and November 1989 in the Bay of Bengal (See table 3.2 for details). Details of sample treatment is given in Chapter 3. In brief, the carbonate, organic matter and biogenic silica fraction was removed by leaching with dilute acetic acid, hydrogen peroxide and sodium carbonate respectively. It is assumed that the refractory portion left after removal of the above components consists essentially of lithogenic material.

Data from shallow traps in the Arabian Sea during the first year was unreliable and hence not included in this study. In subsequent years the shallow traps data continued to be erratic indicating that the samples are being consumed by zooplanktons, or altered by unknown biological processes. Reliable lithogenic flux data for both the shallow and deep traps during the same period are available only for WAST 06 collected between January 15 and September 28 1990 and EAST 03 collected between May 12 and October 21 1987. This data is included in this chapter to show the variation in lithogenic flux with depth.

5.3 Results and discussions

5.3.1 Lithogenic fluxes in the Arabian Sea:

Lithogenic flux patterns in the Arabian Sea (Table 5.1; Fig.5.1 to 5.3)) closely follow that of total fluxes, with high fluxes during the southwest monsoons. They vary by several order of magnitude from less than 0.01 to more than 50 mg m^{-2} day^{-1}, with more pronounced variations in the eastern and western trap located near the continents. Lithogenic fraction in the trap samples range between 4 and 30% with higher percentages in the eastern Arabian Sea. They also showed strong seasonality with 40 to 70% of the annual fluxes during the SW monsoon period. Peak lithogenic fluxes in the western Arabian Sea showed a sharp increase during July and August and low fluxes during the rest of the year. In the eastern Arabian Sea lithogenic fluxes increase at the end of July and showed high fluxes till the end of November whereas the central trap showed moderate fluxes between 5 and 20 mg m^{-2} day^{-1} through most of the year.

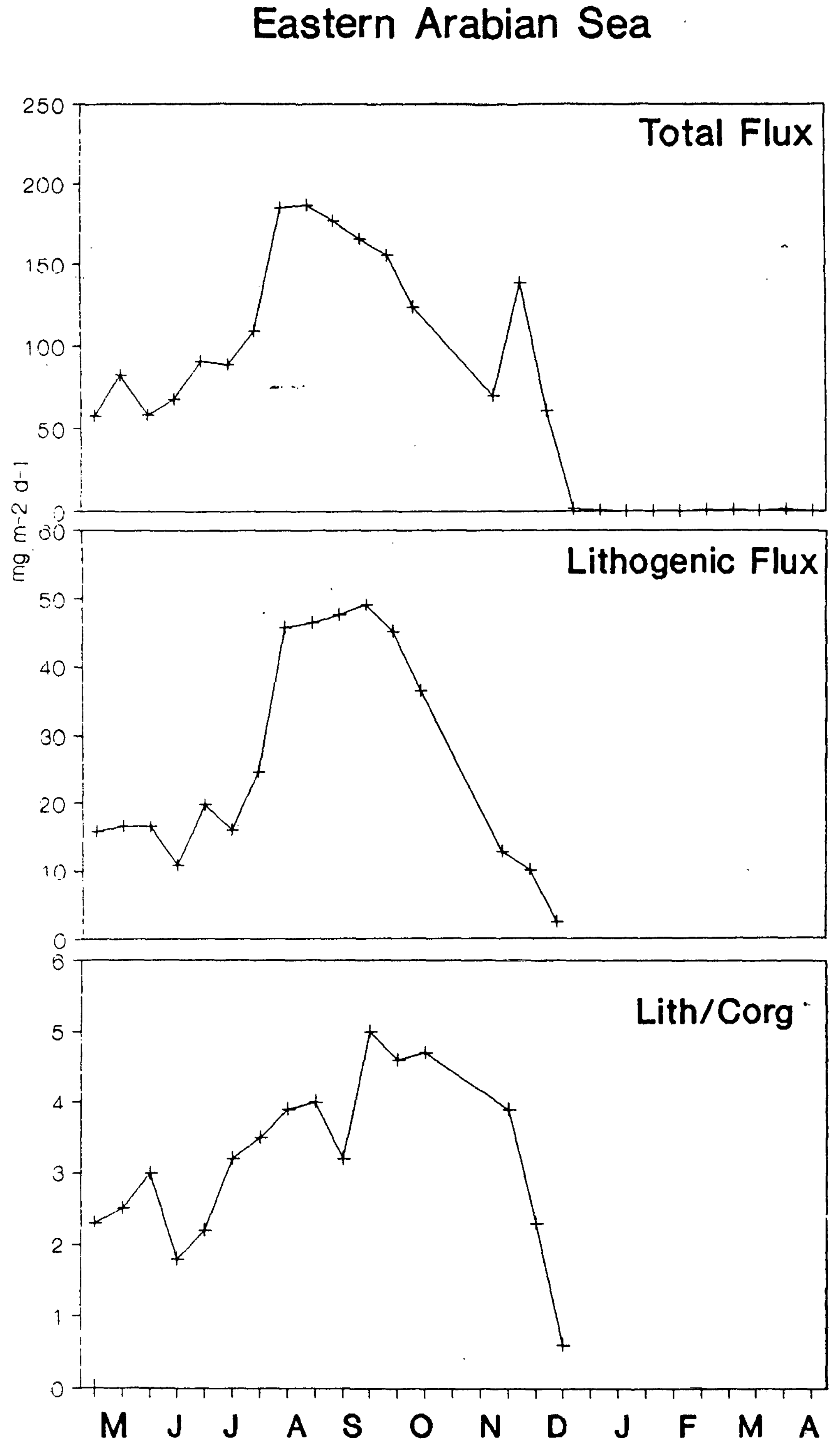


Fig 5.1 Total flux, lithogenic flux and lithogenic/ organic carbon (lith/c.org) ratios in the deep eastern Arabian Sea for the period 1986-87

Central Arabian Sea

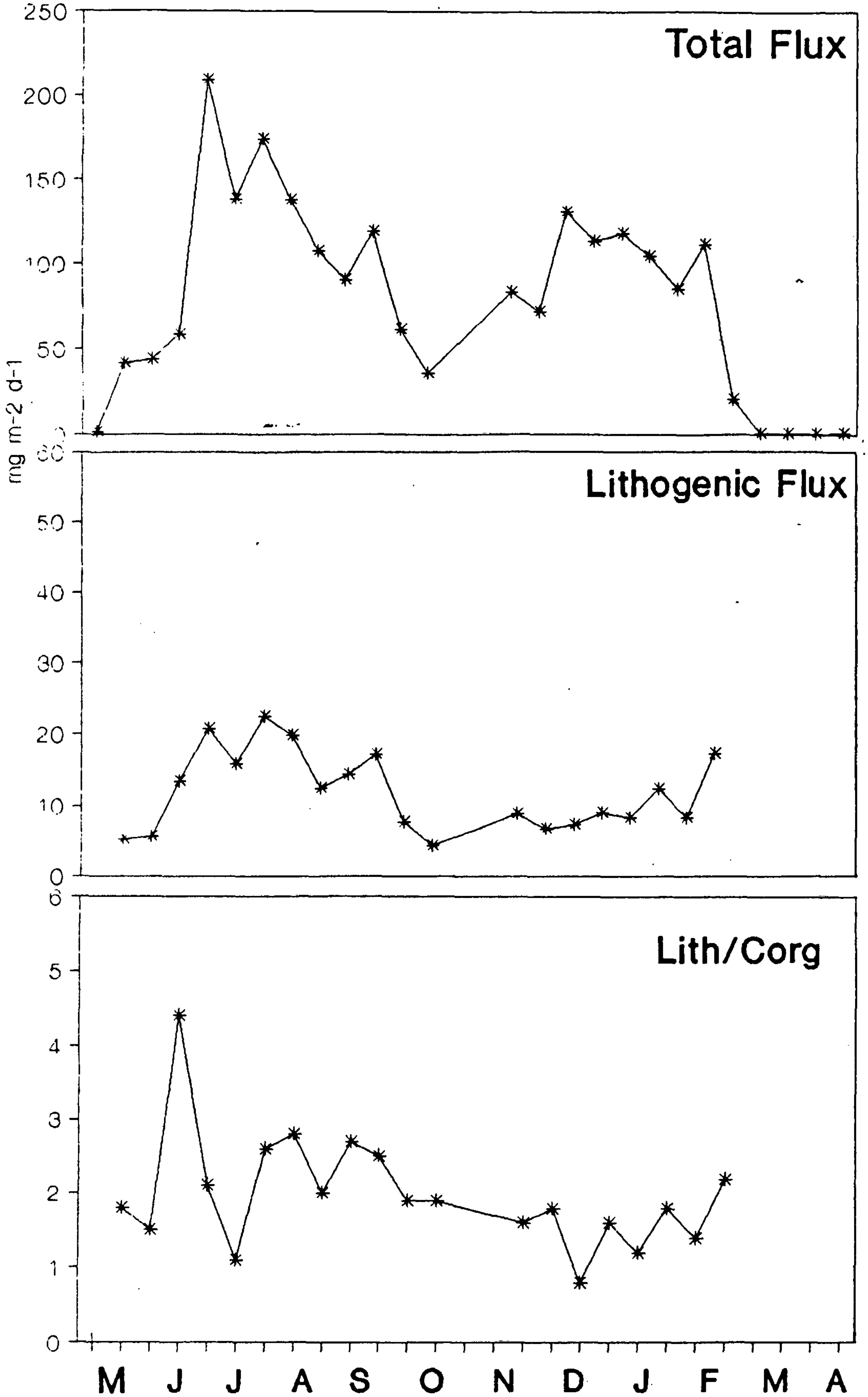


Fig 5.2 Total flux lithogenic flux and lithogenic/ organic carbon (lith/c.org) ratios in the deep central Arabian Sea for the period 1986-87

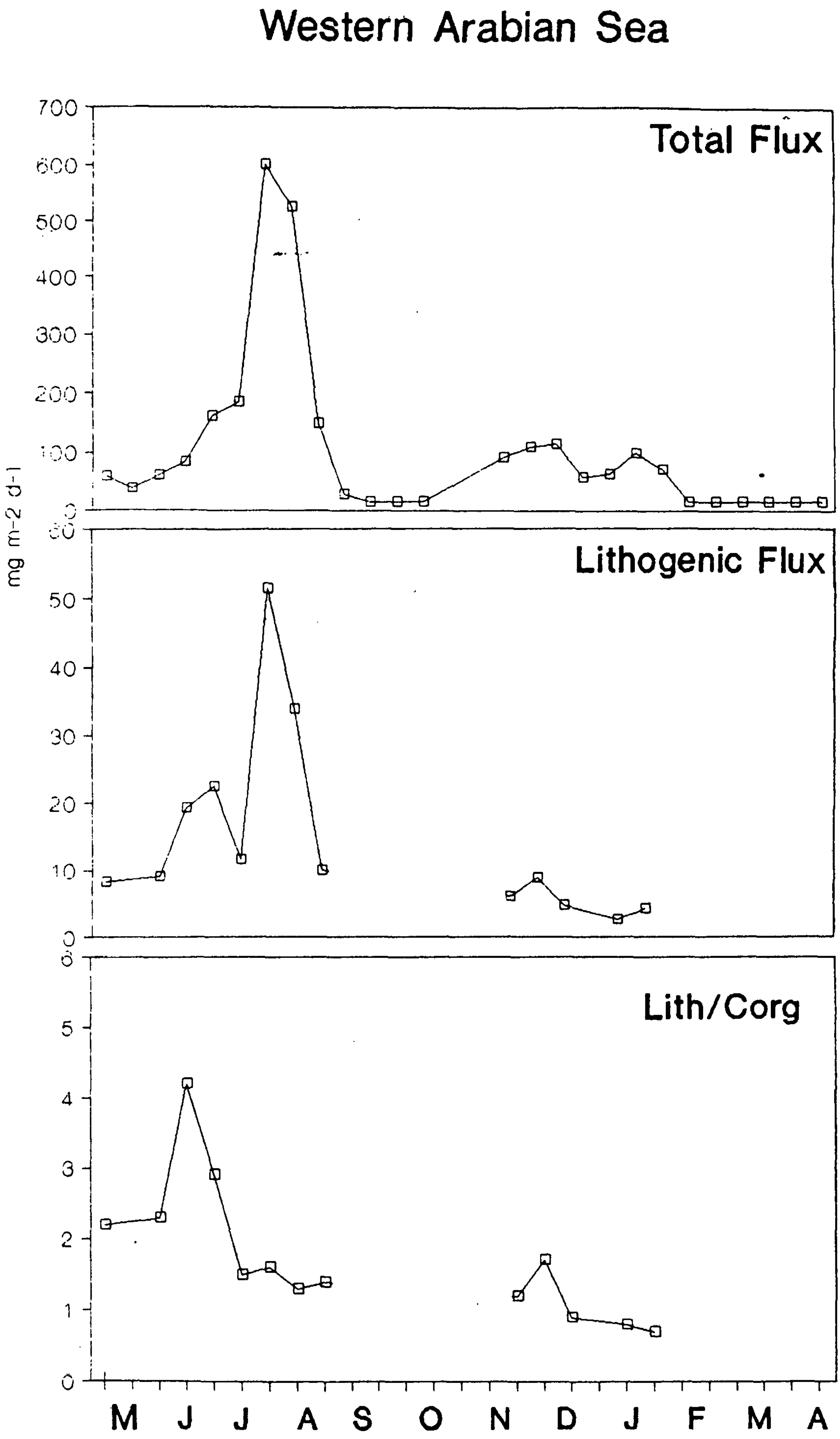


Fig 5.3 Total flux Lithogenic flux and lithogenic/ organic carbon (lith/c.org) ratios in the deep western Arabian Sea for the period 1986-87

Table 5.1 Total and lithogenic fluxes in the eastern and western Arabian Sea simultaneously measured in the shallow and deep traps. Though total flux decrease with depth there is not much variation in the lithogenic flux with depth in this area. The Eastern Arabian Sea moorings were deployed between January 15 and September 28 1990 and the western Arabian Sea moorings were deployed between May 12 and October 21 1987.

EAST 06 Shallow			EAST 06 Deep		
Cup No.	Total Flux	Lithogenic Flux	Cup No.	Total Flux	Lithogenic Flux
		All fluxes in mg m^{-2} day^{-1}			
1	43.53	10.2	1	51.23	12.9
2	61.96	13.3	2	57.59	13.1
3	72.35	16.4	3	64.65	15.3
4	76.55	15.9	4	70.21	13.9
5	132.6	28.0	5	129.92	26.0
6	141.1	30.2	6	127.76	23.7
7	268.75	38.2	7	192.45	35.0
8	195.87	28.6	8	156.17	29.9
9	149.04	30.7	9	110.05	28.1
10	32.72	7.2	10	34.23	8.2
11	3.05	0.0	11	38.37	10.2
12	5.17	1.3	12	11.38	n.d
13	2.6	0.0	13	0.79	n.d

WAST 03 Shallow			WAST 03 Deep		
Cup No.	Total Flux	Lithogenic Flux	Cup No.	Total Flux	Lithogenic Flux
		All fluxes in mg m^{-2} day^{-1}			
1	74.88	8.9	1	30.27	2.5
2	65.37	8.2	2	53.62	4.6
3	45.97	6.6	3	40.22	3.3
4	164.35	20.8	4	136.52	10.6
5	95.62	16.9	5	124.3	18.0
6	220.86	32.8	6	155.01	16.3
7	184.74	16.3	7	163.99	15.8
8	202.71	19.4	8	204.07	22.3
9	309.18	21.6	9	262.78	24.4
10	652.76	44.6	10	404.87	32.5
11	292.02	15.7	11	347.97	27.9
12	80.84	8.7	12	240.56	23.3
13	0.35	0.0	13	199.38	15.8

During 1986-87 an east-west gradient in lithogenic fluxes was observed with annual fluxes decreasing from 5.4 g m^{-2} in the eastern Arabian Sea to 2.64 g m^{-2} in the west. The higher fluxes in the eastern Arabian Sea may be due to the clockwise surface circulation during the southwest monsoon period, which may have diverted the Indus discharge toward the eastern part of the Arabian Sea. This is supported by recent studies on the clay mineral distribution pattern on the western continental shelf of India (*Ramaswamy and Nair, 1989*) which has shown that the inner shelf consists of material derived from the contemporary rivers draining the west coast of India whereas the outer shelf clays appear to have been derived from the Indus river.

Seasonal variations: About 40% of the annual lithogenic flux to the deep western Arabian Sea was brought down within a span of 25 days during the highly productive months of July and August (Fig. 5.3). Organic carbon fluxes also peak during the SW monsoon and has a similar flux pattern. The rapid sedimentation of lithogenic matter in association with organic matter points to the significant role of pelagic organisms in sedimentation of terrigenous matter.

Although peak discharge of the Indus is during July and August (*Ittekkot and Arain, 1986*) high lithogenic fluxes in the eastern trap continue till the end of October (Fig.5.1) indicating that lithogenic particles are removed from the water column only in association with the higher fluxes of biogenic particles during this period. Another reason for the lag in lithogenic fluxes to the eastern Arabian Sea may be the time taken for the suspended sediments from the Indus to reach the trap site. The surface currents in the eastern Arabian Sea during July and August is normally between 18 and 36 km per day (*West Coast of India, Pilot, 1975*). This means that the peak discharge of the Indus in July and August should reach the eastern and central trap site after 30 to 60 days. The dissipating phase of the summer upwelling was noticed off the coast of Goa during September 1986 (*Madhupratap et al., 1992*). This must have provided the necessary 'carrier particles' and may explain the high lithogenic fluxes in the eastern Arabian Sea during September and October. It is also possible that lithogenic particles in the eastern trap are not derived directly from rivers but are resuspended sediments advected from the shelf edge of the west coast of India.

In the western Arabian Sea, eolian sediment dominate. From the study of dust outbreaks from satellite photographs for the year 1979, *Sirocko* (*1989*) has showed that more than 90% of the annual transcoastal flux of dust is transported during the months of June, July and August. This compares well with our data which indicate that about 80% of the annual lithogenic fluxes to this area is during the SW monsoon period.

During the NE monsoon period only about 10% of the annual lithogenic fluxes reach the sea floor. This is surprising since very high concentrations of eolian matter have been reported over the Arabian Sea during this period (*Chester, 1985*). The reason for the above seasonal variations may not be solely due to availability of terrigenous material at the sea surface rather it may depend more on biological factors controlling their sedimentation as will be elaborated on later in this chapter.

Increase in lithogenic fluxes with depth: Simultaneous measurement of lithogenic fluxes to the shallow and deep traps in the eastern and western Arabian Sea is shown in Fig. 5.4 and table 5.1. No significant increase in lithogenic flux with depth, except for a short period towards the end of the SW monsoons, is seen at both the sites. Similar studies in the Bay of Bengal (*Ittekkot et al., 1991* and the present study) show a sharp increase in

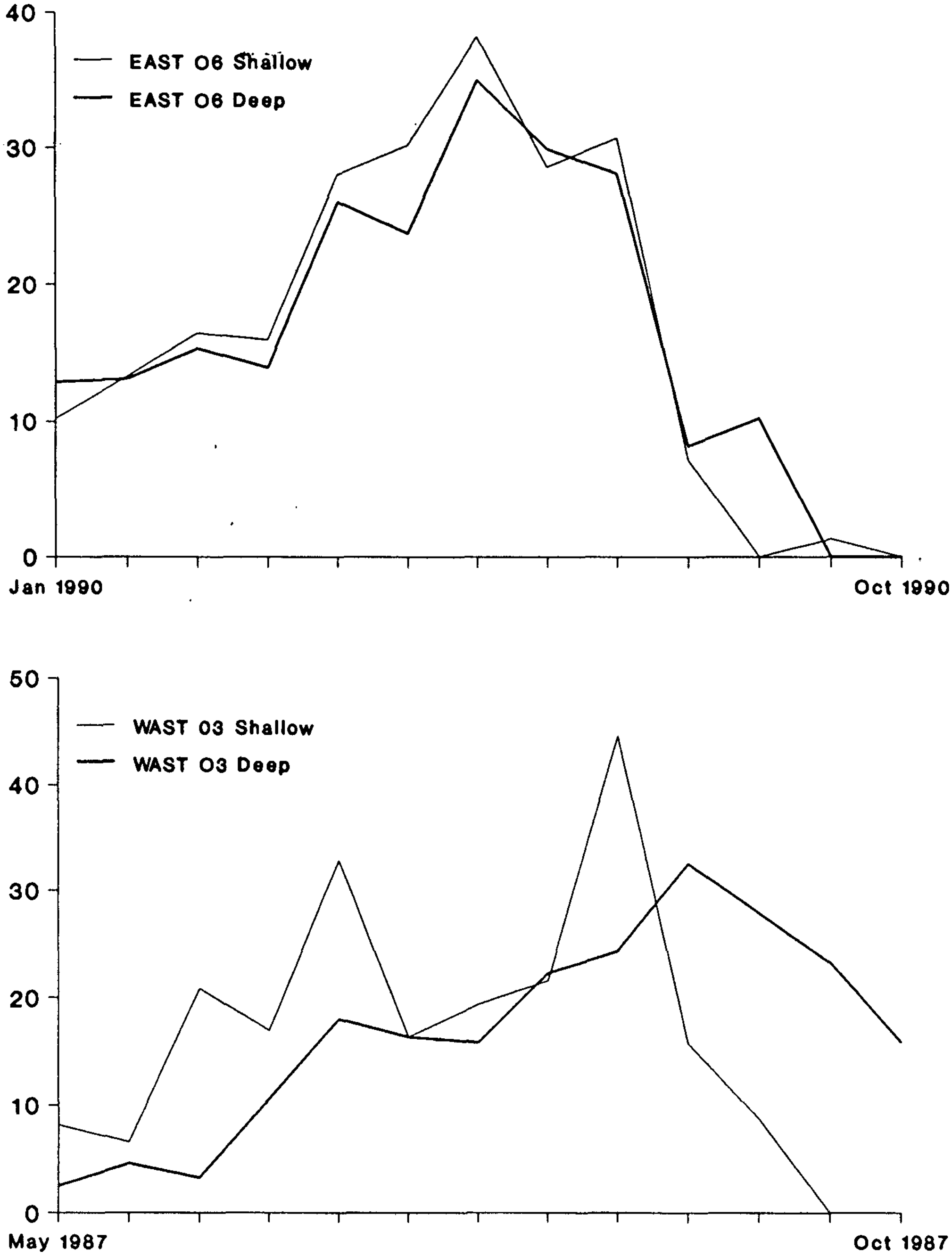


Fig. 5.4 Comparison between lithogenic fluxes in the shallow and deep traps in the eastern (May to October, 1987) and western Arabian Sea (January to September, 1990)

lithogenic fluxes with depth. The reason for the lack of increase of lithogenic flux with depth in the Arabian Sea may be because of low concentrations of suspended lithogenic matter in the deeper waters due to reduced input of fluvial sediments from the Indus. Another reason may be that, presently the dominant mode of terrigenous matter input in this area is by eolian dust which is introduced into the oceans at its surface.

5.3.2. Lithogenic fluxes in the Bay of Bengal

Lithogenic material comprise between 23 to 30% of the total material in the sediment trap samples of the Bay of Bengal (Table 4.5). River discharge obviously accounts for a major part of the lithogenic flux as the traps are located about 400 to 500 km from the mouths of some major rivers like the Ganges-Brahmaputra and Godavari (Fig 1.1). Annual lithogenic fluxes to the deep traps in the northern and central Bay of Bengal during 1988-89 is 14.4 and 16.6 g m^{-2} y^{-1} respectively. Lithogenic fluxes measured here are amongst the highest reported in sediment trap experiments in the World Oceans (*Wefer, 1989*) and are about 3 to 5 times higher than that of the Arabian Sea. Interannual variability is more pronounced in the northern part of the Bay where annual lithogenic fluxes decreased from 27.96 g m^{-2} in 1987-88 to 14.98 g m^{-2} during 1988-89. During 1987-88, when the traps at all three sites functioned correctly, a strong north-south gradient in lithogenic fluxes were seen with annual fluxes decreasing from 27.9 g in the northern Bay of Bengal to 8.5 grams in the southern part of the Bay (*Ittekkot et al., 1991*). The annual lithogenic flux of 27.9 g m^{-2} reported for the northern Bay of Bengal in 1987-88 is the highest lithogenic flux ever reported in the open ocean.

Seasonality of lithogenic fluxes: Lithogenic fluxes follow total flux patterns and peak during the SW monsoon period (Fig. 5.5 to 5.9). Between 40 to 52 % of the total annual lithogenic fluxes to the deep northern and central traps

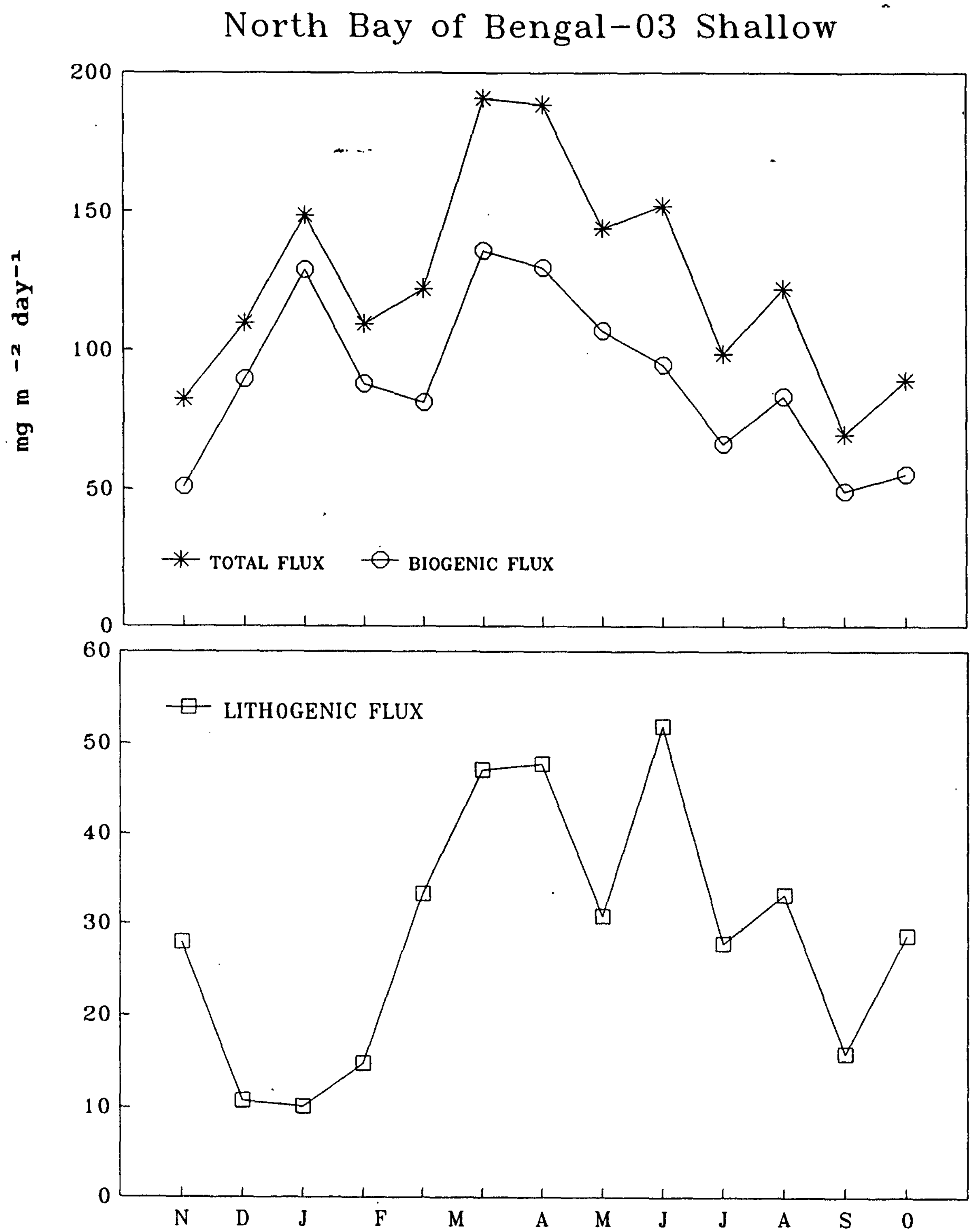


Fig 5.5. Time-series data of total (<1mm fraction), biogenic and lithogenic flux in the shallow northern Bay of Bengal trap for the period November 1988 to October 1989.

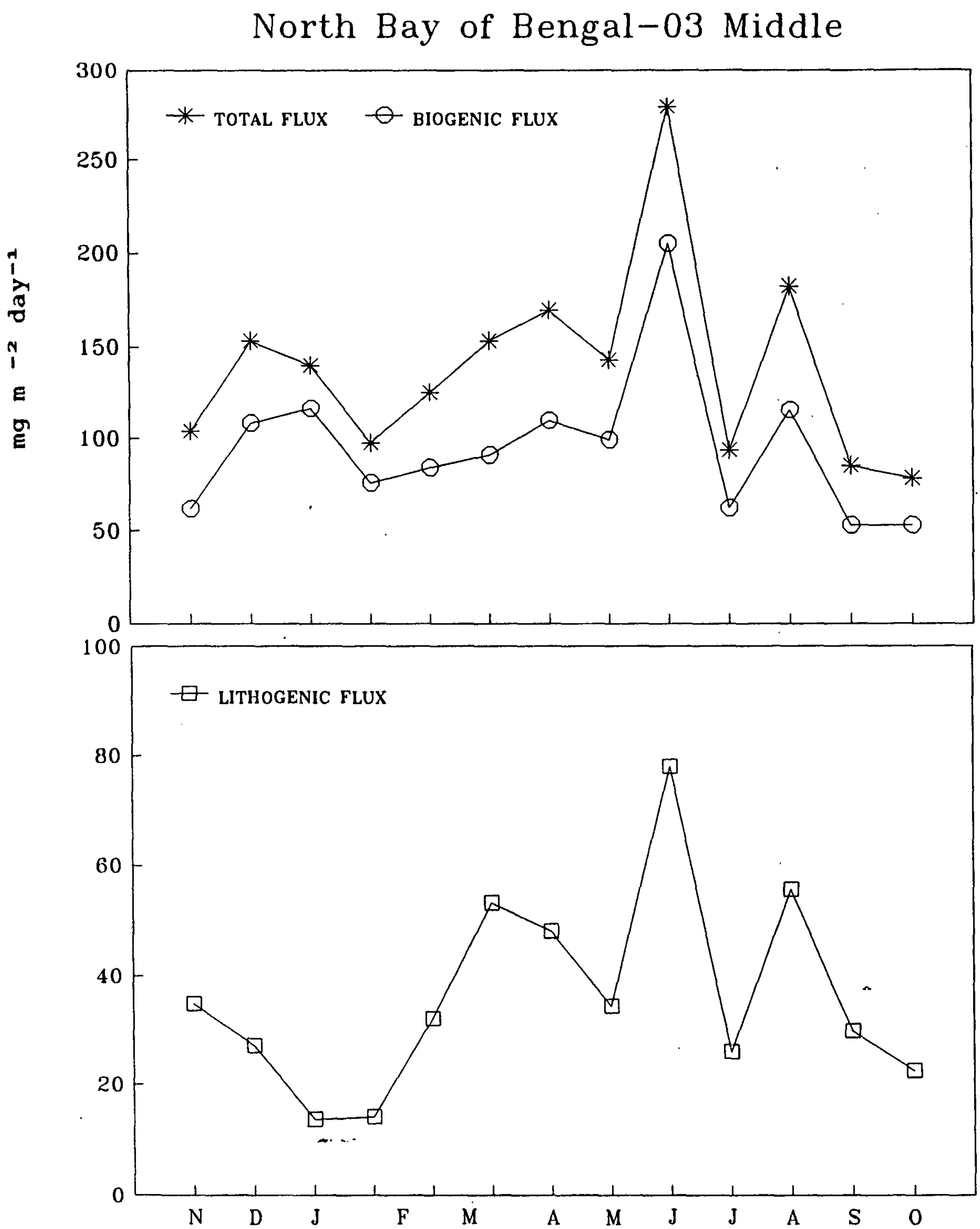


Fig 5.6 Time-series data of total (<1mm fraction), biogenic and lithogenic flux in the middle northern Bay of Bengal trap for the period November 1988 to October 1989.

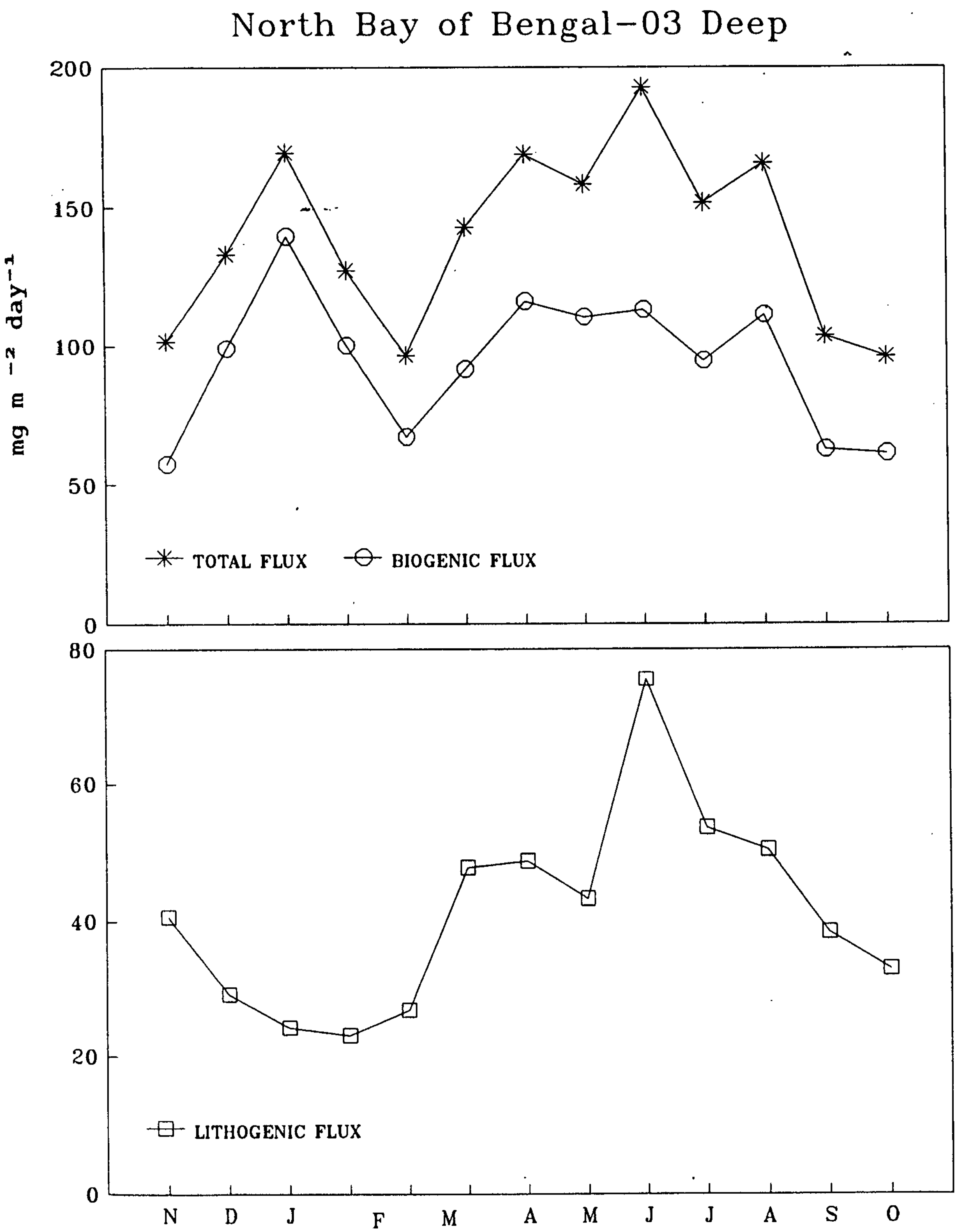


Fig 5.7 Time-series data of total (<1mm fraction), biogenic and lithogenic flux in the deep northern Bay of Bengal trap for the period November 1988 to October 1989.

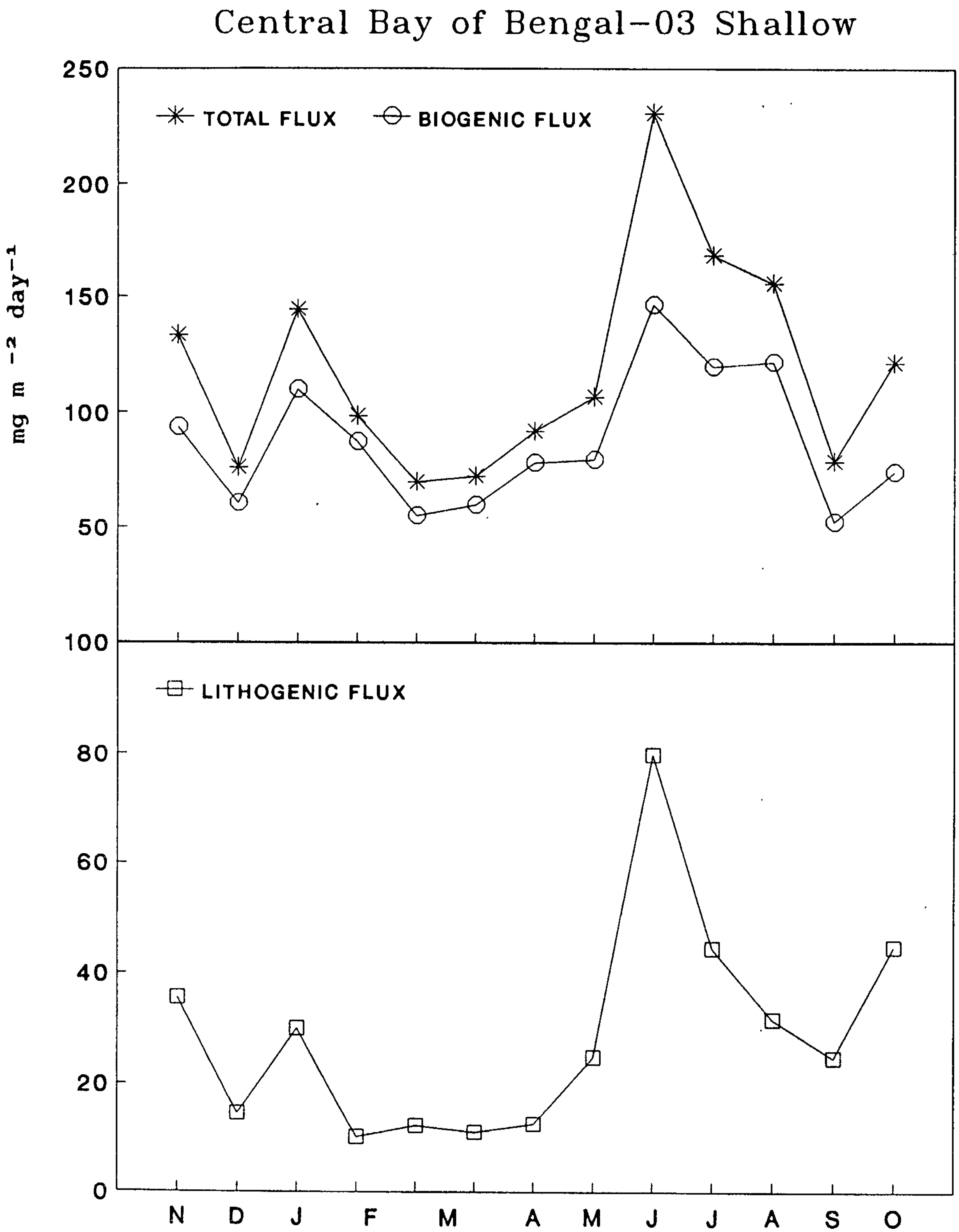


Fig 5.8 Time-series data of total (<1mm fraction), biogenic and lithogenic flux in the shallow central Bay of Bengal trap for the period November 1988 to October 1989.

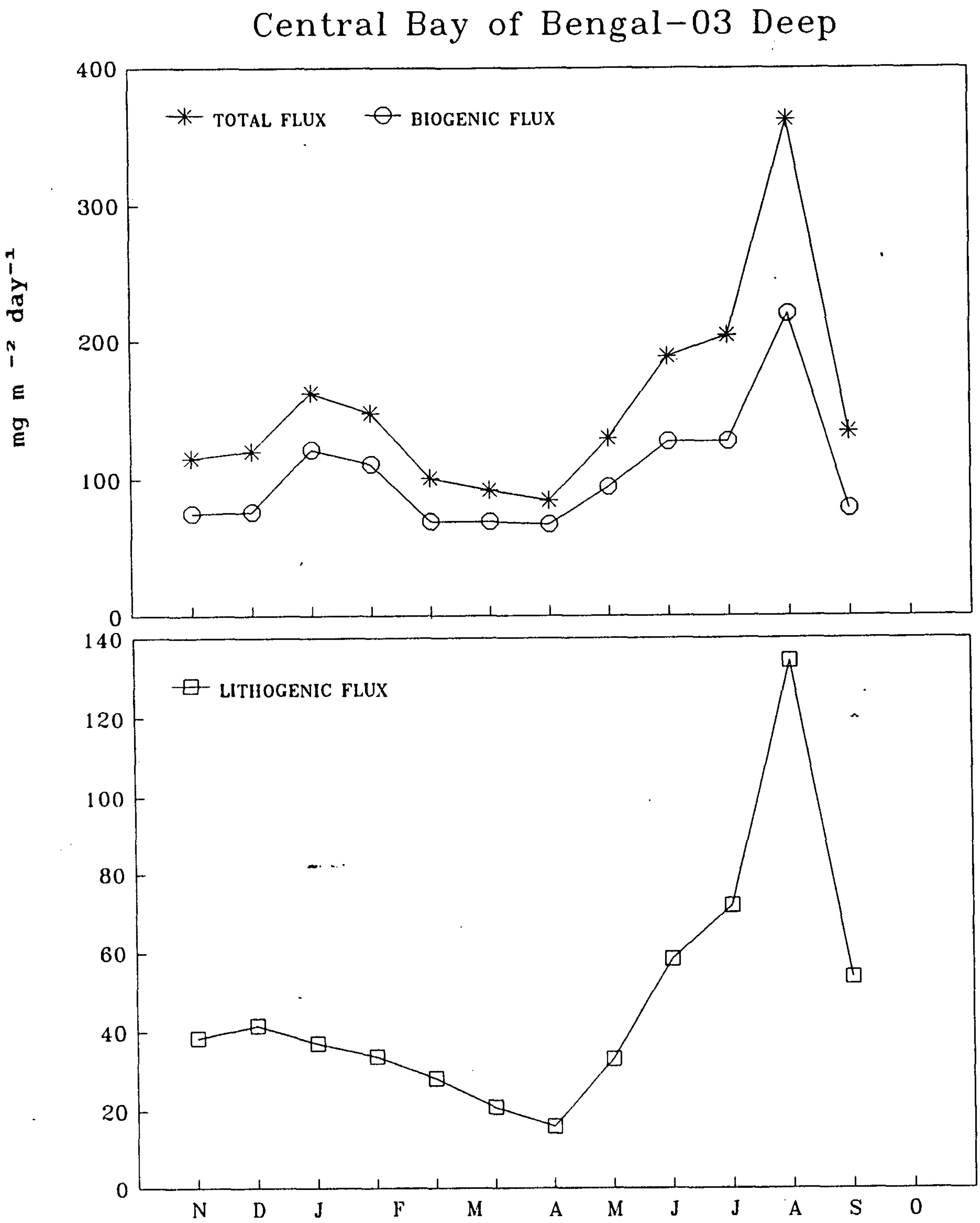


Fig 5.9 Time-series data of total (<1mm fraction), biogenic and lithogenic flux in the deep central Bay of Bengal trap for the period November 1988 to October 1989.

are during the SW monsoon period. The NE monsoon peak is not very prominent as river discharge is low during this period. In the northern part of the Bay, lithogenic fluxes are moderate to high during April and May due to meltwater discharge from the Himalayas (*Rao, 1977*) as well as eolian input from the Indian Peninsula (*Middleton, 1989*).

Increase in lithogenic flux with depth: At both the sites lithogenic fluxes increase with depth. Although increase in lithogenic flux with depth has been reported at many other locations, it is not easy to compare data from different areas as the depth of deployment of the traps are variable. As the rate of increase of lithogenic fluxes with depth is linear, the gradient of this increase can be calculated if we know the trap depth and the lithogenic fluxes at the two depths. Increase in lithogenic flux with depth from different locations can be compared by calculating what *Honjo et al* (*1982 b*) have termed the 'R'factor. R is the increase in lithogenic flux per unit cubic meter of water per day expressed as microgram m^{-3} day^{-1}. It is calculated by the formula R= F2-F1/D2-D1 where, F2 and F1 are the lithogenic fluxes in mg m^{-2} y^{-1} at different depths and D2 and D1 are the trap depth in meters (*Honjo et al., 1982b*).

R values give an indication of the gradient of increase in lithogenic flux with depth. If R values are less than 0 it means that particles are dissaggregating faster than their rate of aggregation or scavenging. Particle flux studies carried out with more than two sediment trap in a single mooring have shown that generally increase in lithogenic flux with depth is linear (*Honjo et al., 1982b*).

Annual average R values for the Bay of Bengal during 1987-88 were 9.3 and 18.1 microgram m^{-3} day^{-1} in the northern and central Bay of Bengal respectively. This shows that the rate of scavenging is extremely high in this area. In the southern Bay of Bengal R values are close to zero suggesting that the rates of disaggregation and scavenging are close to each other. The rate of scavenging of lithogenic particles here is perhaps the highest in the world because of the extremely high fluvial discharge as well as high biological productivity of the surface waters. For comparison, it may be noted that the lowest R values reported for the open ocean is 0.13 microgram m^{-3} day^{-1} in the Mid-Pacific east of Hawaii while the highest value reported is 17 microgram m^{-3} day^{-1} for the Panama Basin (*Honjo et al., 1982*).

Seasonal variation in the rate of scavenging of lithogenic particles has not been previously reported in literature. For the first time, time-series measurements of R values are being reported in this thesis. Time-series data on the gradient of lithogenic flux can give important information on the seasonality of scavenging in the water column by settling particles.

Time-series measurements of R values in the northern and central Bay of Bengal during 1988-89 (Fig. 5.10) show that scavenging processes dominate for a major part of the year. In the central Bay of Bengal aggregation processes dominate throughout the year, except for one cup interval in June (Fig. 5.10). The rates of scavenging is, however, not constant. The rate of scavenging is extremely high during the SW monsoon period and comparatively low during the intermonsoon period (Fig. 5.10). R values may be as high as 80 microgram m^{-3} day^{-1} during the SW monsoon and close to or less than zero during the pre SW monsoon period. During peak river discharge periods, not only is suspended lithogenic matter introduced into the oceans but the freshwater and associated nutrients can change productivity patterns and increase availability of organic aggregates or marine snow in the oceans (*Ittekkot et al, 1991*). High R values during the SW monsoon can therefore be attributed to simultaneous increase in primary

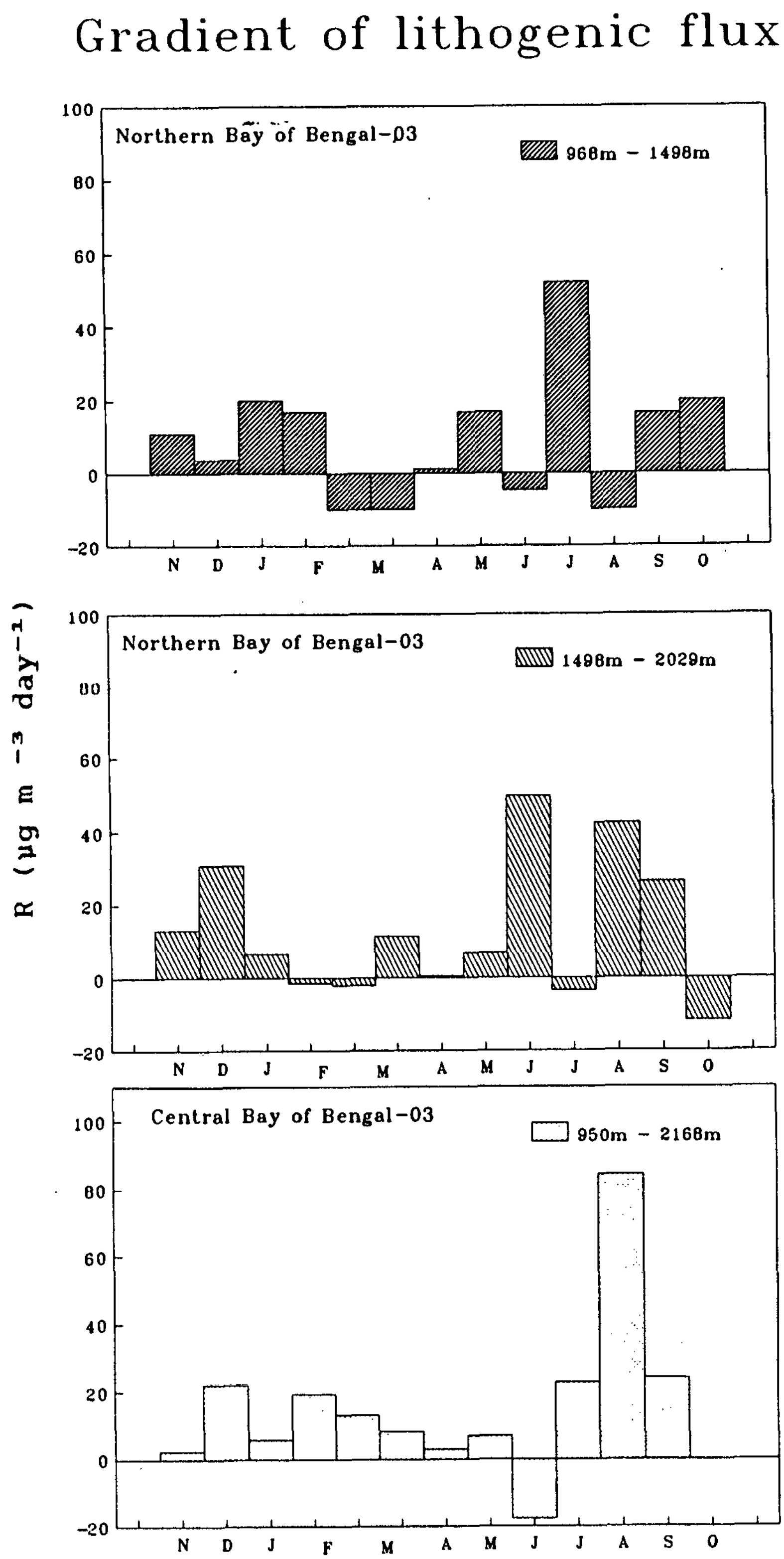


Fig 5.10. Seasonal variations in gradient of lithogenic flux (R values in μg m^{-3} day^{-1})) in the northern and central Bay of Bengal. Positive values indicate aggregation and negative values indicate disaggregation of particles.

productivity and suspended load of the Bay of Bengal during the SW monsoon period.

5.3.3. Biological control on lithogenic particle sedimentation:

Fine inorganic particles like clays sink at very slow rates (<0.5 m day $^{-1}$) in the water column and may take years to reach the sea floor on their own (*Lal, 1977*). These fine particles could however, collide with larger, faster sinking particles due to differential settlement or shear and adhere to them because of the sticky organic mucous coating them or get incorporated into the matrix of marine snow (*Lal, 1977; Alldredge and Silver, 1988*). Grazing by pelagic organisms could further repackage these aggregates and accelerate their sinking rates and transport the fine lithogenic particles to the sea floor within a few weeks (*McCave, 1975, 1984*). Foraminifera and radiolarians, which make up more than 50 % of the total flux in the Arabian Sea, may not play an active role in the transport of clays. Fecal pellets contain clay particles, but their overall contribution to the total flux is not very important. Fecal pellets have been shown to contribute only 0.5 to 4% of the total lithogenic flux in the Pacific (*Pilskaln and Honjo, 1987*). Large amorphous aggregates of marine snow are the most likely particles responsible for the transport of clays to the sea floor (*Honjo, 1982; Alldredge and Silver 1988*).

If we assume that the amount of organic aggregates responsible for transportation of clays is proportional to the organic carbon values, then, the lithogenic flux / organic flux (Lith./C.org) ratios of the trap material will reflect the importance of biological processes in the transport of lithogenic particles (Fig 5.1 to 5.9). There is a good correlation between lithogenic and organic carbon fluxes at all the sites in the northern Indian Ocean (Fig. 5.11 to 5.13). The Lith./C.org values in the world oceans generally range between 0.2 and 10 (*Honjo et al., 1980, 1982, 1987,1988*). Low values (< 0.5) are generally

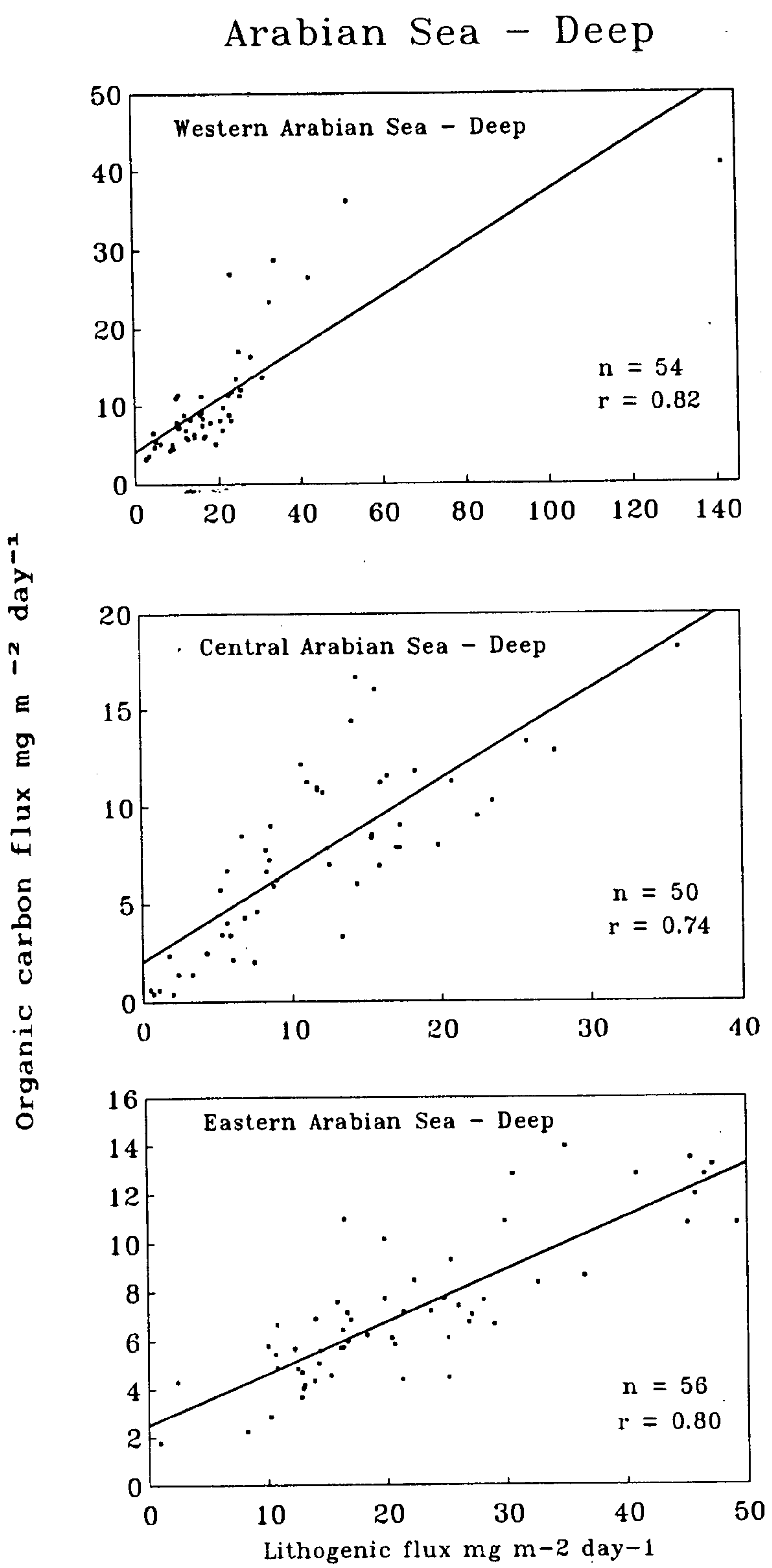


Fig 5.11. Lithogenic flux versus organic carbon flux in the western, central and eastern Arabian Sea. A positive correlation between lithogenic and organic carbon fluxes is seen at all the three sites.

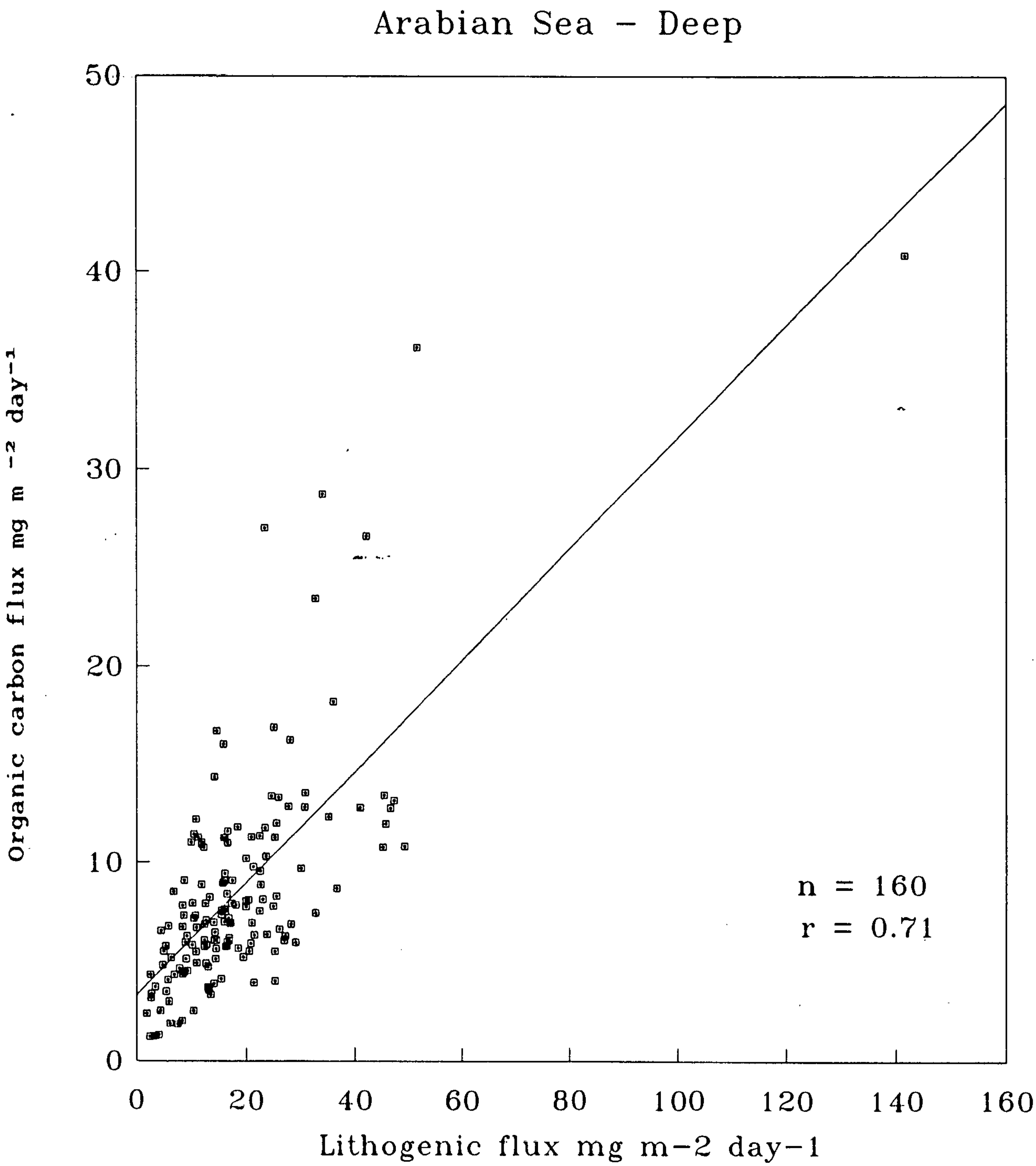


Fig.5.12. Correlation between lithogenic and organic carbon fluxes in the Arabian Sea . All available data for the Arabian Sea (samples collected between May 1986 and September 1990) has been included in this figure.

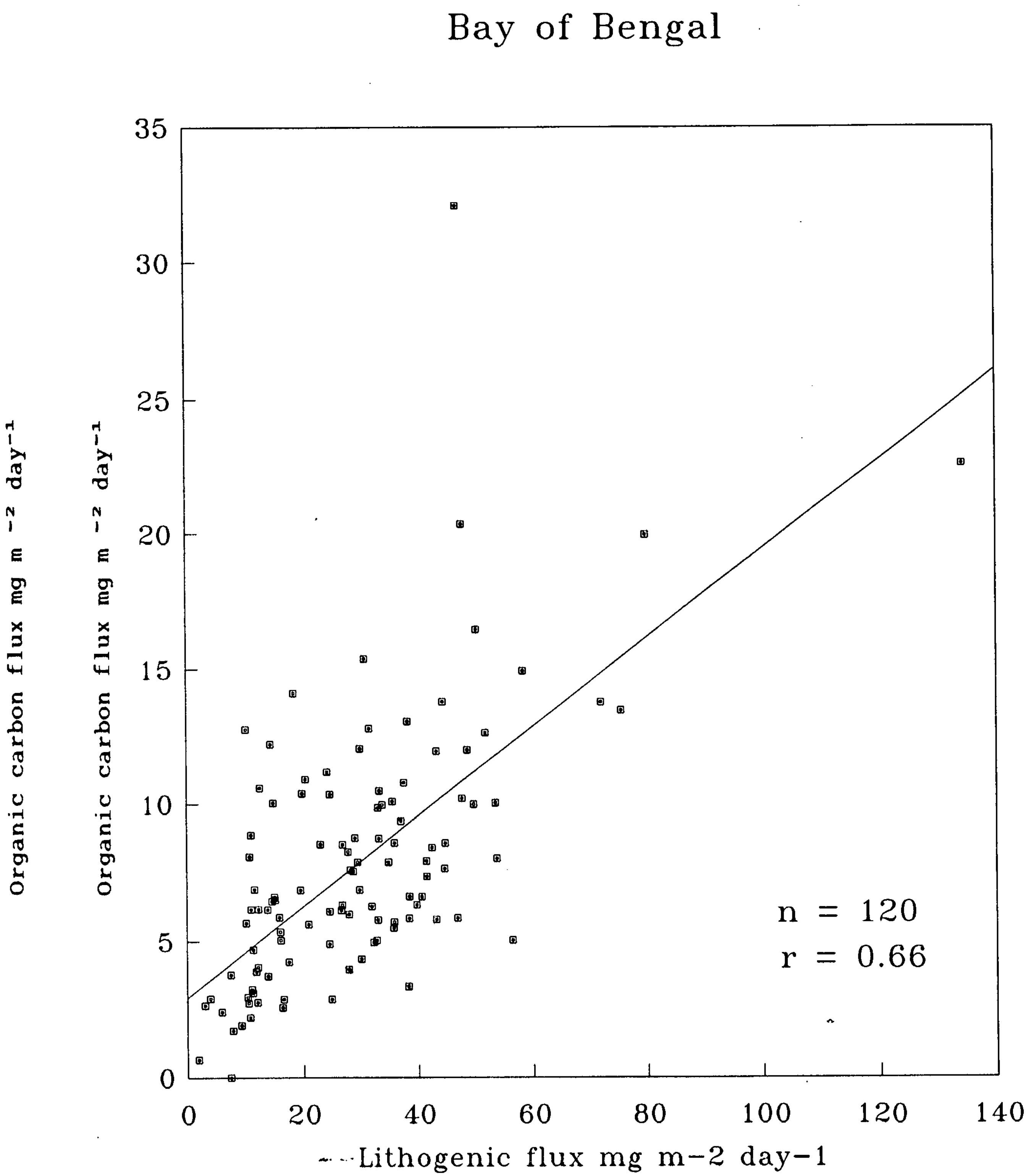


Fig 5.13 Correlation between lithogenic and organic carbon fluxes in the Bay of Bengal. All available data for the Bay of Bengal (samples collected between May 1987 and September 1989) has been included in this figure.

found in areas far away from continental masses and in the Antarctic. High values are found in traps deployed in nepheloid layers or in areas with strong advection from nearby continental margins (*Honjo, 1986*).

In the Arabian Sea, Lith/C org. values range between 1 and 5. In the eastern Arabian Sea, particles exhibit much higher Lith/ C org. ratios reflecting greater availability of lithogenic particles which can get into the matrices of large amorphous aggregates. The sharp increase in the lith /C org. values in the central and western trap during the month of June (Fig. 5.2 and 5.3) coincides with the beginning of eolian input from the Arabian deserts by the SW monsoon winds. What is surprising is that the peak lithogenic fluxes in the western trap during August is not reflected in higher Lith/C org. ratios. If the high lithogenic fluxes were solely due to lateral advection or a dust outbreak the Lith/C.org ratios would have been raised substantially. The simultaneous increase of other components including organic carbon suggests the absence of any episodic lithogenic input. Rather, the high lithogenic fluxes are a consequence of its association with export of biological matter from the upper mixed layer. Recent papers have shown that significant export of diatom floccules can occur as episodic seasonal events following phytoplankton blooms (*Honjo et al., 1982; Billet et al., 1983; Alldredge and Gotschalk, 1989*). *Takahashi* (*1983*) has shown that some diatom species like *Rhizosolenia* are the main contributors of silica during bloom periods. In our traps we have noticed up to 40% opal during peak fluxes in July and August. These rapidly settling (>100m day $^{-1}$) floccules of diatoms as well as clay enriched fecal pellets could provide the necessary matrix for aggregation of discrete clay particles.

Compared to other regions Lith./ C.org values observed in the Bay of Bengal are very high and generally ranges between 1 to 2 during low flux periods to

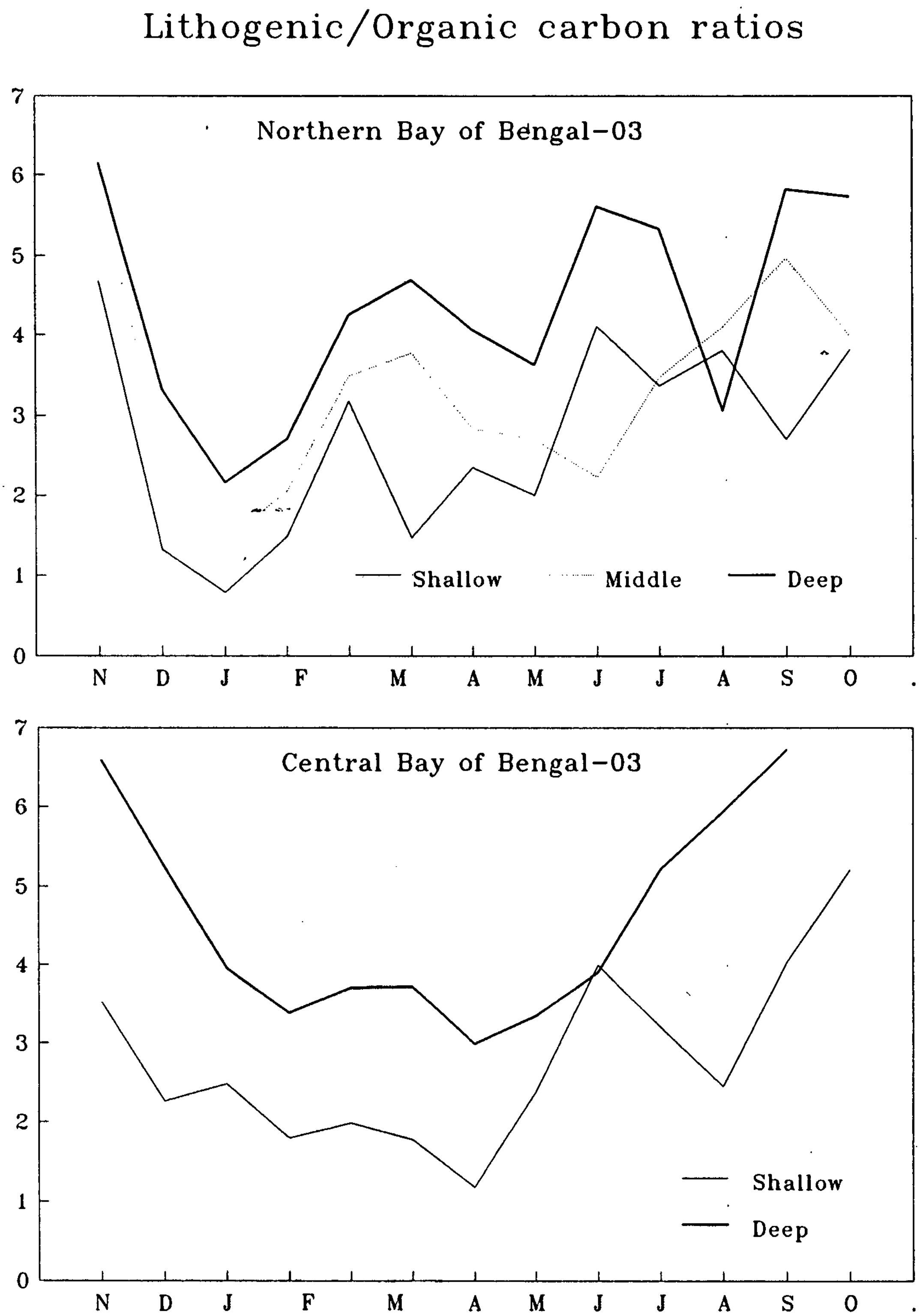


Fig 5.14 Lithogenic/Organic carbon ratios in settling particles in the northern and central Bay of Bengal.

over 4 during the SW monsoon (Fig. 5.14). The correlation between lithogenic and organic carbon fluxes (Fig. 5.13) though good are not as high as in the Arabian Sea. Cooperative sedimentation of lithogenic and biogenic particles is obscured in the northern and central Bay of Bengal during the SW monsoon period when lithogenic particles were overwhelmingly dominant.

Increase in Lith/ C.org values with depth is very clearly seen in both the northern and central Bay of Bengal. This increase is due to increase in lithogenic flux with depth due to scavenging by organic aggregates as well as decrease of organic matter with depth due to remineralization in the water column.

In the northern Bay of Bengal lowest lith/C.org values are observed during January when suspended sediment discharge of rivers is minimal while in the central Bay of Bengal lowest ratios were observed in April. In the northern Bay of Bengal lith./C.org values increase from May onwards indicating greater availability of suspended lithogenic particles in the water column due to increase in freshwater and suspended sediment discharge of the Ganges-Brahmaputra rivers. The peak in March may be related to eolian dust input from the Indian subcontinent. In the central Bay of Bengal peak lithogenic fluxes are during July-August (Fig. 5.8 and 5.9) but lith/C.org values continue to be high till November. This shows that though biological primary productivity in the surface layers decrease after the monsoon, the water column still has high concentrations of suspended lithogenic particles.

Our studies strongly support earlier works by *Deuser et al.* (*1983*) that the flux of lithogenic materials in the open ocean is dependant not only on its concentration at the surface, rather, it depends more on the concentration and efficiency of scavenging by biogenic matter. Higher lithogenic percentages in marine snow or fecal pellets should enhance their sinking rates as the density of lithogenic particles is higher than organic matter. This may lower the residence times and degradation rates of these particles in the water column. On the sea floor 75 to 85 % of the freshly sedimented organic material can be decomposed within a year (*Cole, Honjo and Erez, 1987*, see also table 8.1)). Higher lithogenic content in these particles should enhance preservation of organic carbon by burying it (*Calvert, 1987*). High organic carbon contents in the sediments of the Arabian Sea (*Paropkari, 1992*) may thus result from high Lith./C.org ratios in sedimenting particles.

Importance of scavenging processes in marine sedimentation: It will be of interest to know the proportion of lithogenic particles derived from the surface layers by aggregation and that scavenged by organic aggregates while settling through the water column. In shallow areas near river mouths because of high suspended matter concentration most of the lithogenic matter settles out due to flocculation or biological processes of aggregation while in the areas dominated by eolian particles scavenging rates in the water column is expected to be low.

Figure 5.15 shows the gradient of increase of lithogenic flux with depth at three sites in the Bay of Bengal and in the eastern and western Arabian Sea. In areas dominated by fluvial sediments, i.e. the northern and central Bay of Bengal, lithogenic fluxes show a sharp increase with depth while at other sites lithogenic flux is almost constant with depth.

The annual average increase in lithogenic flux with depth in the northern Bay of Bengal for the year 1988-89 is 10.9 microgram m^{-3} day^{-1}. If we extrapolate the gradient of lithogenic flux for this station we see that the lithogenic flux at the base of the photic zone (100m) is expected to be 6.3 g m^{-2} y^{-1}. This means that of the 14.4 g of sediment trapped in the deep sediment trap 6.3 g is derived from the photic zone while the remaining 8.1 g is derived by

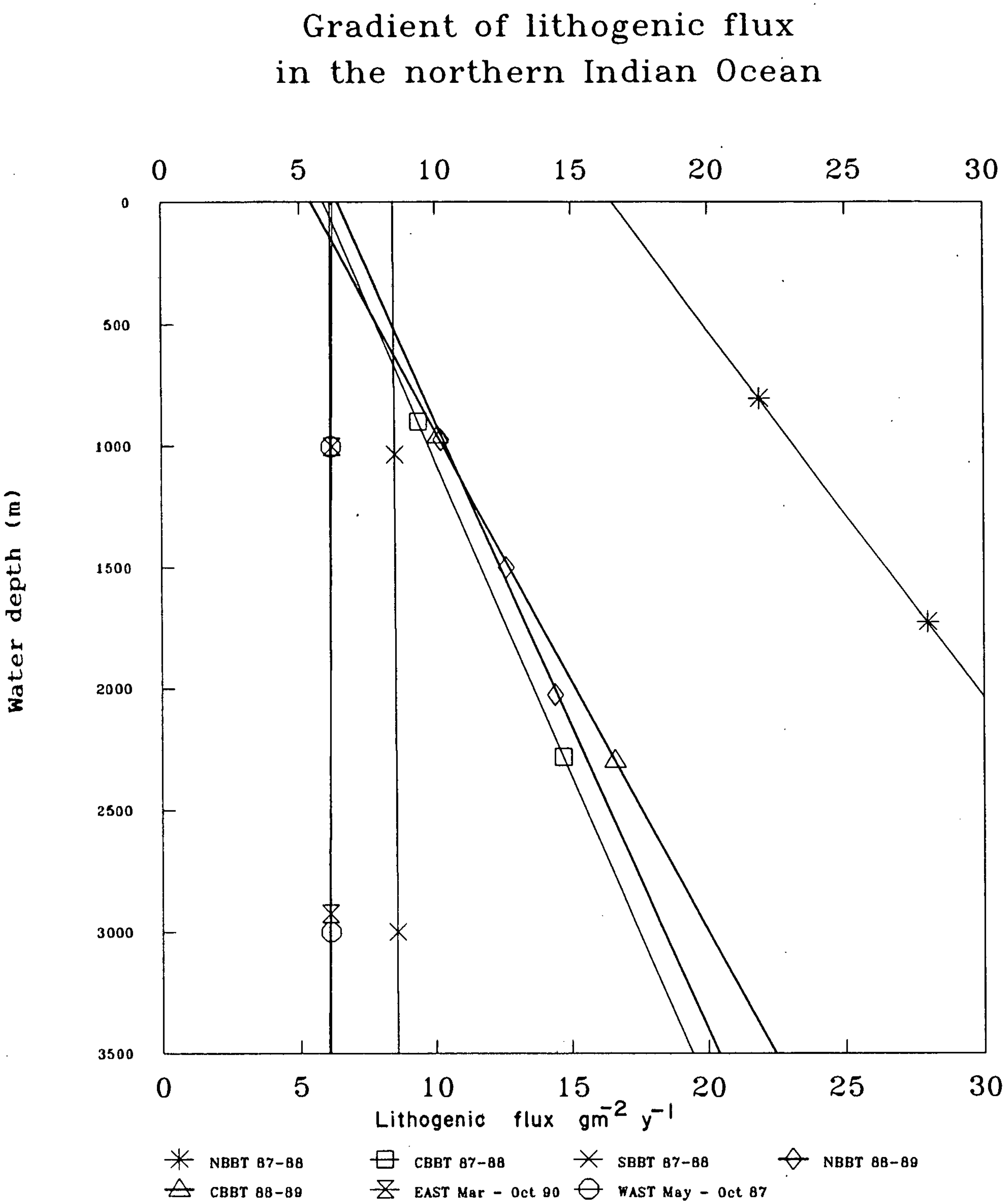


Fig 5.15 Gradient of lithogenic fluxes in the Bay of Bengal and Arabian Sea. Lithogenic fluxes in the northern and central Bay of Bengal show a steep gradient while the southern Bay of Bengal and Arabian Sea show hardly any variation with depth.

scavenging processes in the water column. Thus about 60 per cent of the lithogenic flux to the deep trap in the northern and central trap may be scavenged from the water column and only 40 per cent may be directly derived from the surface layers. In reality the process is more complicated with particles undergoing repeated aggregation and disaggregation. Still the above calculations show that scavenging processes are very important in settling of lithogenic particles in the oceans.

In the southern Bay of Bengal (*Ittekkot et al., 1991*) there is not much increase or decrease of lithogenic fluxes with depth suggesting a balance between the processes of aggregation and disaggregation. At both sites in the Arabian Sea the gradient of lithogenic flux is low due to the dominance of eolian sediments and reduced input of fluvial material in recent times (*Milliman et al., 1984*)

As shown above clay minerals are rapidly removed from the photic zone because of higher biological activity. This is an ecologically significant mechanism by which marine biota clear the photic zone of abiotic particles, which would otherwise hinder availability of light and nutrient. Sediment trap studies in the Mediterranean seas show that residence time of aluminosilicates in the photic zone range between 5 to 70 days with minimum residence times during periods of high biological activity (*Buat-Menard et al.,1989; Davies and Buat-Menard, 1990*). In deeper waters their residence times is about 15 years (*Spencer, 1984*). As scavenging of lithogenic particles in the deeper waters is not as efficient, advective processes below the photic zone can transport lithogenic particles far away from the point of introduction into the ocean. The horizontal flux of fine particles is then converted to vertical flux due to scavenging or other deep water processes of aggregation Conversion of horizontal flux to vertical flux is also seasonal and is mostly during high flux periods. It is therefore very much probable that most of the clays will settle out during periods of high productivity or in areas of high biological productivity. Higher Al mass accumulation rates have been recorded in the equatorial Pacific divergence zone compared to adjacent areas with low productivity (*Dymond et al., 1993*).

Advection of particles can also take place at intermediate depths below the photic zone. *Pak et al., (1980)* have reported intermediate particle maxima at depths of about 200 and 400 m extending hundreds of kilometers from the continental slope out into the open sea off the western South American coast. Advection of clays at intermediate depths may be important in our study area as very high lithogenic fluxes is seen even in the shallow southern Bay of Bengal traps (*Ittekkot et al., 1991*).

From the sediment trap flux data it is estimated that out of more than 2500 million tonnes of lithogenic matter discharged into the northern Indian Ocean, between 30 to 40 million tonnes of the lithogenic matter is being deposited in the deeper part of the ocean as hemi-pelagic sediments.

Of the 200 million tonnes of terrigenous material discharged into the Arabian Sea only 4 to 6.5 million tonnes reaches the deeper part of the northern Arabian Sea, the rest being trapped in estuaries, deltas and continental shelves. Similarly by extrapolating the lithogenic fluxes to the entire Bay, it is seen that out of more than 1500 million tonnes of lithogenic material discharged into the Bay of Bengal only 55 million tonnes is deposited in the sediment column. This is less than 5% of the total lithogenic input to the Bay of Bengal. In the Arabian Sea cross shelf transport is negligible and hemi-pelagic sediments are dominated by eolian sediments in the western and central Arabian Sea. In the Bay of Bengal cross shelf transport of fluvial sediments is considerable and the high fluxes observed in the deep Bay of

Bengal traps is due to scavenging of this advected material by biogenic aggregates. Other transport processes of lithogenic material like high density lutite flows, turbidity currents, transport through submarine channels etc is not yet well understood and needs to be further investigated.

5.3.4. Role of monsoons in marine sedimentation.

The effect of monsoons on marine sedimentation is shown in figures 5.16 and 5.17. In the Arabian Sea monsoon winds trigger phytoplankton blooms at the sea surface by bringing in nutrients from deeper waters. A part of the phytoplankton sinks out of the euphotic zone forming hemi-pelagic sediments. At the same time the monsoon winds also bring in large amounts of eolian dust. The simultaneous introduction of iron at the time of phytoplankton bloom leads to a large bloom than would otherwise occur. Furthermore dust particles can get incorporated into biogenic aggregates increasing their settling rates.

In the Bay of Bengal high rainfall during the monsoons leads to increase in the amount of freshwater and suspended solids introduced into the oceans. Nutrients introduced along with freshwater enhances primary productivity in the seas. Sedimentation of the particles produced is accelerated by the lithogenic material introduced along with the river discharge. Thus we see that monsoons are directly or indirectly not only influence particle production at the sea surface but also influence their rapid sedimentation.

In summary, the northern Indian Ocean has extremely high lithogenic fluxes due high input of terrigenous material into the area by fluvial and eolian processes. The quantity of lithogenic matter settling to the sea floor depends on monsoon circulation and distance from shore (river mouths). Lithogenic fluxes increase with depth due to scavenging, especially in areas dominated by fluvial sediments. Settling of lithogenic matter is controlled by availability

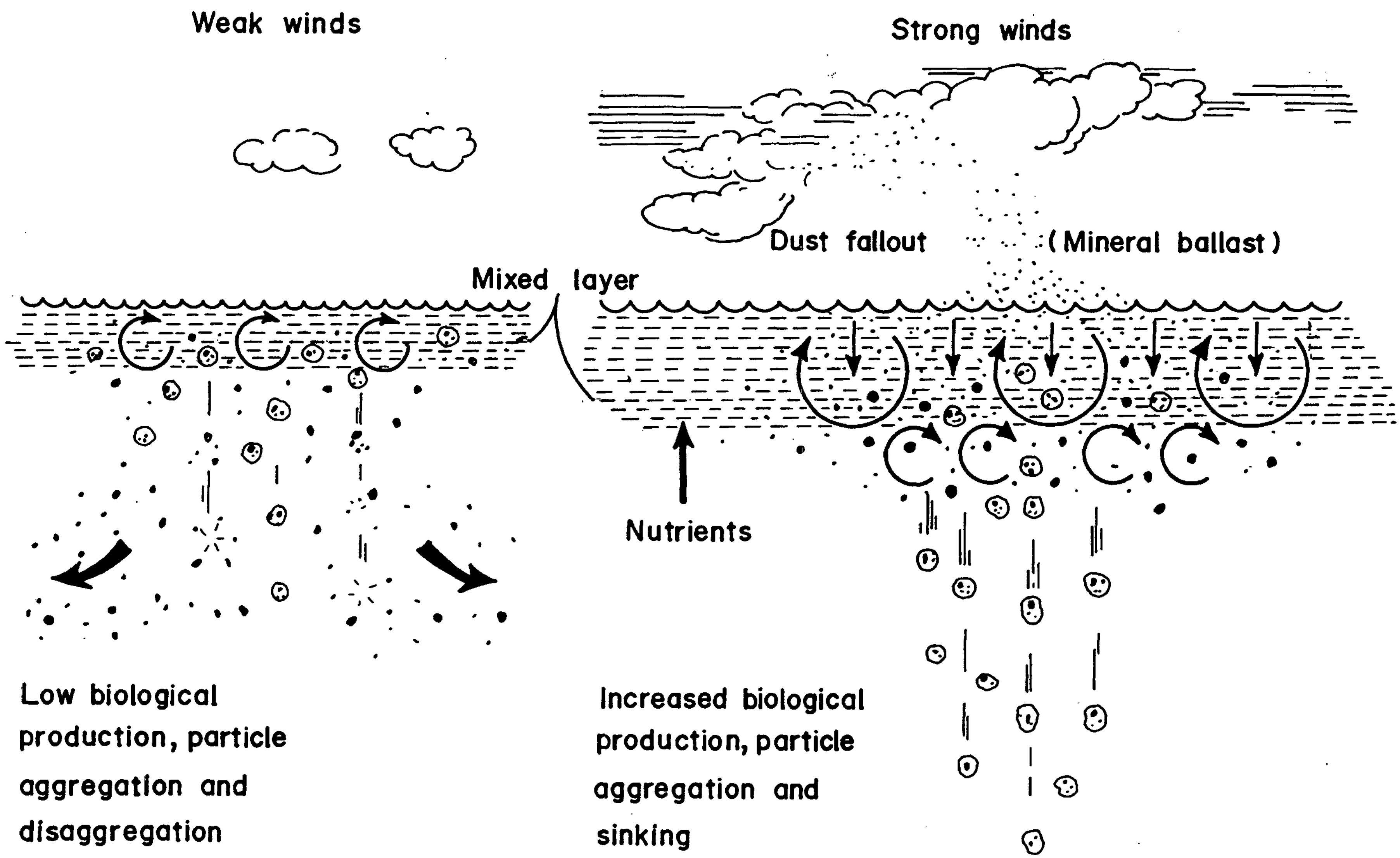


Fig. 5.16 Cartoon showing effect of monsoon winds on marine sedimentation. Monsoon winds deepen the mixed layer introducing nutrients into the photic zone. The monsoon winds also bring in eolian dust from the Arab-Somali region which increases the density of organic aggregates thereby enhancing their sinking rates.

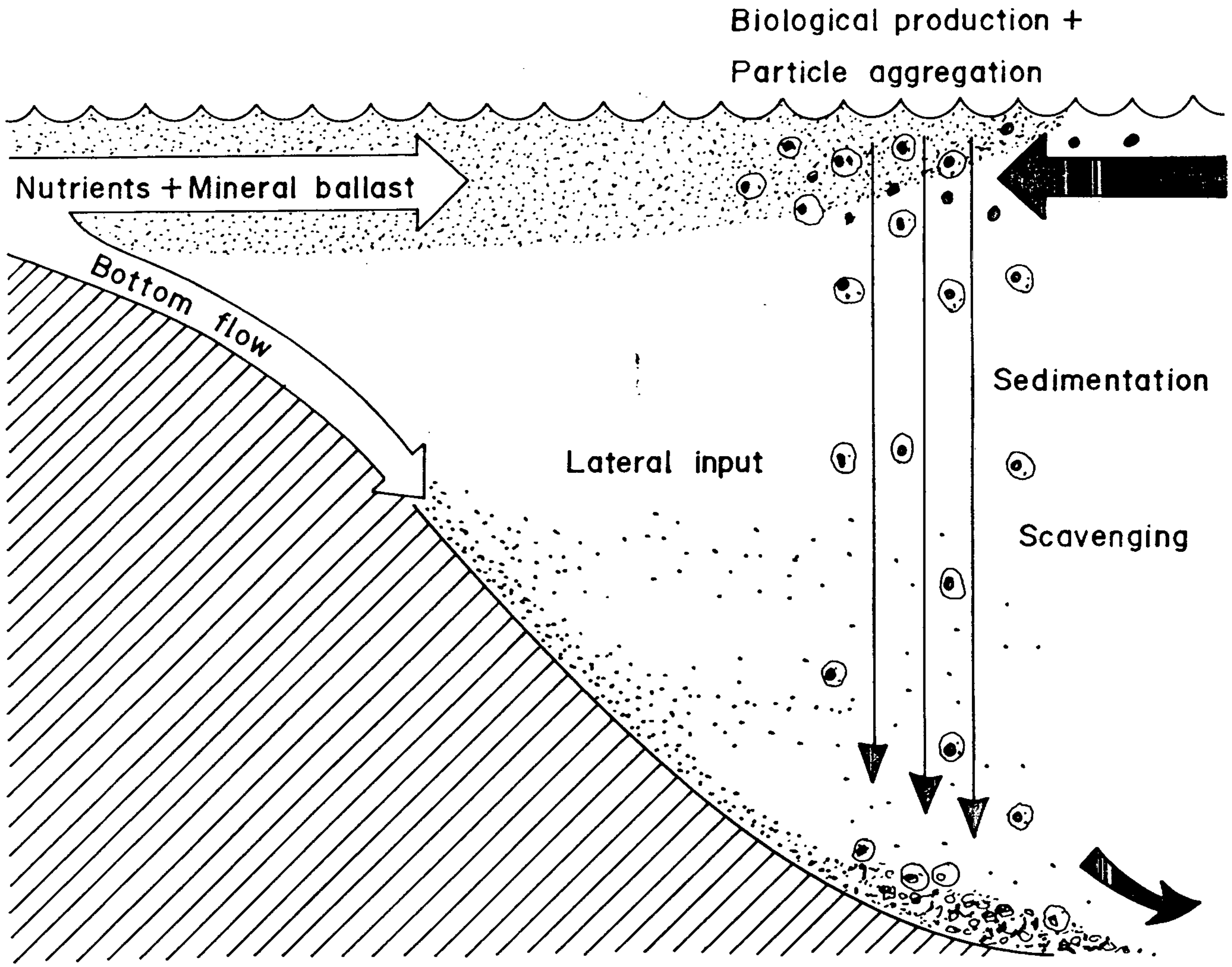


Fig. 5.17 Figure showing effect of river discharge on marine sedimentation. Nutrient supplied by rivers increases primary production and aggregate formation in river plumes. The lower salinity enhances stratification of surface waters and keeps the nutrients in the photic zone for a longer period. Biogenic aggregates incorporate lithogenic particles supplied by rivers leading to rapid sedimentation of both types of particles. In addition the biogenic aggregates also scavenge lithogenic particles advected from continental shelves.

of biogenic aggregates which have higher settling rates. This biological control on lithogenic particle sedimentation is clearly seen at all locations in the northern Indian Ocean.

Chapter 6

Mineralogy of the Lithogenic Fraction of Sediment Trap Samples Collected in the Arabian Sea and Bay of Bengal.

6.1. Introduction

The lithogenic fraction of marine sediments is derived from land and comprises the erosional products of continental rocks. More than 90 per cent of the lithogenic particles are introduced into the oceans near the boundaries, the largest input being at river mouths. Once in the oceans, lithogenic material undergo very little change, therefore the mineralogy of surficial marine and adjacent land sediments are very similar.

The most widely distributed terrigenous minerals in the oceans are clays. Clays are hydrous silicates of aluminium and are derived from the weathering or alteration products of rocks. The major clay minerals found in marine sediments are illite, kaolinite, chlorite and smectites. Quartz, fledspars, manganese and iron oxide are the principal non-clay minerals. The relative proportions of clays will vary according to the prevailing climatic regime in the source region and to mixing process in the oceans.

There has been a lot of debate regarding the formation of clays in the ocean. It is now generally accepted that they are formed by weathering of rocks on land and transported to the oceans with little or no subsequent alteration. Near mid-ocean ridges and in areas where volcanic rocks are predominant, smectites can form due to submarine alteration of oceanic basalts.

There is a latitudinal zonation of clay minerals in the oceans with kaolinites being dominant at low latitudes and illite and chlorite being dominant at high latitudes. This is because chlorite and illite are unstable clay minerals formed due to mechanical weathering rather than chemical alteration. The acidic, warm and humid conditions of the tropics enhance the alteration of chlorite to kaolinite.

Not much work has been carried out on the mineralogy of the lithogenic fraction of sediment trap samples. *Honjo et al., (1992b)* have shown that there is a deep advective transport of smectites in the Panama Basin. *Aoki et al. (1992)* have shown that in the Japan trench at a depth of 8700 m there is hardly any change in clay mineral composition with depth. The main source of clays to the Japan Trench are eolian dust and resupended turbidites.

There are three main sources of lithogenic material to the northern Indian Ocean each having a characteristic mineral assemblage. The most important source is the rivers draining the Himalayas which have high percentages of illite and chlorite and very low kaolinite and smectite (*Naidu et al., 1985*). The sediments of the Peninsular Indian rivers have high smectite and kaolinite content but are low in illite and chlorite (*Naidu et al., 1985*). Eolian dust from north Africa and Arabia have high illite, palygorskite and quartz percentages (*Sirocko, 1989*). From the clay mineral assemblage of the sediment trap samples an attempt has been made to trace the provenance of the lithogenic fraction.

6.2. Material and Methods

The samples used for study of mineralogy of the lithogenic particles were collected between May 1986 and May 1987 in the Arabian Sea and between November 1988 and November 1989 in the Bay of Bengal (See table 3.2 for details).

McManus (1991) has suggested that all authors reporting clay mineral percentages by X-Ray diffraction must give full details of samples preparation as quantification of clay minerals is highly dependant on the method of sample preparation and operator skills. He cautions that the results though repeatable and precise may not be accurate.

The clay mineral percentages reported in this thesis are only semi-quantitatve. Since all the samples had similar treatment and preparation, the seasonal and spatial variations can be compared. Minor variation in mineral percentages are not compared and only major trends are discussed.

For clay mineral analyses, a 1/8 aliquot of the sample was treated with hydrogen peroxide and dilute acetic acid to remove organic carbon and calcium carbonate. The material remaining after removal of calcium carbonate and organic matter was made into a thick slurry on a glass slide and run on a Phillips XRD unit. It was noticed that in some of the Arabian Sea samples, especially those collected during the SW monsoon period, biogenic silica masked the clay mineral signals. This was corrected in the Bay of Bengal samples by removing biogenic silica with a 1M solution of sodium carbonate heated to 90 ^{o}C for two hours.

No sample disaggregation or grinding was required as the samples were not kept in a dried state at any time. However, prior to slide preparation the samples were disaggregated thoroughly by treatment in an ultrasonic bath for 2 to 5 minutes.

The entire sample was analysed, and no particle size separation prior to analysis was carried out. The samples are composed mostly of clays with minor quantities of fine silt (see Chapter 8).

The samples were run in a Philip X-Ray diffraction unit at from 3 ^{o} to 40^{o} 2 theta at the rate of 2^{o} 2 theta per minute using 40 kV and 20 mA settings. Samples suspected to have significant amounts of smectite were glycolated at 150^{o} C for one hour and rerun using the same settings.

For mineral identification the following peaks in the diffractogram were taken (Fig. 6.0).

Smectites	15 A°
Mixed layer clays	14 A°
Palygorskite	10.16A°
Illite-	9.9 A°
Kaolinite and Chlorite	7.14A°
Quartz	4.26A°
Plagioclase	3.2 A°

Semi-quantitative determination of mineral percentages were made using the method of *Biscaye* (*1965*). The peak area of each mineral measured above an appropriate baseline was multiplied by factors given by Biscaye as follows:

Illite	4
Smectite	1
Kaolinite + chlorite	2.5

No factors were applied for the non-clay minerals. The fluxes of all minerals were calculated by assuming that clay minerals, quartz and plagioclase made up 100% of the total lithogenic flux. Kaolinite and chlorite ratios were determined by resolving the combined 25° 2 theta peak at a slow scan. This ratio was used to calculate the actual mineral percentage of these two minerals, the combined percentage having been previously calculated using the 12.5° 2 theta peak.

The accuracy of the X-Ray diffraction method was checked using external standards prepared using pure clay minerals. Pure mica, vermiculite, kaolinite, Wyoming bentonite, chlorite, quartz and orthoclase were ground to a fine powder, mixed in known proportions and run on the XRD. The results

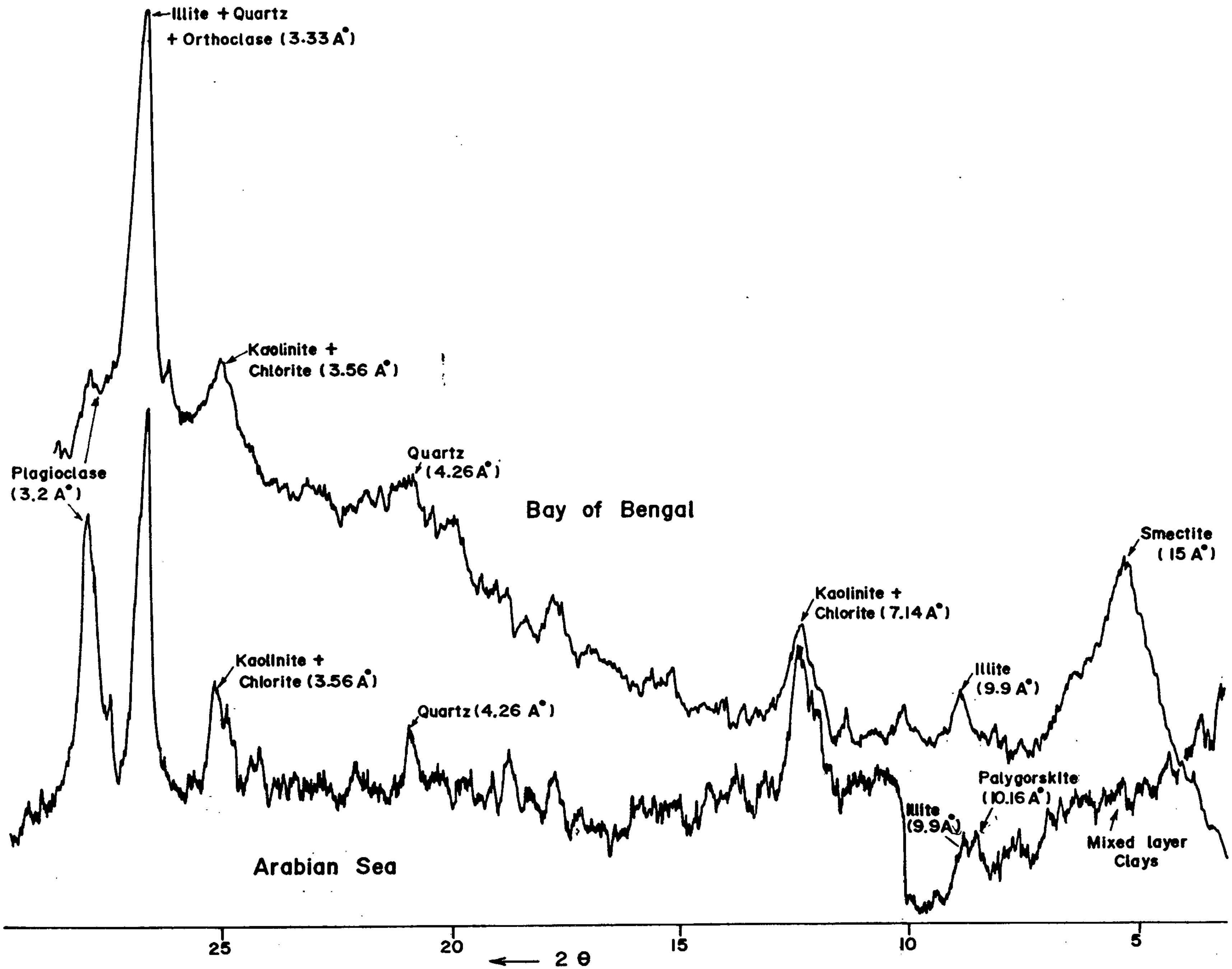


Fig. 6.D Sample X-ray diffractogram of the lithogenic fraction of sediment trap samples from the Arabian Sea and Bay of Bengal. The major peaks used for mineral identification and quantification are shown.

though not very accurate (< 5%) are well within the acceptable range of semi-quantitative analysis.

6.3. Results and discussions

6.3.1. Mineralogy of the lithogenic fractions of the Arabian Sea sediment trap samples

X-ray diffraction studies of the lithogenic fraction of the trap samples in the Arabian Sea show illite to be the dominant mineral which range from 25 to 50% (Table 6.1 to 6.3 & Fig. 6.1). The other minerals identified are quartz, feldspar, smectites, palygorskite and chlorite. All mineral fluxes follow a similar pattern with peak fluxes during the SW monsoon.

Smectites. Smectite percentages are generally less than 10% of the total lithogenic flux but higher percentages up to 24% were found in the eastern and central Arabian Sea during the SW monsoons. They are negligible in the western Arabian Sea amounting to less than 3% of the annual lithogenic fluxes. The estimated annual fluxes of smectites are 0.42, 0.31 and 0.07 g m^{-2} y^{-1} for the eastern, central and western Arabian sea sites respectively. Smectites are derived from the weathering of Deccan trap basalts and are transported to the Arabian Sea mostly through the rivers Narmada and Tapti. The suspended load of these rivers contain more than 75% smectites (*Naidu et al., 1985*). About 80% of the rainfall in the drainage basins of these rivers is during the SW monsoon period, and this is seen in the higher fluxes of smectites on the eastern and central traps during this period.

Illite/smectite ratios in the eastern Arabian Sea (Fig. 6.2) drop steadily from about 20 in July to less than 4 in September due to increase in the flux of smectites. The current during this period flow towards the SE (*West coast of India Pilot, 1975*)) and is favorable for resuspending and transporting smectite rich sediments from the eastern continental margins of India.

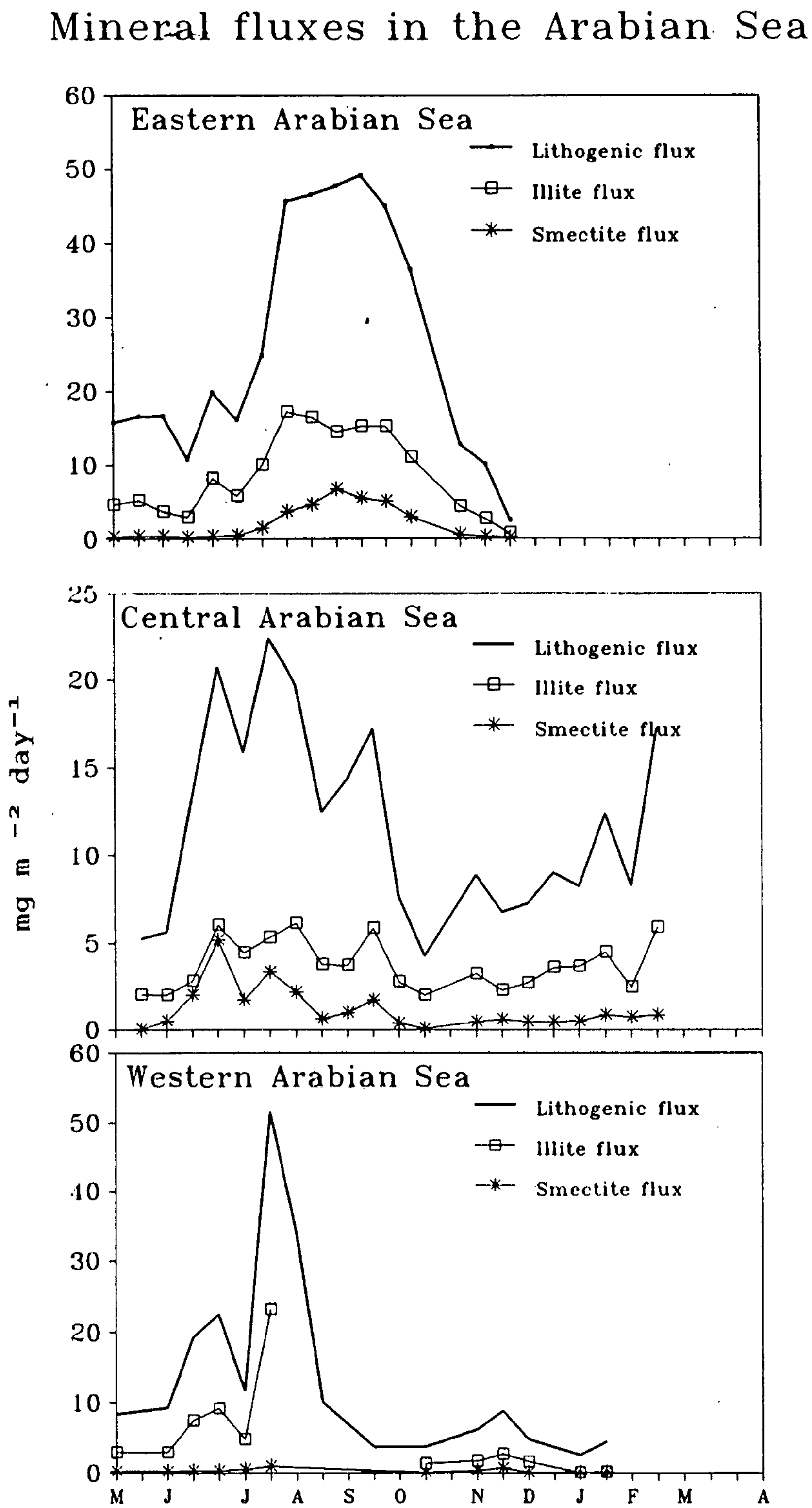


Fig. 6.1. Time-series data on total lithogenic, illite and smectite fluxes in the eastern, central and western Arabian Sea for the period May 1986 to April 1987..

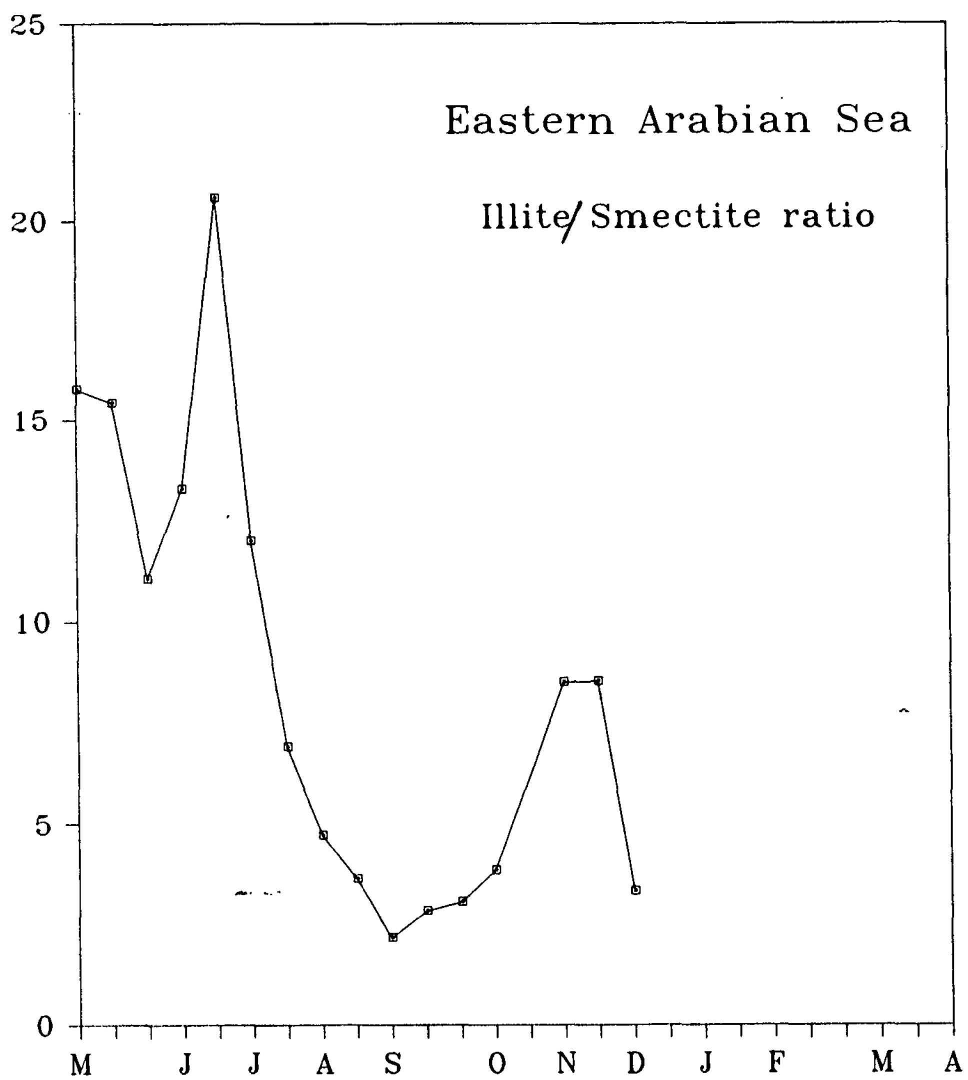


Fig. 6.2 Illite/Smectite ratio in settling particles from the eastern Arabian Sea. Note the steady fall in the ratio from June to September.

Table 6.1. Clay mineral fluxes to the eastern Arabian Sea. All fluxes are in mg m-2 day-1.
K= Kaolinite; Ch= Chlorite; MLC= Mixed layer clays.

	TOTAL LITH.	ILLITE FLUX	%	PALYGORSKITE FLUX	%	K+Ch FLUX	%	MLC FLUX	%	SMECTITE FLUX	%	FELDSPAR FLUX	%	QUARTZ FLUX	%
10 May-23 May	15.8	4.6	29.0	1.1	7.0	3.2	20.0	1.0	6.1	0.3	1.8	1.4	9.0	4.3	27.0
23 May-05 Jun	16.7	5.2	31.0	1.5	9.2	2.8	16.5	0.9	5.1	0.3	2.0	1.5	9.2	4.5	26.8
05 Jun-18 Jun	16.7	3.7	22.0	1.2	7.0	2.7	16.0	1.3	8.0	0.3	2.0	2.2	13.0	5.3	32.0
18 Jun-01 Jul	10.8	2.9	27.0	0.6	6.0	1.5	14.0	0.9	8.0	0.2	2.0	1.2	11.0	3.4	32.0
01 Jul-14 Jul	19.9	8.1	41.0	1.0	5.0	3.2	16.0	0.8	4.0	0.4	2.0	3.0	15.0	3.4	17.0
14 Jul-27 Jul	16.1	5.8	36.0	0.8	5.0	2.9	18.0	1.0	6.0	0.5	3.0	1.9	12.0	3.2	20.0
27 Jul-09 Aug	24.8	10.0	40.4	3.2	13.1	4.1	16.5	1.7	6.8	1.4	5.8	1.6	6.7	2.7	10.7
09 Aug-22 Aug	45.7	17.2	37.6	2.5	5.5	8.2	18.0	2.3	5.0	3.6	8.0	5.0	11.0	6.9	15.0
22 Aug-04 Sep	46.5	16.5	35.5	2.4	5.1	6.2	13.3	2.7	5.9	4.6	9.8	3.2	6.9	11.0	23.6
04 Sep-17 Sep	47.7	14.4	30.2	4.7	9.8	7.6	16.0	3.8	8.0	6.7	14.0	1.9	4.0	8.6	18.0
17 Sep-30 Sep	49.1	15.2	31.0	3.4	7.0	8.4	17.0	4.9	10.0	5.4	11.0	3.4	7.0	8.4	17.0
30 Sep-13 Oct	45.1	15.2	33.8	2.6	5.8	6.4	14.1	4.5	10.1	5.0	11.1	3.2	7.0	8.2	18.2
13 Oct-26 Oct	36.5	11.0	30.1	2.2	6.1	4.3	11.8	3.6	9.8	2.9	7.9	3.6	9.8	9.0	24.5
19 Nov-01 Dec	12.8	4.3	34.0	0.0	0.0	2.6	20.1	0.8	6.0	0.5	4.0	1.9	15.0	2.7	21.0
01 Dec-14 Dec	10.0	2.6	26.0	0.6	6.0	2.6	26.0	0.6	6.0	0.3	3.0	0.6	6.0	2.7	27.0
14 Dec-26 Dec	2.4	0.7	30.0	0.0	0.0	0.3	13.9	0.2	8.0	0.2	9.1	0.2	9.1	0.7	30.0
26 Dec-08 Jan	*														
08 Jan-21 Jan	*														
21 Jan-02 Feb	*														
02 Feb-15 Feb	*														
15 Feb-27 Feb	*														
27 Feb-12 Mar	*														
12 Mar-25 Mar	*														
25 Mar-06 Apr	*														
06 Apr-19 Apr	*														
19 Apr-01 May	*														
Annual Flux	5400	1791	33.2	362	6.7	867	16.1	395	7.3	419	7.8	467	8.7	1096	20.3

Table 6.2. Clay mineral fluxes to the central Arabian Sea. All fluxes are in mg m-2 day-1.
K= Kaolinite; Ch= Chlorite; MLC= Mixed layer clays.

	TOTAL	ILLITE		PALYGORSKITE		K+Ch		MLC		SMECTITE		FELDSPAR		QUARTZ	
	LITH.	FLUX	%	FLUX	%	FLUX	%	FLUX	%	FLUX	%	FLUX	%	FLUX	%
10 May-23 May	*														
23 May-05 Jun	5.3	2.0	38.6	0.4	7.0	1.2	23.8	0.2	2.9	0.1	1.0	0.5	8.9	0.9	17.8
05 Jun-18 Jun	5.6	2.0	36.1	0.2	2.9	1.1	19.0	0.3	5.0	0.5	9.0	0.5	9.0	1.1	19.0
18 Jun-01 Jul	13.4	2.8	21.0	0.9	7.0	3.1	23.0	0.8	6.0	2.0	15.0	1.3	10.0	2.4	18.0
01 Jul-14 Jul	20.8	6.0	29.0	0.0	0.0	3.7	18.0	2.7	13.0	5.2	25.0	1.2	6.0	1.9	9.0
14 Jul-27 Jul	15.9	4.4	28.0	1.4	9.0	2.7	17.0	1.0	6.0	1.7	11.0	1.4	9.0	3.2	20.0
27 Jul-09 Aug	22.4	5.3	23.8	1.1	5.0	3.8	16.8	0.9	4.0	3.3	14.9	2.0	8.9	6.0	26.7
09 Aug-22 Aug	19.8	6.1	31.0	2.4	12.0	3.2	16.0	1.0	5.0	2.2	11.0	1.2	6.0	3.8	19.0
22 Aug-04 Sep	12.5	3.8	30.1		0.0	2.5	20.0	0.9	7.0	0.6	5.0	2.0	16.0	2.7	22.0
04 Sep-17 Sep	14.4	3.7	26.0	0.6	4.0	2.9	20.0	0.6	4.0	1.0	7.0	1.1	8.0	4.5	31.0
17 Sep-30 Sep	17.2	5.8	33.9		0.0	2.9	16.9	0.9	5.2	1.7	10.0	2.9	16.9	2.9	16.9
30 Sep-13 Oct	7.7	2.8	36.3		0.0	2.1	27.3	0.2	3.0	0.4	5.1	1.2	16.2	0.9	12.1
13 Oct-26 Oct	4.3	2.0	47.0		0.0	0.7	17.1	0.1	2.9	0.1	2.0	0.5	11.0	0.9	20.0
19 Nov-01 Dec	8.9	3.2	36.4	1.0	11.1	1.8	20.3	0.7	8.1	0.4	5.1	0.6	7.0	1.1	12.1
01 Dec-14 Dec	6.8	2.3	33.7		0.0	1.4	20.9	0.8	11.8	0.6	9.0	0.5	6.8	1.2	17.8
14 Dec-26 Dec	7.3	2.7	37.4		0.0	1.4	19.6	0.5	6.6	0.5	6.6	0.8	11.2	1.4	18.6
26 Dec-08 Jan	9.0	3.6	40.0	0.5	6.0	1.9	21.0	0.5	6.0	0.5	5.0	0.5	6.0	1.4	16.0
08 Jan-21 Jan	8.2	3.7	44.4		0.0	1.4	17.2	0.3	3.9	0.5	6.1	0.5	6.1	1.8	22.2
21 Jan-02 Feb	12.4	4.5	36.0	1.5	12.0	1.6	13.0	0.7	6.0	0.9	7.0	0.7	6.0	2.5	20.0
02 Feb-15 Feb	8.3	2.5	30.0	0.8	9.9	1.2	15.0	0.7	9.0	0.7	9.0	0.8	9.9	1.4	17.1
15 Feb-27 Feb	17.3	5.9	34.1	2.1	12.0	3.5	20.1	1.0	5.8	0.9	5.0	1.2	7.0	2.8	16.1
27 Feb-12 Mar	*														
12 Mar-25 Mar	*														
25 Mar-06 Apr	*														
06 Apr-19 Apr	*														
19 Apr-01 May	*														
ANNUAL	3050	965	31.7	158	5.2	567	18.6	189	6.2	310	10.2	280	9.2	577	18.9

Table 6.3. Clay mineral fluxes to the western Arabian Sea. All fluxes are in mg m-2 day-1.
K= Kaolinite; Ch= Chlorite; MLC= Mixed layer clays.

	TOTAL LITH.	ILLITE FLUX	%	PALYGORSKITE FLUX	%	K+Ch FLUX	%	MLC FLUX	%	SMECTITE FLUX	%	FELDSPAR FLUX	%	QUARTZ FLUX	%
10 May-23 May	8.3	2.9	34.8	0.6	7.1	1.6	19.4	0.3	4.0	0.3	3.0	1.0	12.2	1.6	19.4
23 May-05 Jun	*														
05 Jun-18 Jun	9.2	2.9	31.4	1.3	14.2	1.9	20.2	0.7	8.0	0.3	3.0	0.6	7.1	1.5	16.2
18 Jun-01 Jul	19.3	7.4	38.1	1.1	5.7	3.3	17.1	2.2	11.4	0.4	1.9	2.2	11.4	2.8	14.3
01 Jul-14 Jul	22.5	9.1	40.4	2.3	10.1	2.7	12.1	1.6	7.1	0.4	2.0	1.4	6.1	5.0	22.2
14 Jul-27 Jul	11.7	4.8	41.0	0.5	4.0	1.3	11.0	1.3	10.9	0.5	4.0	1.2	10.0	2.2	19.1
27 Jul-09 Aug	51.5	23.2	45.0	2.6	5.0	7.7	15.0	2.6	5.0	1.0	2.0	4.1	8.0	10.3	20.0
09 Aug-22 Aug	33.8	*Mineral percentages not determined due to masking by excess biogenic silica*													
22 Aug-04 Sep	10.0														
04 Sep-17 Sep	*														
17 Sep-30 Sep	3.7														
30 Sep-13 Oct	3.7														
13 Oct-26 Oct	3.7	1.3	36.4	0.4	11.1	0.6	17.1	0.3	8.0	0.1	2.1	0.3	7.2	0.7	18.1
19 Nov-01 Dec	6.1	1.7	28.0	0.8	12.8	1.3	20.7	0.4	6.3	0.4	6.2	0.7	11.4	0.9	14.6
01 Dec-14 Dec	8.8	2.7	30.4	0.4	4.0	1.2	14.1	0.7	7.9	0.7	8.1	0.9	10.1	2.2	25.4
14 Dec-26 Dec	4.8	1.6	32.2	0.1	2.6	0.9	18.8	0.2	4.4	0.1	1.7	0.8	16.0	1.2	24.2
26 Dec-08 Jan															
08 Jan-21 Jan	2.6	1.1	40.7	0.3	9.9	0.3	12.3	0.2	7.0	0.1	2.0	0.2	6.0	0.6	22.2
21 Jan-02 Feb	4.4	1.8	41.2	0.2	3.9	0.5	11.0	0.5	11.0	0.2	3.9	0.4	10.1	0.8	19.0
02 Feb-15 Feb	*														
15 Feb-27 Feb	*														
27 Feb-12 Mar	*														
12 Mar-25 Mar	*														
25 Mar-06 Apr	*														
06 Apr-19 Apr	*														
19 Apr-01 May	*														
ANNUAL	2640	1050	39.8	175	6.6	402	15.2	186	7.1	73	2.8	237	9.0	514	19.5

Taking a higher estimate of 0.43 g m^{-2} y^{-1} as the average annual flux of smectites for the Arabian Sea it is computed that the total input to the deep Arabian Sea north the Carlsberg Ridge excluding the shelf areas is less than 0.6 million tonnes per year. This is less than 1% of the annual discharge of smectites from the Narmada and Tapti rivers. The balance of smectites is probably retained in the inner shelf of the west coast of India.

Illite: Illite is the major mineral in all the traps throughout the year and is derived from the Indus river which has up to 50% illite (*Konta, 1985*) as well as through eolian dust having around 45% illite (*Chester, 1985*). The annual illite fluxes to the eastern, central and western trap are 1.79, 0.96 and 1.05 g m^{-2} day^{-1} respectively. Illite percentages are maximum in the western trap where it accounts for 40% of the total lithogenic flux. In the eastern Arabian Sea illite fluxes decrease from about 40% in the beginning of the monsoons to approximately 30% towards the end of the monsoon. We suspect that the eastern Arabian Sea trap may have illite derived mostly from the Indus while the western Arabian Sea trap may have higher eolian illite, but the illite from these two sources cannot be differentiated on the basis of our data.

Palygorskite: Palygorskite is an important indicator of eolian transport and is inferred to have been derived from the Somali-Arabian area (*Goldberg and Griffin, 1970; Kolla et al.,1981b; and Weser, 1974*). Concentrations of palygorskite in the sediment trap samples show that eolian material derived from Arab-Somali regions cover the entire Arabian Sea. Even the eastern Arabian Sea traps received palygorskite transported to this site by monsoon winds. Compared to bottom sediments, palygorskite content in the eastern Arabian Sea sediment trap samples are much higher. This may be mainly due to the present reduced input of illite to the eastern Arabian Sea. The main source of illite to this region, the Indus river, delivers less than 50 million tonnes of suspended sediments compared to more than 400 million tonnes before construtions of major dams and barrages (*Milliman et al., 1983*)

Quartz and feldspar: Quartz and feldspar contribute about 20% and 8% of the total lithogenic flux. Their distribution pattern is similar to that of illite. The annual average clay to framework silicate ratio is 2.5 to 1. In the eastern Arabian Sea, prior to the onset of the monsoon, the ratio is approximately 1.8 but towards the end of the monsoon this ratio increases to 2.8 because of increased fluvial input.

6.3.2. Mineralogy of the lithogenic fractions of the Bay of Bengal sediment trap samples.

X-Ray diffraction studies of the lithogenic fraction show illite to be the major mineral in all the traps (Table 6.4 and 6.5). Other minerals identified are smectites, chlorite, kaolinite quartz and plagioclase. Mineralogy of the lithogenic fraction show significant variations due to varying river inputs and change in surface and subsurface circulation.

Illite: Illite is derived mainly from the Ganges-Brahmaputra river as the suspended load of these rivers contain more than 65% illite (*Konta, 1985 Naidu et al., 1985*). Concentrations of illite in Peninsular Indian rivers is generally less than 25% and less than 5% in the Godavari and Krishna. As a consequence, illite flux shows a decreasing trend from north to south (Table 6.4 and 6.5). As expected, illite flux peaks during the SW monsoon in all the traps (Fig. 6.3 to 6.7). Also, their fluxes increase earlier in the northern trap compared to the central traps as they are located closer to the source.

Smectite: Smectite fluxes are higher in the central traps (Table 6.4 and 6.5; Fig. 6.3 to 6.7) as the major input of smectite to the Bay is from the nearby Krishna and Godavari rivers. Though they are ultimately derived from the

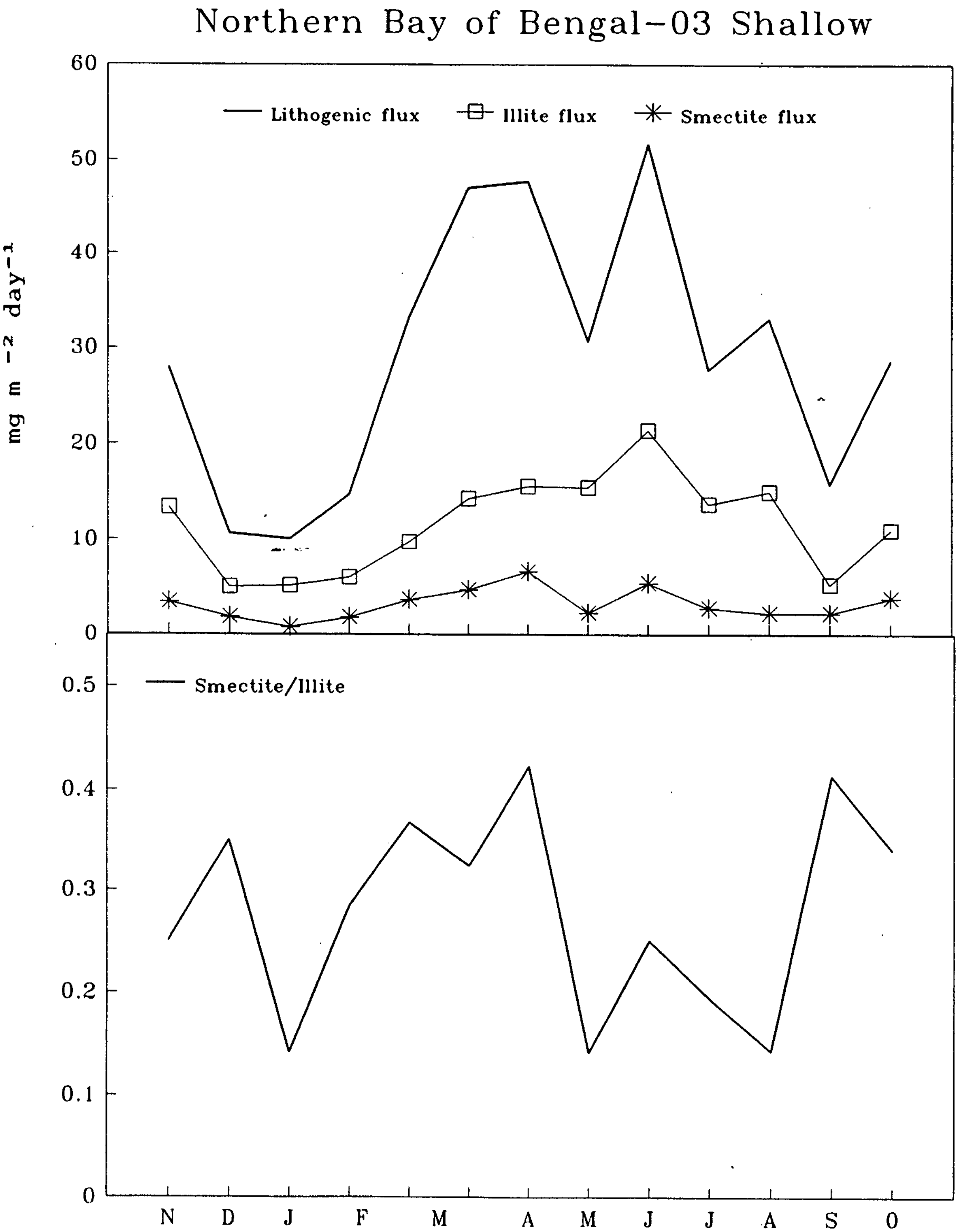


Fig. 6.3. Total lithogenic, illite and smectite flux and smectite/illite ratio in the shallow northern Bay of Bengal trap samples for the period November 1988 to October 1989.

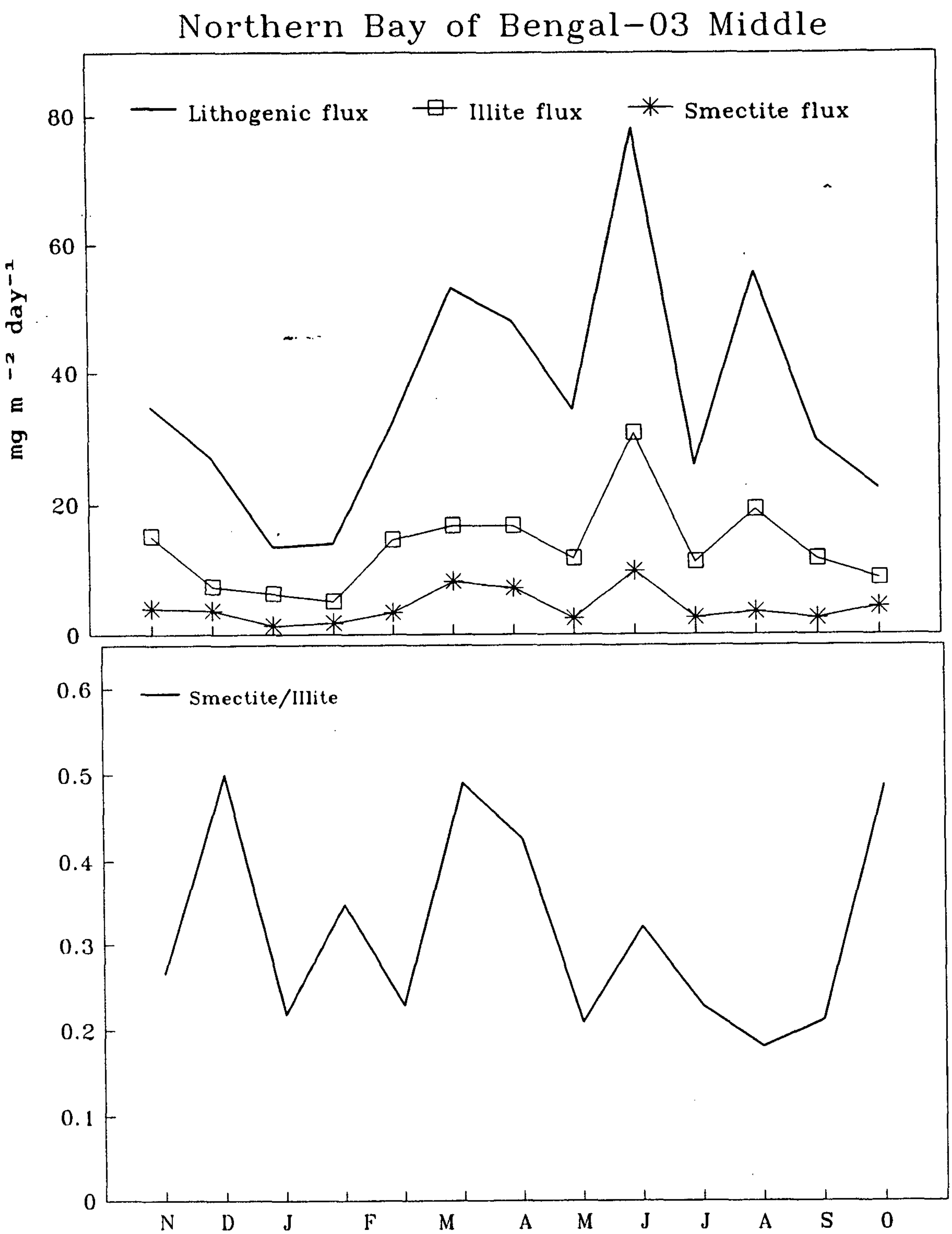


Fig. 6.4 Total lithogenic, illite and smectite flux and smectite/illite ratio in the middle northern Bay of Bengal trap samples for the period November 1988 to October 1989.

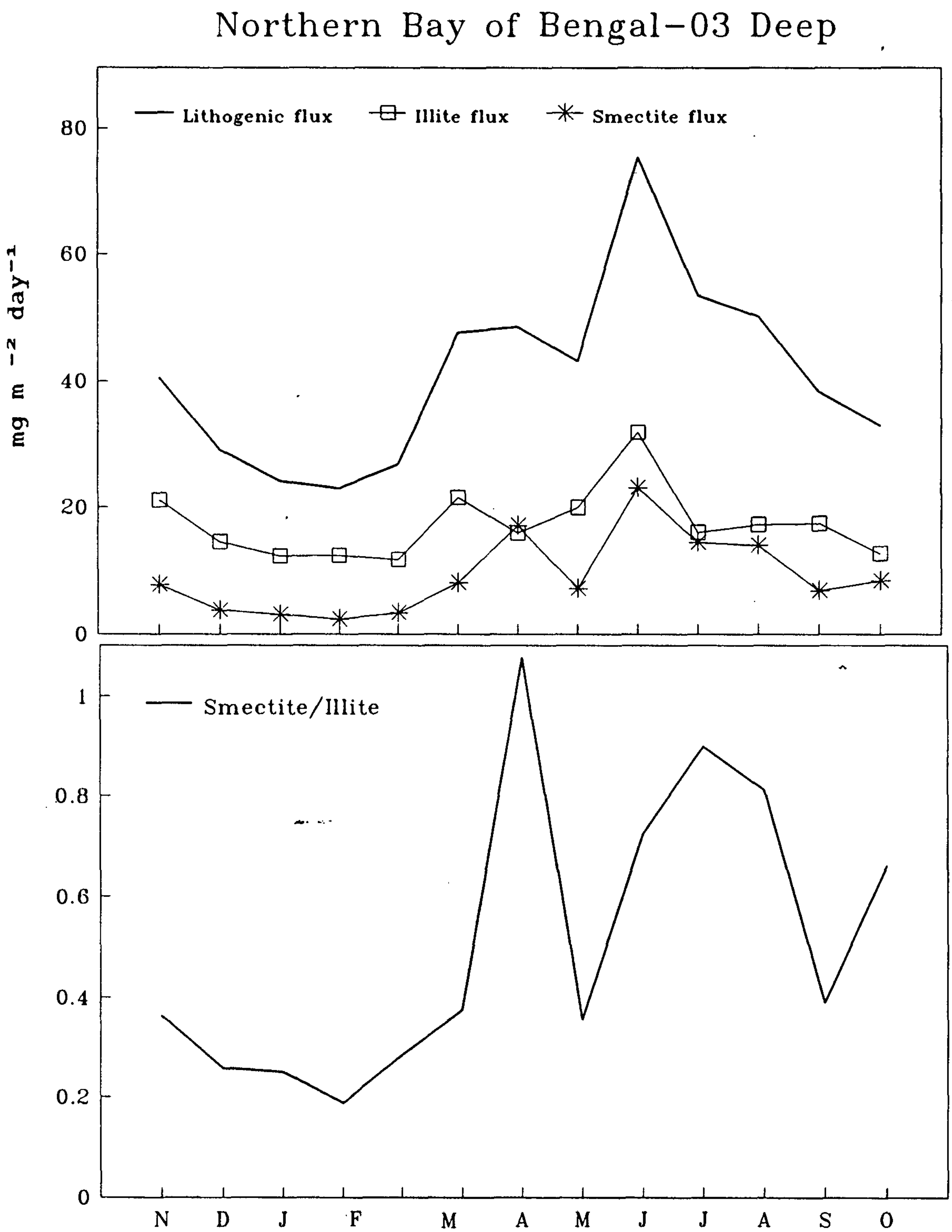


Fig. 6.5 Total lithogenic, illite and smectite flux and smectite/illite ratio in the deep northern Bay of Bengal trap samples for the period November 1988 to October 1989.

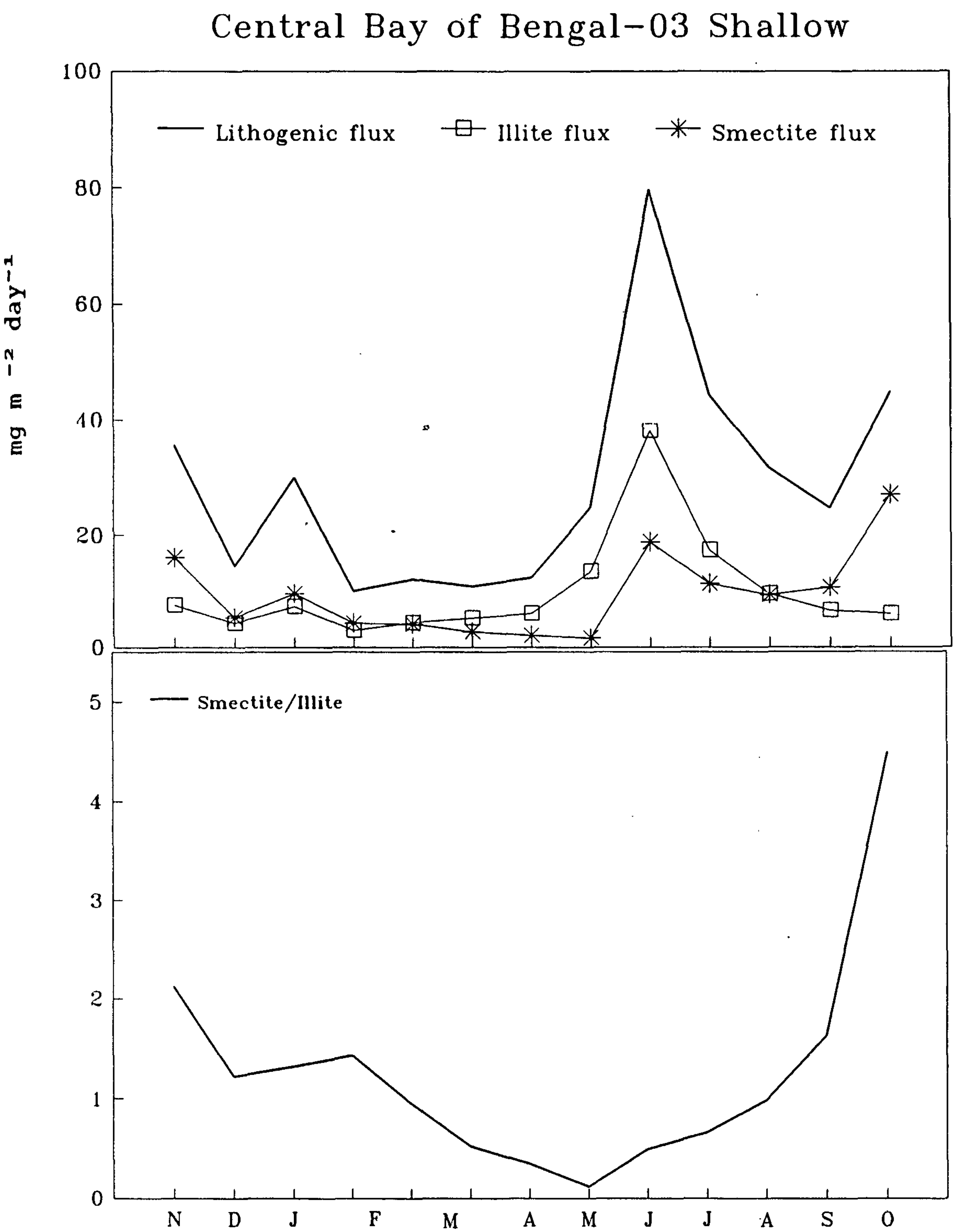


Fig.6.6 Total lithogenic, illite and smectite flux and smectite/illite ratio in the shallow central Bay of Bengal trap samples for the period November 1988 to October 1989.

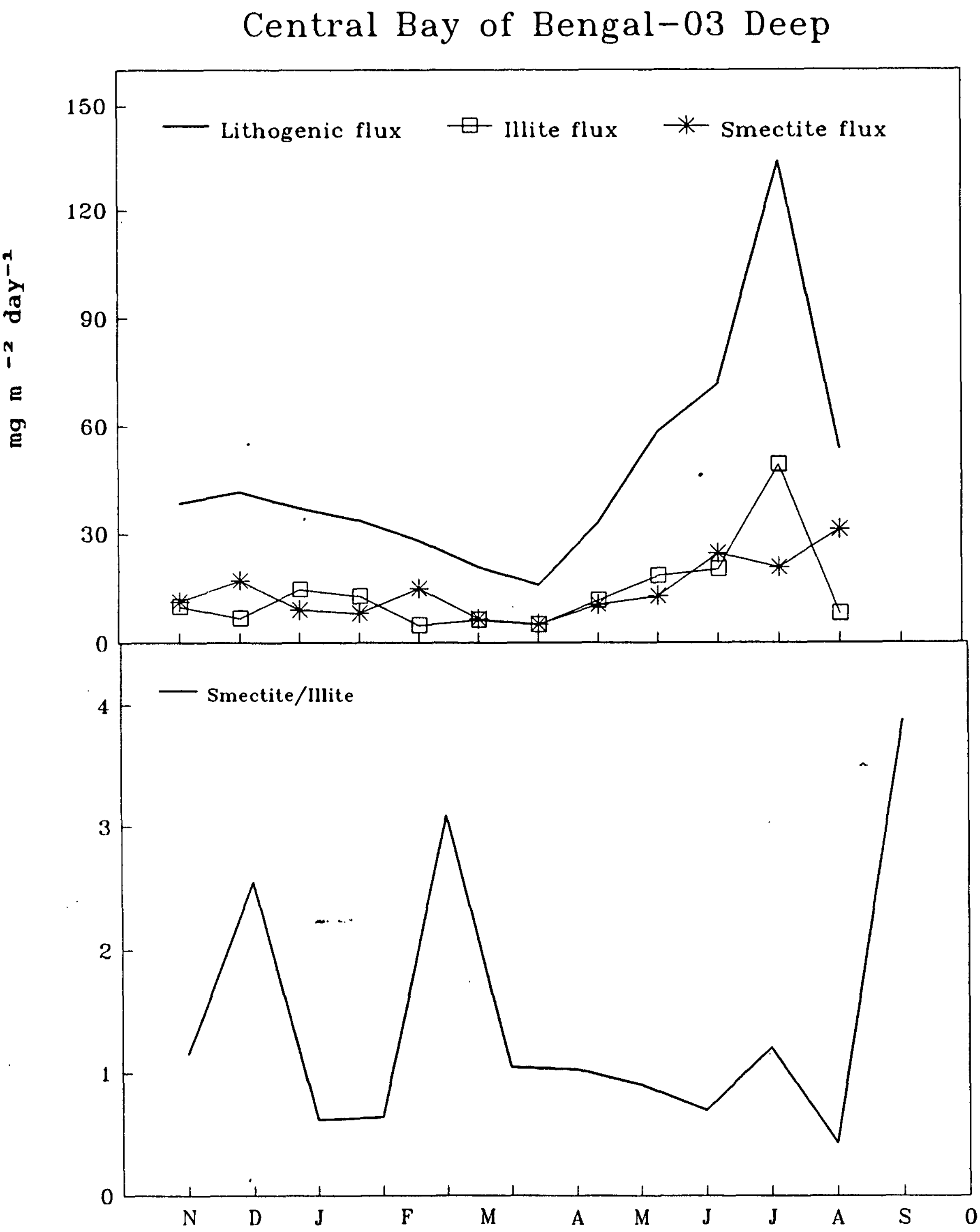


Fig. 6.7. Total lithogenic, illite and smectite flux and smectite/illite ratio in the deep central Bay of Bengal trap samples for the period November 1988 to October 1989.

Table 6.4. Clay mineral fluxes to the northern Bay of Bengal. All fluxes are in mg m-2 day-1.

North Bay of Bengal- 03 Shallow

	Total flux <1mm	Lith. flux	Smectite flux	Illite flux	Chlorite flux	Kaolinite flux	Quartz flux	Plagioclase flux
02 Nov-29 Nov	82.57	27.91	3.4	13.4	3.5	2.3	2.6	2.8
29 Nov-26 Dec	109.77	10.68	1.7	5.0	0.6	0.9	1.0	1.4
26 Dec-22 Jan	148.37	10.06	0.7	5.1	1.0	0.9	1.7	0.7
22 Jan-18 Feb	109.33	14.78	1.7	6.0	2.0	1.3	2.4	1.4
18 Feb-17 Mar	121.80	33.21	3.6	9.7	5.3	3.0	7.1	4.6
17 Mar-13 Apr	190.88	47.03	4.6	14.3	9.8	4.5	8.3	5.4
13 Apr-10 May	188.57	47.71	6.6	15.6	6.2	5.6	8.0	5.8
10 May-06 Jun	143.65	30.72	2.2	15.5	2.6	2.3	5.0	3.1
06 Jun-03 Jul	151.98	51.77	5.4	21.4	7.4	4.6	5.9	7.1
03 Jul-30 Jul	98.52	27.73	2.8	14.2	2.8	2.5	2.6	3.0
30 Jul-26 Aug	121.63	33.11	2.1	15.0	4.9	2.5	4.1	4.5
26 Aug-22 Sep	69.70	15.82	2.1	5.2	2.6	1.6	2.0	2.4
22 Sep-19 Oct	88.99	28.65	3.8	11.3	4.4	2.1	3.1	4.0
Annual (g m^{-2})	45.65	10.65	1.14	4.26	1.50	0.96	1.51	1.3

North Bay of Bengal- 03 Middle

	Total flux <1mm	Lith. flux	Smectite flux	Illite flux	Chlorite flux	Kaolinite flux	Quartz flux	Plagioclase flux
02 Nov-29 Nov	103.68	34.75	4.0	15.2	4.6	3.0	3.9	3.9
29 Nov-26 Dec	152.84	27.01	3.8	7.7	3.6	2.0	5.9	4.0
26 Dec-22 Jan	139.87	13.59	1.4	6.5	1.1	0.7	2.8	1.0
22 Jan-18 Feb	97.01	14.06	1.8	5.3	1.6	1.3	2.4	1.6
18 Feb-17 Mar	124.77	32.08	3.4	14.9	5.4	3.2	2.1	3.0
17 Mar-13 Apr	152.85	53.04	8.6	17.5	7.9	6.2	8.3	4.5
13 Apr-10 May	169.23	47.95	7.1	16.7	7.6	4.2	7.2	5.1
10 May-06 Jun	142.61	34.31	2.5	12.1	5.2	3.6	5.8	5.1
06 Jun-03 Jul	279.20	78.06	9.8	30.7	10.2	7.8	12.7	6.8
03 Jul-30 Jul	93.21	25.84	2.5	11.2	4.8	2.8	2.0	2.5
30 Jul-26 Aug	181.92	55.56	3.5	19.3	11.9	3.6	8.9	8.3
26 Aug-22 Sep	84.76	29.65	2.5	11.7	6.2	2.9	3.0	3.4
22 Sep-19 Oct	77.72	22.38	4.2	8.6	3.5	2.2	1.7	2.1
Annual (g m^{-2})	50.53	13.15	1.55	4.98	2.07	1.22	1.87	1.44

Table 6.4. Continued. Clay mineral fluxes to the northern Bay of Bengal. All fluxes are in mg m^{-2} day^{-1}

North Bay of Bengal- 03 Deep

	Total flux <1mm	Lith. flux	Smectite flux	Illite flux	Chlorite flux	Kaolinite flux	Quartz flux	Plagioclase flux
02 Nov-29 Nov	101.59	40.64	7.7	21.1	3.7	2.4	3.7	2.1
29 Nov-26 Dec	133.01	29.04	3.8	14.9	3.1	1.8	3.7	1.7
26 Dec-22 Jan	169.02	24.16	3.0	12.1	2.0	1.3	3.8	1.9
22 Jan-18 Feb	127.32	22.99	2.3	12.3	2.0	1.9	3.0	1.5
18 Feb-17 Mar	96.69	26.76	3.3	11.7	3.2	2.5	4.0	2.1
17 Mar-13 Apr	142.57	47.67	8.1	21.5	5.2	3.1	7.3	2.6
13 Apr-10 May	168.50	48.59	17.1	15.9	5.0	4.0	4.3	2.3
10 May-06 Jun	157.76	43.23	7.1	19.9	4.7	3.5	4.9	3.1
06 Jun-03 Jul	192.61	75.45	23.1	31.8	6.3	3.2	5.7	5.3
03 Jul-30 Jul	151.12	53.48	14.3	15.9	6.0	4.5	6.8	5.9
30 Jul-26 Aug	165.36	50.23	14.0	17.2	7.0	2.2	5.2	4.6
26 Aug-22 Sep	103.68	38.41	6.8	17.4	5.0	1.9	4.3	2.9
22 Sep-19 Oct	95.96	32.93	8.4	12.7	4.3	2.3	2.6	2.7
Annual (g m^{-2})	50.68	14.98	3.34	6.30	1.61	0.97	1.67	1.1

Table 6.5. Clay mineral fluxes to the central Bay of Bengal. All fluxes are in mg m-2 day-1.

Central Bay of Bengal- 03 Shallow

	Total flux <1mm	Lith. flux	Smectite flux	Illite flux	Chlorite flux	Kaolinite flux	Quartz flux	Plagioclase flux
02 Nov-29 Nov	133.14	35.46	16.1	7.6	2.4	2.9	3.0	3.5
29 Nov-26 Dec	76.35	14.54	5.3	4.3	1.4	1.7	0.9	1.0
26 Dec-22 Jan	144.51	29.87	10.3	7.8	2.4	3.0	3.8	2.6
22 Jan-18 Feb	98.56	10.11	4.3	3.0	0.5	1.0	0.6	0.7
18 Feb-17 Mar	69.98	12.18	4.2	4.4	0.9	1.1	0.8	0.8
17 Mar-13 Apr	72.56	10.91	2.6	5.5	0.7	1.1	0.5	0.5
13 Apr-10 May	92.05	12.42	2.0	6.8	0.5	1.0	1.0	1.1
10 May-06 Jun	106.76	24.66	1.5	14.6	2.4	2.2	1.9	2.1
06 Jun-03 Jul	230.96	79.77	18.9	38.5	4.5	6.3	5.6	5.9
03 Jul-30 Jul	168.34	44.31	11.4	17.3	2.3	4.4	5.1	3.9
30 Jul-26 Aug	156.22	31.45	9.8	10.0	2.8	3.0	3.1	2.8
26 Aug-22 Sep	78.92	24.55	10.7	6.6	1.6	2.4	1.6	1.7
22 Sep-19 Oct	121.29	44.67	26.2	7.3	2.6	3.2	2.2	3.2
Annual (g m^{-2})	43.51	10.526	3.46	3.75	0.70	0.93	0.85	0.84

Central Bay of Bengal- 03 Deep

	Total flux <1mm	Lith. flux	Smectite flux	Illite flux	Chlorite flux	Kaolinite flux	Quartz flux	Plagioclase flux
02 Nov-29 Nov	115.10	38.39	10.8	11.6	4.0	5.8	2.9	3.4
29 Nov-26 Dec	120.23	41.50	17.2	6.8	3.8	6.3	4.8	2.6
26 Dec-22 Jan	161.91	37.00	9.4	15.3	2.9	2.9	3.8	2.7
22 Jan-18 Feb	147.32	33.72	8.1	12.8	2.9	3.9	3.4	2.6
18 Feb-17 Mar	100.36	28.16	13.6	7.3	1.1	1.8	2.1	2.3
17 Mar-13 Apr	91.87	20.90	7.4	8.2	1.2	1.4	1.6	1.2
13 Apr-10 May	84.07	15.95	5.3	5.1	1.1	1.4	2.0	1.1
10 May-06 Jun	129.08	33.00	10.2	11.3	3.3	2.2	3.6	2.4
06 Jun-03 Jul	188.43	58.32	12.3	21.4	6.0	7.4	5.3	5.9
03 Jul-30 Jul	203.72	71.90	25.4	21.0	6.3	6.9	6.9	5.4
30 Jul-26 Aug	362.38	134.16	20.8	49.2	11.0	22.1	14.0	17.0
26 Aug-22 Sep	133.98	53.78	29.6	10.6	3.1	5.1	1.9	3.5
22 Sep-19 Oct	Sample not recovered.							
Annual (g m^{-2})	55.92	17.24	5.17	7.74	1.42	2.04	1.59	1.52

Peninsular Indian Rivers a number of facts point towards the fact that smectites in the sediment traps are not directly derived from the rivers but are resuspended from the smectite rich muds on the continental slope of the east coast of India. i) Smectite fluxes in the northern Bay of Bengal traps peaks during the pre-monsoon period when river discharge of the Peninsular rivers is at a minimum. Moreover, smectite fluxes even in the central Bay of Bengal traps during this period are at a minimum. ii) A high smectite flux in the deep central Bay of Bengal traps trap during March (Cup No. 5) is not reflected in the shallow trap iii) Smectite flux increases with depth at both the stations and especially in the northern Bay of Bengal a sharp increase of smectite flux is seen only below 1500 m. iv) Peninsular Indian rivers contribute only 3% of the total lithogenic input to the Bay of Bengal but smectite fluxes to the deep traps constitute over 20% percent of the annual lithogenic flux. As mentioned before smectite is below detection limits in the suspended load of the Himalayan rivers.

Illite/ Smectite ratio: Illite/ smectite ratios show two trends (Fig. 6.3 to 6.7). A decrease from north to south and a decrease with depth especially during high flux periods. These ratios are generally low during the pre-monsoon months and high during the SW monsoon period. The decrease in the ratios towards the south is related to the decreasing influence of the Ganges-Brahmaputra river. Decrease in the Illite/ smectite ratios with depth means that either illite particles are getting disaggregated or smectite is additionally being scavenged at depth. In fact, between April and August, Illite/ smectite ratios of the deep northern trap is lower than the shallow Central trap.

Quartz: Quartz fluxes in the northern Bay of Bengal peaks during the premonsoon period (Table 6.4). Flux patterns of quartz do not follow those of either illite or quartz indicating a possible eolian source. *Goldberg and Griffin*

(*1970*) measured high eolian dust content over the Bay of Bengal but thought it to be insignificant compared to the fluvial input. Eolian fluxes can be significant during summer when river discharge is low and number of dust storms are more. Additional support for an eolian source is from the haze charts (*McDonald 1938*) which show maximum number of hazy days during the premonsoon period especially in the northern Bay of Bengal. *Middleton* (*1989*) studied the distribution of dust storms over India and showed that considerable amounts of dust is raised over the northern India due to 'Andhi' (localized convective cells) and 'Loo' winds(a pressure gradient airflow) Maximum number of aandhi's were recorded between April and June. The source area of these dust are in the Thar Desert of Rajasthan. Of the three main trajectories of dust from this area the main flow is towards ENE down the Indo-Gangetic Plains. Part of the lithogenic flux in the northern Bay of Bengal may have been derived from the plains of north India and deliverd to the trap site by these winds.

Kaolinite and Chlorite: Kaolinite in the Bay of Bengal is derived mostly due to chemical weathering of felsic rocks of the Indian Peninsula. The peninsular rivers have a higher kaolinite content (6 to 25%) compared to the Himalayan rivers (below detection limits). Chlorite on the other hand is derived mostly due to physical disintegration of rocks and therefore up to 25% chlorite is found in the suspended load of the Ganges-Brahmaputra rivers (*Konta, 1985*). Kaolinite and chlorite rations therefore show a clear north-south gradient, being consistently higher in the central part of the Bay (Fig 6.8).

Variation of clay mineral flux with depth. As stated in the previous chapter, lithogenic flux in the northern and central Bay of Bengal increase linearly with depth. Such linear increase in lithogenic flux with depth has also been reported in other parts of the world (*Honjo et al., 1982b*). However, the

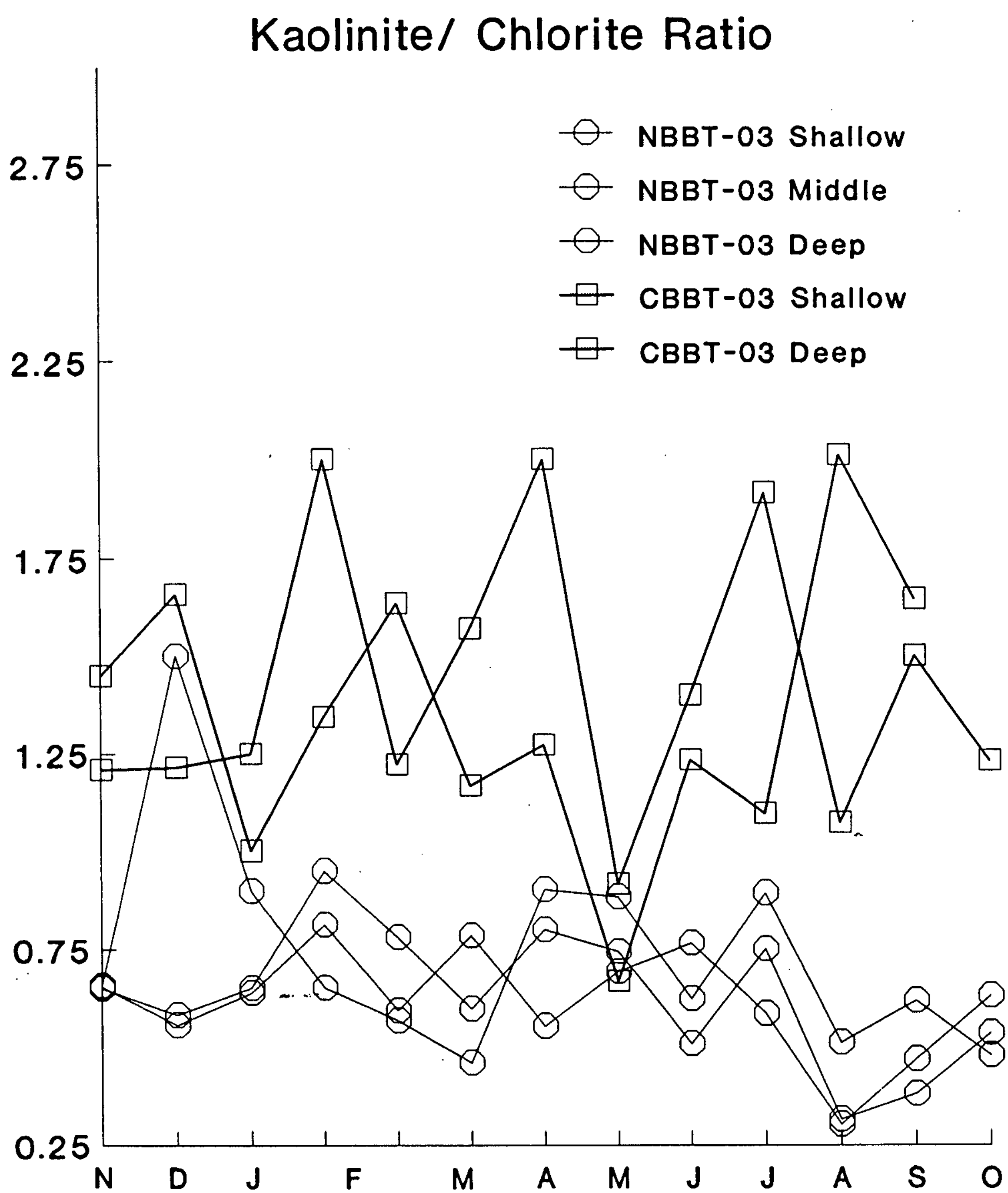


Fig. 6.8 Kaolinite and chlorite ratios in the northern and central Bay of Bengal

present study shows that increase in lithogenic flux with depth is not linear if we consider individual mineral species.

During the SW monsoon the gradient of increase with depth of smectites is higher than that of illites at both the sites in the Bay of Bengal (Fig. 6.9). Furthermore, extrapolation of the smectite flux to the surface shows that most of the smectite is derived from the deeper layers by scavenging processes and a negligible amount of smectite is introduced at the surface. In the northern Bay of Bengal almost the entire smectite flux is derived from subsurface waters. Varying gradients for different mineral species can only be explained by change with depth of the composition and concentration of suspended sediments in the water column.

The surficial low saline layer over the entire Bay of Bengal is formed mainly by the river discharge of the Ganges-Brahmaputra rivers. Isohaline patterns during the SW monsoon period show that most of the low saline water at the surface is transported towards the south (*Murthy et al.,1990*). Therefore the finer fraction of the suspended load of the Ganges-Brahmaputra rivers should be advected towards the south within this layer. This is seen in the higher concentrations of illite even in the central shallow trap where it is expected that smectite should have been the dominant clay mineral. In may therefore be concluded that illite rich Ganges-Brahmaputra sediments in the surficial low saline layer are aggregated by biogenic processes and transported to the sea floor. During their descent the large aggregates scavenge smectite rich suspended sediments advected over the trap sites by deep water currents.

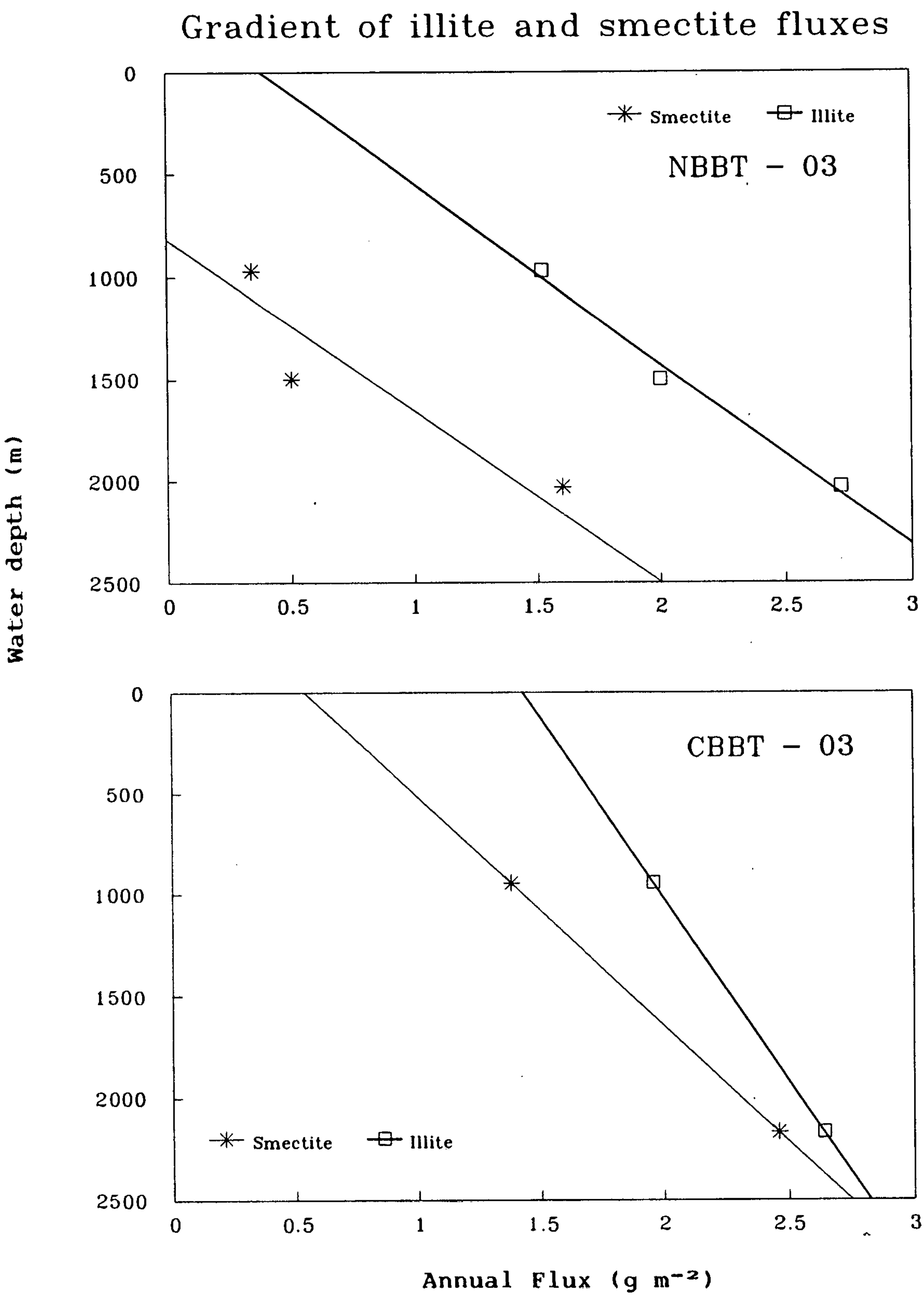


Fig. 6.9. Gradient of illite and smectite fluxes in the northern and central Bay of Bengal. Note the steeper gradient for smectite fluxes at both the sites.

6.3.3. *Relation between mineral input, mineral fluxes in the water column and mineralogy of sea floor sediment*

Smectite, form only 3 to 6% of the total lithogenic input into the Bay of Bengal, but constitute more than 20% of the annual lithogenic flux in the sediment traps. The Deccan trap province which extends from 18 °N to south of the equator contains more than 70% smectites (*Venkatrathnam and Biscaye, 1973; Kolla and Rao, 1990*). The high smectite percentages in the sediment traps and cores deserve an explanation. Most of the illite rich sediments of the Ganges-Brahmaputra rivers are deposited initially in the estuaries and shelf regions and later transported directly to the deep sea via the Swatch of no Ground and deposited as turbidites. The fine fraction (<15 micrometer) of the Ganges Brahmaputra river discharge which bypasses the shelf is carried in the surficial low saline layer and tends to be rapidly removed near the coastal high productive zone in the northern Bay of Bengal. Smectite rich sediments on the other hand are easily resuspended from the shelf and slope regions on the east coast of Indian and advected at depth for long distances by coastal and other boundary currents. The major difference between illite and smectite sources is that illite is delivered to the deep Bay of Bengal by one major river across a broad shelf, while smecites are deliverd by many rivers across a narrow shelf. It is seen that only turbidites in the Bengal fan contain high illite percentages whereas hemi-pelagic assemblages contain relatively low illite but up to 50% smectite (*Brass and Raman, 1990*).

Kolla et al., (*1976*) have speculated upon the existence of a southward flowing bottom current in the western Bay of Bengal which transports smectites to the distal Bengal Fan. Our data on the other hand shows that during the premonsoon and SW monsoon periods, smectites are advected at depth towards the north. Similar conclusions have also been reported by *Rao et al.* (*1990*) and *Mallick* (*1976*) from the study of shelf and slope sediments. Probably the bulk of the southward transport of smectites takes place south of 15° N where surface circulation patterns are different from that of the northern Bay of Bengal. A cyclonic gyre is present between 15° and 18° N and south of this an anticyclonic gyre is present in the western Bay region (*Murthy et al., 1990*).

In summary, it is seen that in most of the terrigenous particles in the northern Indian Ocean is derived due to physical weathering of rocks in the Himalayas and to a lesser extent derived from the Indian Peninsula and the deserts of Arabia and north Africa. At present the Arabian Sea is dominated by eolian sediments and dust particles from the Arab-Somalia region reach even the eastern part of the sea.

The Bay of Bengal is dominated by fluvial sediments derived from the Ganges-Brahmaputra and Peninsular Indian rivers. In this area there is a marked change in the clay mineral compositon of settling particles with depth. Suspended solids form the Ganges-Brahmaputra rivers are carried in the surfical low saline layer while sediments from the Peninsular Indian rivers is laterally advected at depth.

Chapter 7

Grain Size Analysis of the Lithogenic Fraction of Settling Particles and Surface Sediment in the Arabian Sea and Bay of Bengal

7.1. Introduction

Particles size in the oceans have a size range of 8 to 9 orders of magnitude (*Lambert et al., 1981*) from the submicrometer macromolecule to the largest whales. Of this, only the lowermost part of this broad spectrum is relevant to the study of biogeochemical cycles in the oceans. Particles settle according to their hydrodynamic characteristics, mainly a function of its size and density, and form sediment deposits like sand silt or clay. When sediments from contrasting parent deposits are redeposited, mixture of sizes known as population are formed. For example muds may consist of a mixture of clay and silt. Thus, size distribution in sediments can be related to i) The availability of different sizes of particles in parent material and ii) physical processes operating at the site where the sediments are deposited.

Statistical analysis of grain size data has been successfully used in coarse grained sediments to decipher the origin of sands and to differentiate between beach, river and eolian deposits (*Folk and Ward, 1962*). One of the reasons why such techniques have not been routinely applied to fine grained sediments is that conventional methods of measuring grain size of clays is tedious and time consuming.

A number of new automated electronic instruments to measure grain size of sediments have recently been developed. They include the Sedigraph, Coulter Counter and laser based systems like Malvern particle size analyzer and GALAI-CIS. Each of these instruments measures a different physical property of particles to arrive at its grain size distribution. For example the

Sedigraph uses the Stokes settling velocity to measure grain size, while the Coulter Counter measures the voltage drop between an orifice when a particle partially blocks the orifice; the voltage drop being proportional to the volume of the particle. Grain size distribution measured using different techniques are not easy to compare because principle of measurement and treatment of data are different. This is particularly difficult for fine grained sediments. For example the GALAI-CIS laser particle size analyzer gives a mean particle diameter which is about 1.5 to 2 times to that measured using a Coulter Counter (*Jantschik et al., 1992*).

7.2. Material and methods

Grain size measurements were carried out on the lithogenic fraction of the Arabian Sea and Bay of Bengal sediment trap and core top samples. In the Arabian Sea the samples chosen were the deep traps of the eastern, central and western Arabian Sea collected between May 1986 and May 1987 (for details see table 3.1). In the Bay of Bengal, grain size measurements were made on the shallow and deep sediment traps in the northern and central Bay of Bengal, collected between November 1988 and October 1989. In addition 9 core tops from the Arabian Sea and 3 core tops from the Bay of Bengal were provide by Dr. Ittekkot of Hamburg University for similar analysis (See Table 7.1 for location of cores). The core tops had been vacuum dried so there was no significant alteration in their grain size during storage.

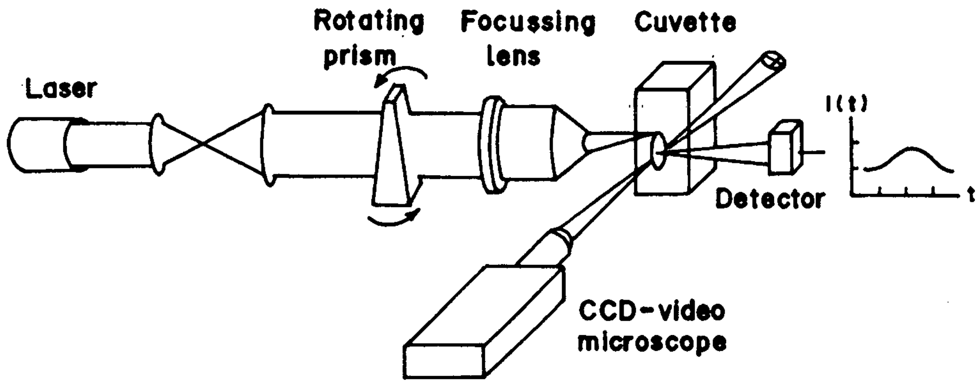
The lithogenic fraction of each sample was isolated by removing all other major components, i.e. carbonates, organic matter and opal by treatment with acetic acid, hydrogen peroxide and sodium carbonated respectively (see Chapter 3 for details). The lithogenic fraction was thoroughly disaggregated by treatment in an ultrasonic bath for half an hour. The sample was then diluted with a peptising solution (2% sodium hexa meta phosphate in distilled water) and introduced into the sampling chamber of the grain size analyzer.

Table 7.1 Station locations and water depth of core tops in the northern Indian Ocean which have been analysed for grain size .

Station No.	Latitude	Longitude	Water depth (m)
KL87	10^0 30.05′	57^0 44.22′	3773
KL82	12^0 41.09′	58^0 40.62′	4416
KL79	13^0 38.84′	58^0 19.56′	4351
KL74	14^0 19.26′	57^0 20.82′	3212
KL71	16^0 14.17′	60^0 15.35′	4029
KL70	17^0 30.69′	61^0 41.82′	3810
KL57	20^0 54.47′	63^0 07.32′	3422
KL51	20^0 57.92′	65^0 33.54′	2644
KL36	17^0 04.49′	69^0 02.68′	2055
KL26	15^0 30.86′	68^0 45.61′	3772
KL15	14^0 52.82′	64^0 44.79′	3920
NBBC	17^0 60.00′	89^0 01.00′	2263
CBBC	13^0 09.00′	84^0 01.00′	3259
SBBC	04^0 06.00′	87^0 09.00′	4017

7.2.1. Grain size measurements

Grain size of the lithogenic fraction was measured by a GALAI TM (model CIS-1) laser particle size analyzer (*Karasikov et al. 1987*). This instrument has arrived in the market only about 4 to 5 years ago and is not many laboratories have started using this instrument due to its high cost. This instrument employs a time-size mapping called the Time-Of-Transition theory to directly measure particle size. A He-Ne laser beam is scanned circularly by rotating a wedge prism and focussed down to a 1.2 micrometer spot. The time taken for a particle to transit a laser beam moving at fixed velocity, depends on the particle's diameter and generates a interaction signal. These signals are detected by a photodiode. Since the beam rotates at a constant speed the duration of interaction provide a direct measurement of each particles size. Knowing precisely the angular velocity of the beam, the size of



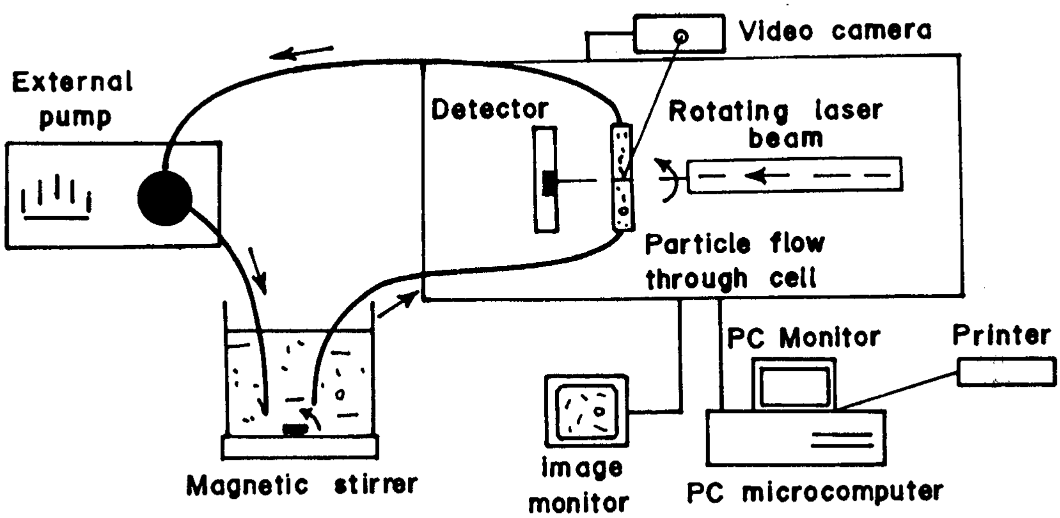


Fig. 7.1 Figure showing working principle of the GALAI-CIS laser particle size analyser.

each particle scanned at the focus can be calculated from the duration of obscuring of the beam. Sophisticated pulse analysis algorithms are employed to reject out of focus and off center interactions. Particle-laser interactions are processed by a desktop computer to yield particle diameter size. The standard measuring range extends from 0.5 micrometer to 150 micrometer in 300 steps of 0.5 micrometer each. A CCD-TV microscope is also incorporated into the basic unit allowing the operator to monitor the sample while it is being measured to observe the nature of particles and to visually control the occurrence of aggregation. It is possible to observe the occurrence of micro bubbles (formed mostly due to difference in sample and room temperature or due to the peristaltic pump or stirrer) in the sample which would give erroneous results. The sample is introduced into a liquid flow-through cell embedded into a closed circuit configuration. A peristaltic pump recycles the sample through the sampling cell, thus facilitating analysis of minute quantity of sample.

For each sample, the diameter of about 100,000 particles were measured routinely and converted to volume percentage. Although the instrument is permanently calibrated at the factory, the calibration was checked using latex spheres of know diameters. An excellent callibration was obtained wit latex spheres of 2.03 and 4.98 micron diameter showing a mean diameter of 2.3 and 5.5 micrometer. The small difference between the two was probably due to the fact that this instrument measures the maximum diameter of particles; The latex spheres had probably lost their shape but not their volume (*Karashikov, personal communication*).

Jantschik et al. (*1992*) have made a preliminary comparison of the performance of the GALAI-CIS system with that of the Coulter Counter by measuring the same suspended sediment sample on both the machines.

Both the system showed a bimodal distribution between 1 to 20 micrometer. However the 3.5 micron peak in the Coulter Counter corresponded to the 7 micron peak in the GALAI-CIS system. The authors were not sure what exactly is the reason for this discrepancy and advocated further studies. They were also not sure which machine is more reliable.

The major problem with grain size analysis of clays is due to their shape. Most modern instruments are designed to measure spherical objects while clays are mostly in the form if sheets. It is difficult to correct for shapes of clays as their aspect ratio (length: bredth: width) is different for various mineral species. Since interpretation of the grain size data in the present thesis is based on seasonal variation in grain size and not on the absolute diameter of the particle the results presented should be acceptable.

7.2.2. Data processing and presentation

The grain size data collected by the GALAI-CIS was processed on a personal computer to eliminate spurious values and to calculate the volume percentage. The grain size information over a size range of 0.5 to 150 micrometer is collected in 300 channels, with each channel representing a size interval of 0.5 micrometer. The basic information provided by the instrument is the number of particles in each size range. This parameter is not very useful as it is seen that the number of particles increase monotonously with decrease in grain size in all the samples. The number distribution was therefore converted to volume percentage by multiplying the number of particles in each size range by its specific volume ($4/3\pi r^3$). Volume percentage of each sample was plotted on a log-noraml scale with grain diameter (in micrometer) on the **X** axis and Relative Volume distribution (Volume %) on the **Y** axis.

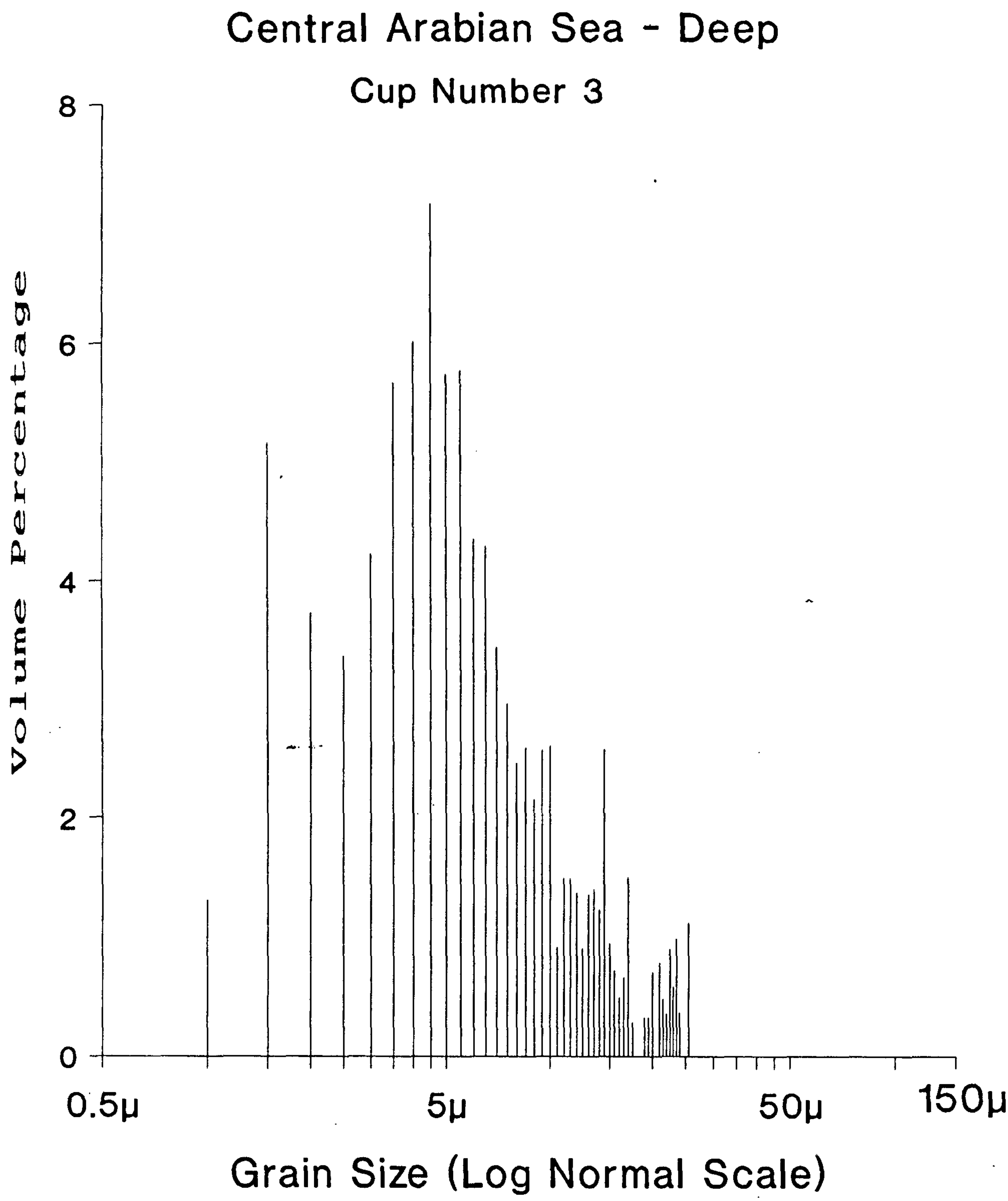


Fig. 7.2 Histogram showing grain size distribution of lithogenic particles in the deep central Arabian Sea trap (Cup No. 3 sampling period 5th to 18th June, 1986). On the X axis is the grain diameter (log scale) and on the Y axis is the volume percentage of particles.

Fig. 7.2 shows grain size distribution of the lithogenic fraction from the central Arabian Sea. On the X axis is the grain size diameter from 0.5 to 150 micrometer in steps of 0.5 micrometer in a logarithmic scale. On the Y axis is the volume percentage of particles for each size range of 0.5 micrometer on a normal scale. The grain size distribution pattern followed a normal distribution pattern between 0.5 and 20 micrometer. In some of the samples a steady increase in volume percentage is seen above 20 micrometer. These values appear to be spurious as the number of particles for each size range is exactly one. These points were discarded and volume percentage recalculated. In subsequent figures only data from 0.5 to 20 micrometer is presented as most of the samples do not show any particles coarser than 20 micrometer. For comparison purpose the graphs were stacked one above the other. The size of the figures are not uniform as the number of samples at each location varied between 12 to 18. Gaps in the figures indicate that the quantity of sample available for analysis was too small for reliable measurements to be made. Notice that the Y scale is also different for each figure.

7.3. Results and discussions

The grain size of the lithogenic fraction in the Arabian Sea is extremely fine as particles greater than 20 micrometer account for less than 1% of the volume percentage (Fig. 7.3 to 7.5) indicating deep water deposition in a quiescent environment. In the case of Bay of Bengal, it is even more finer with particles less than 14.5 micrometer accounting for over 99% of the volume percentage in most of the samples (Fig 7.6 to 7.9). A similar trend is seen in the grain sizes of the core tops of sediments of the Arabian Sea and Bay of Bengal (Fig 7.11 and 7.12).

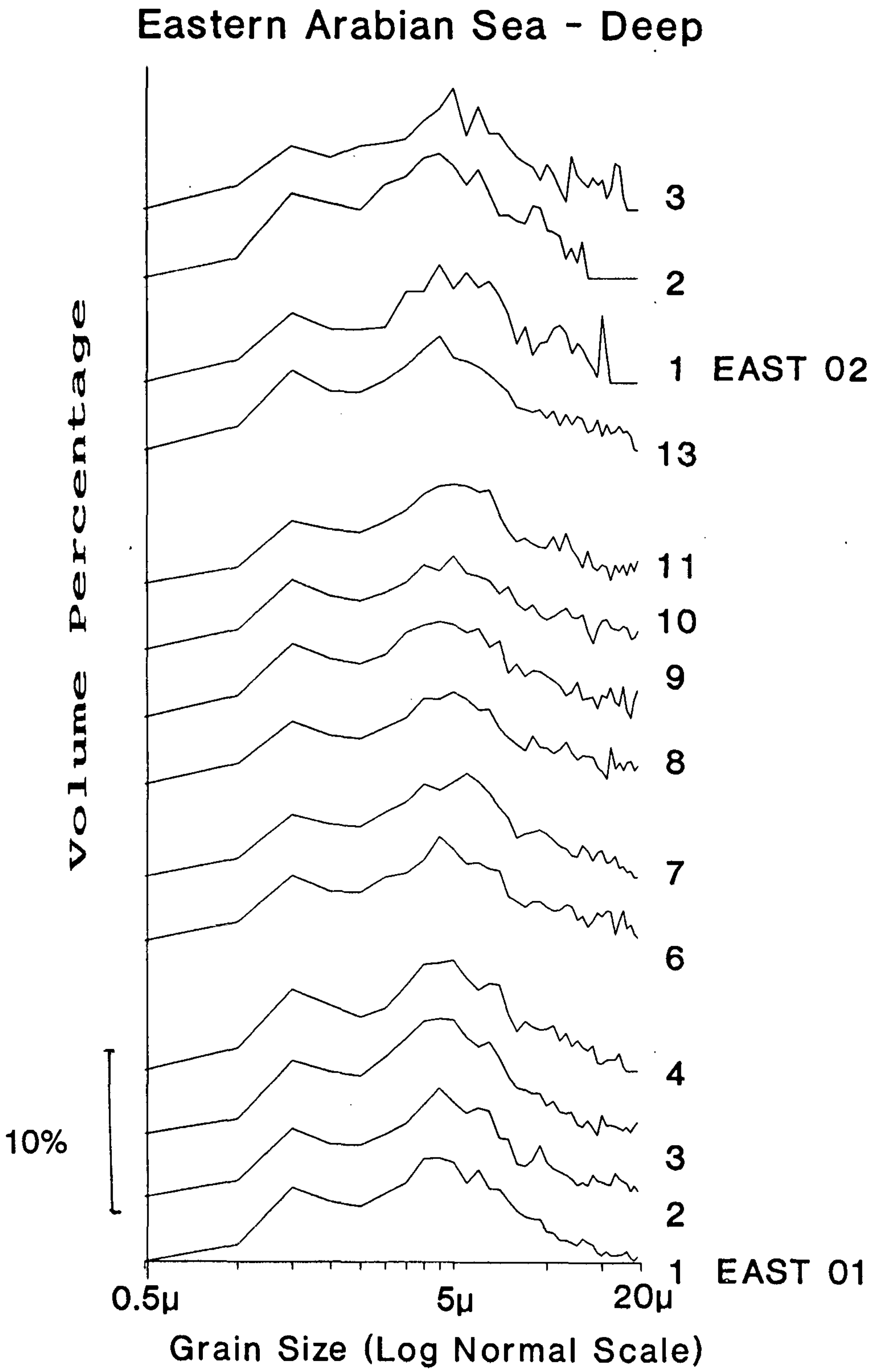


Fig. 7.3 Grain size distribution of the lithogenic fraction of the eastern Arabian Sea deep trap on a log-normal scale. Y scale is indicated by a vertical bar. EAST 01 and EAST 02 are the deployment number of the moorings. Numbers 1 to 13 represent sample numbers. For sampling period see table 6.1.

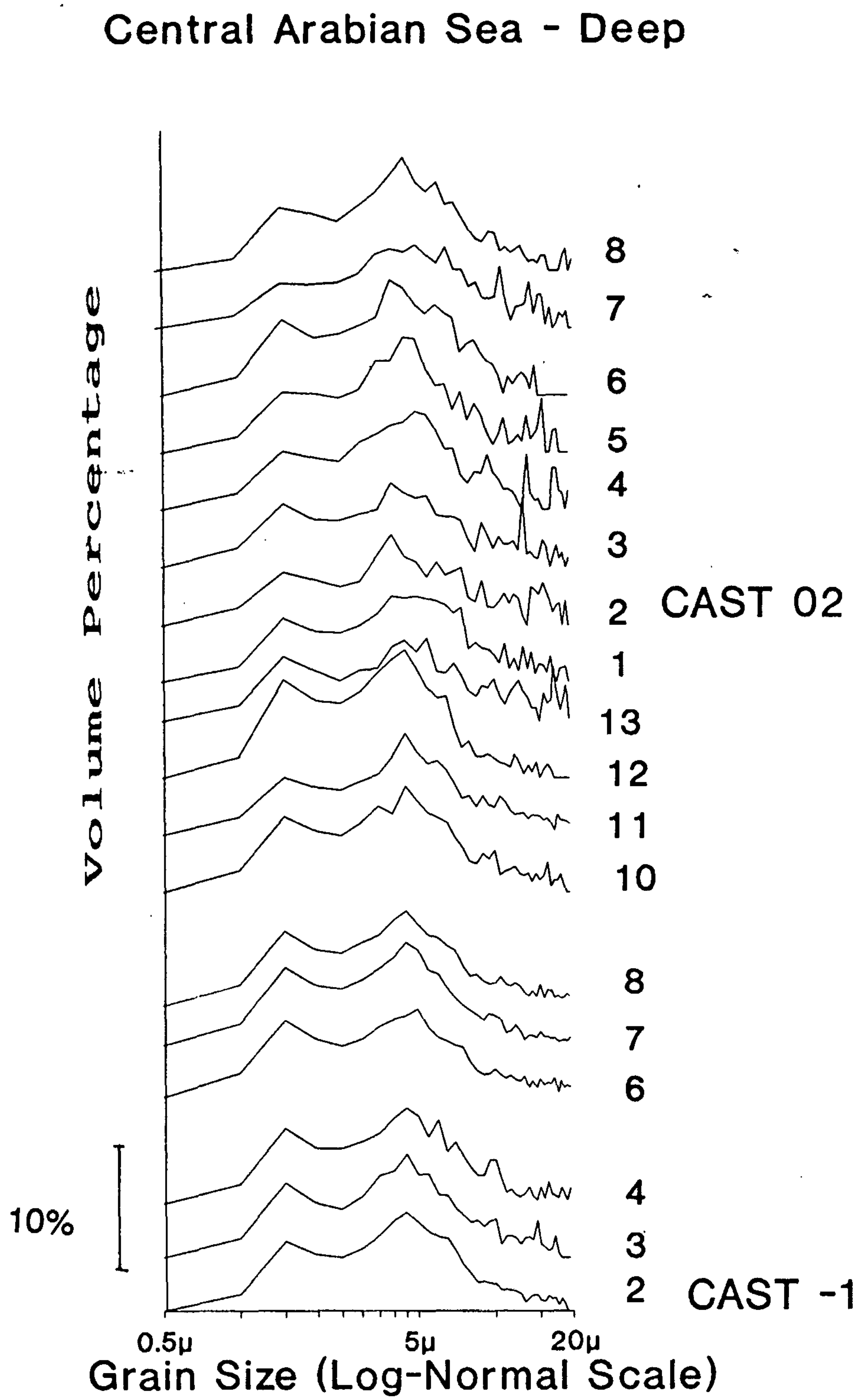


Fig. 7.4 Grain size distribution of the lithogenic fraction of the central Arabian Sea deep trap. Y scale is indicated by a vertical bar. For sampling period see table 6.2.

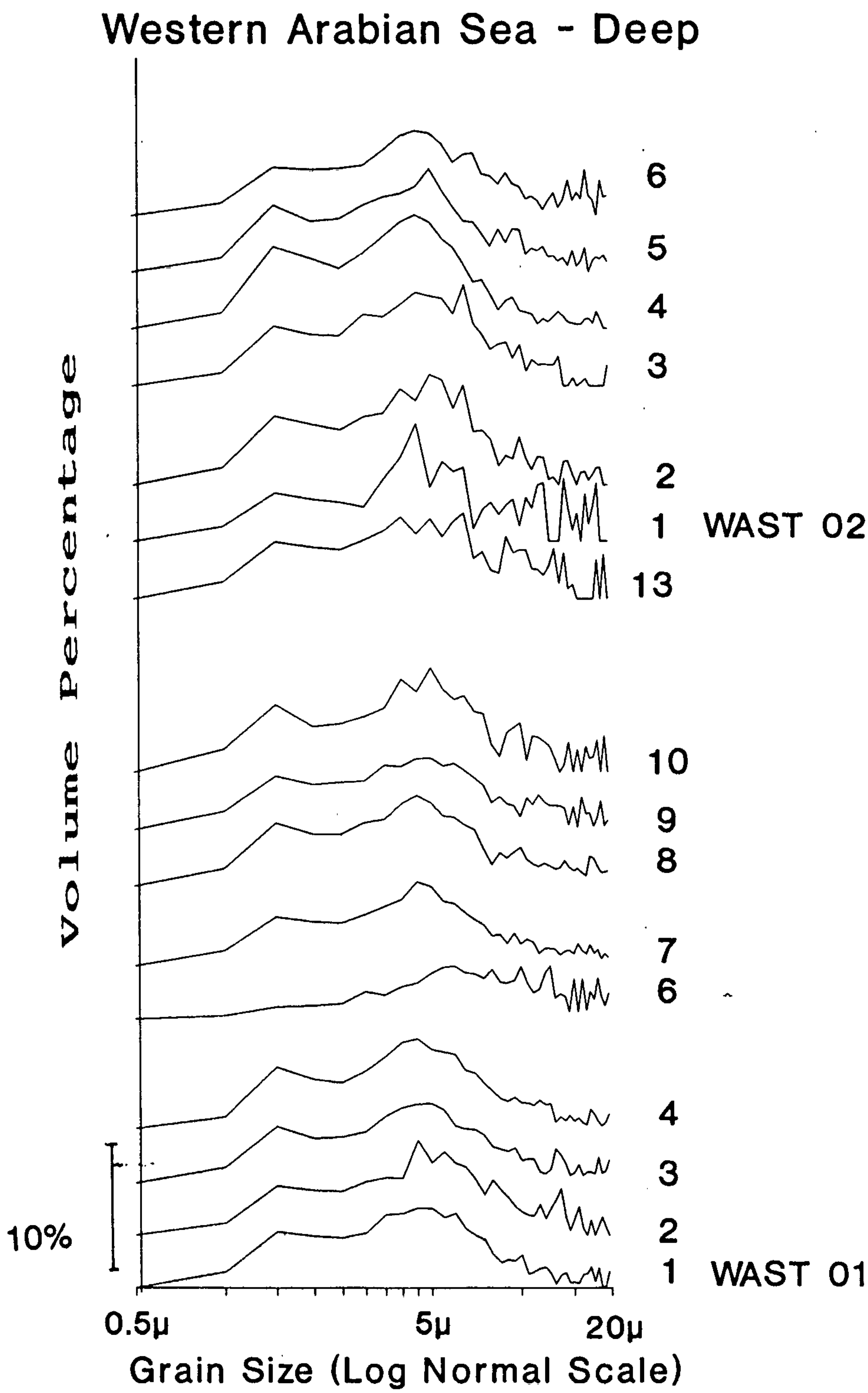


Fig. 7.5 Grain size distribution of the lithogenic fraction of the western Arabian Sea deep trap. For sampling period see table 6.3. Scatter between 14 to 20 microns is probably due to eolian particles.

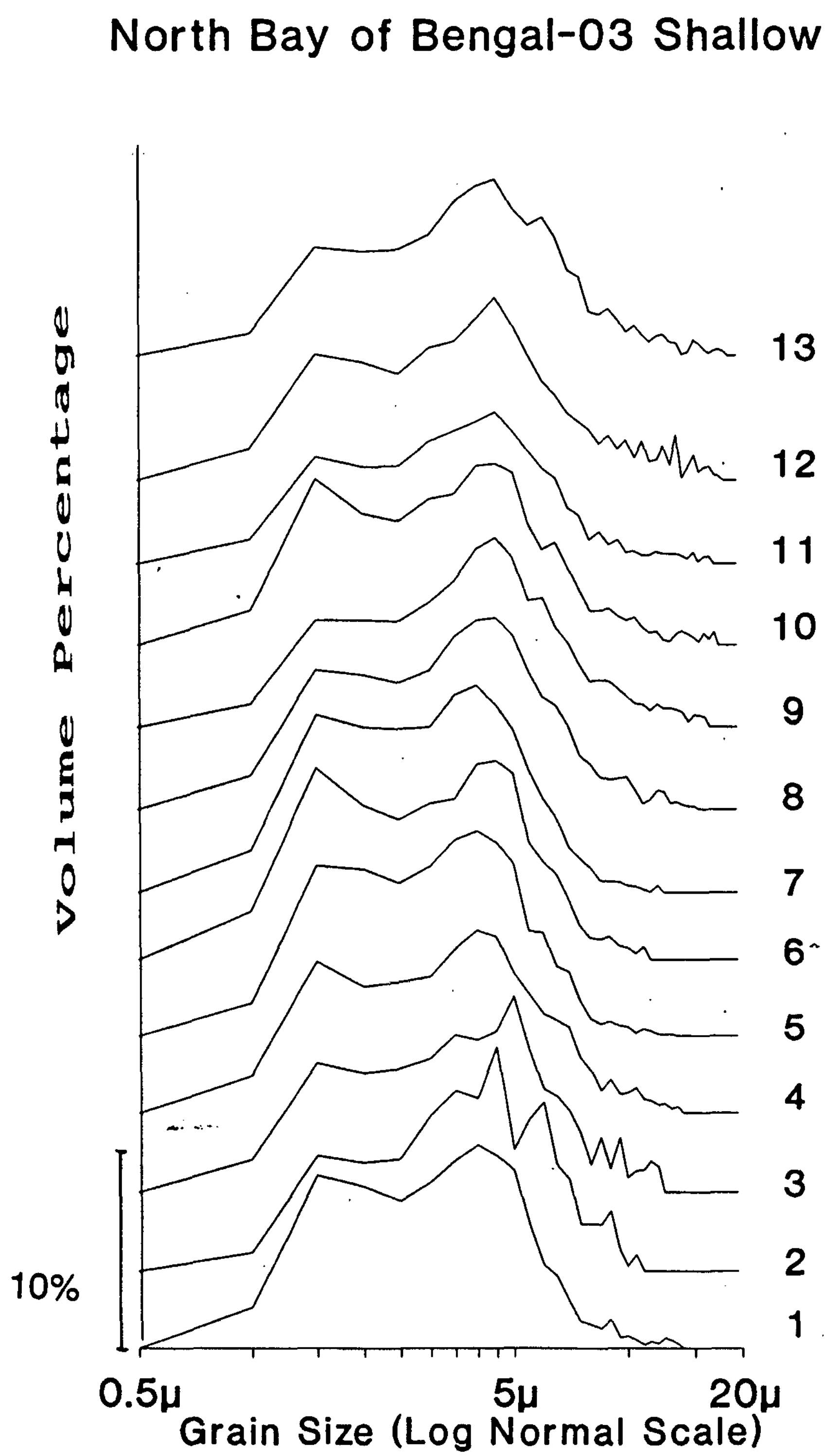


Fig. 7.6 Grain size distribution of the lithogenic fraction of the shallow northern Bay of Bengal trap. For sampling period see table 6.4. Notice that there are hardly any particles greater than 15 micro meters.

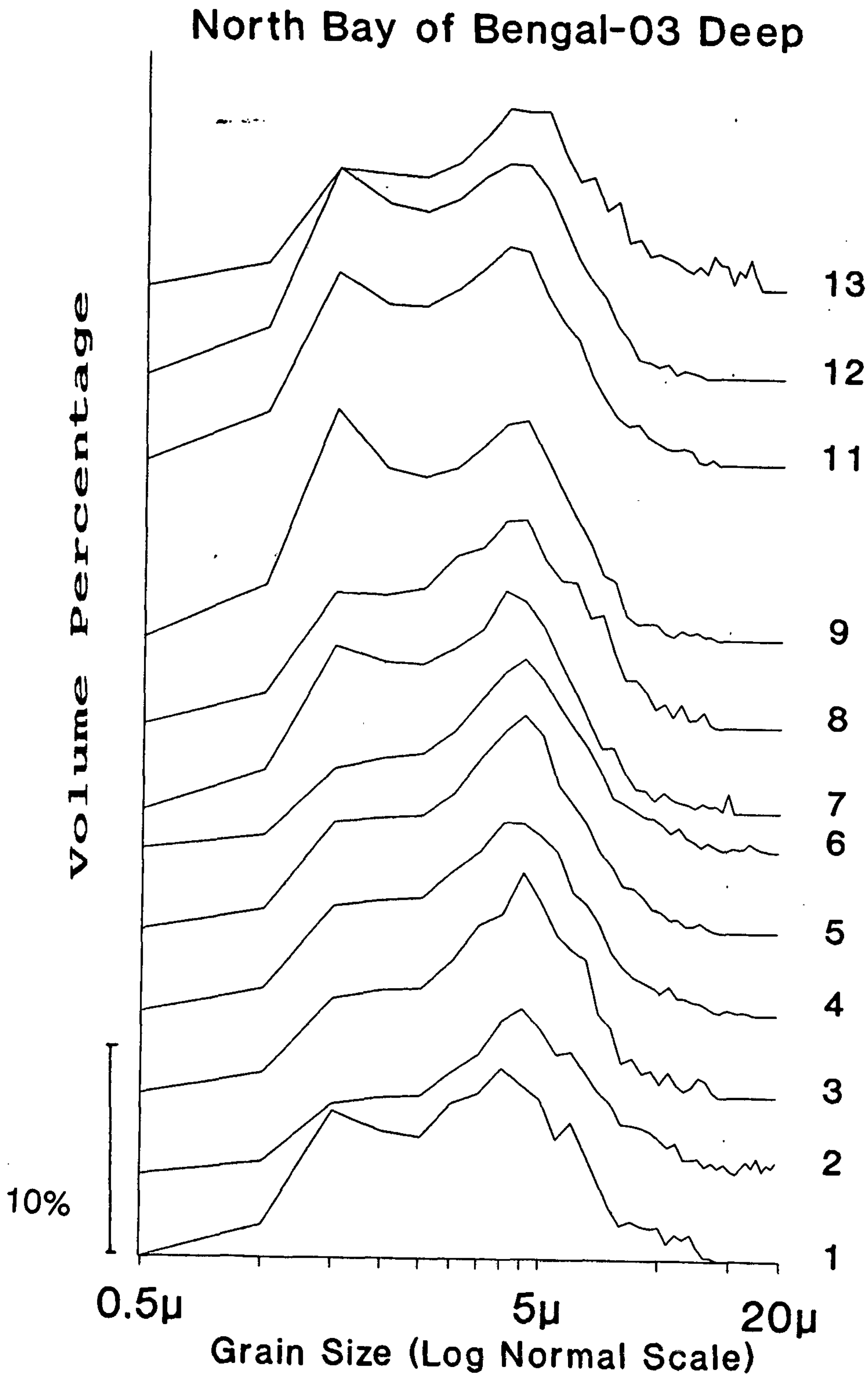


Fig. 7.7 Grain size distribution of the lithogenic fraction in the deep northern Bay of Bengal trap. For sampling period see table 6.4. Note the increase in percentage of particles in the 1.5 to 2.0 micrometer range during the late SW monsoon period (Cup No. 9-12).

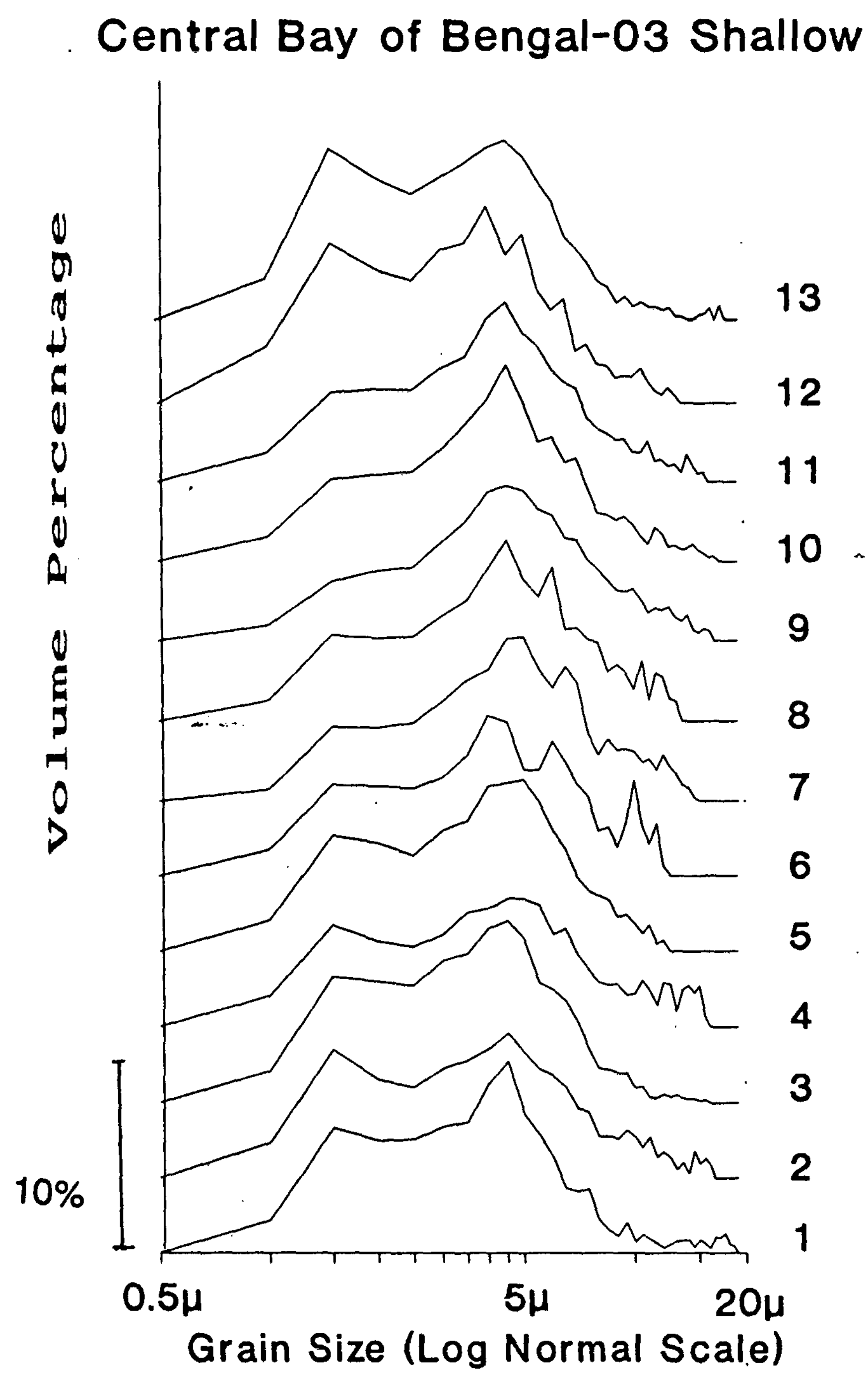


Fig. 7.8 Grain size distribution of the lithogenic fraction in the shallow central Bay of Bengal trap. For sampling period see table 6.5. Y scale is indicated by a vertical bar.

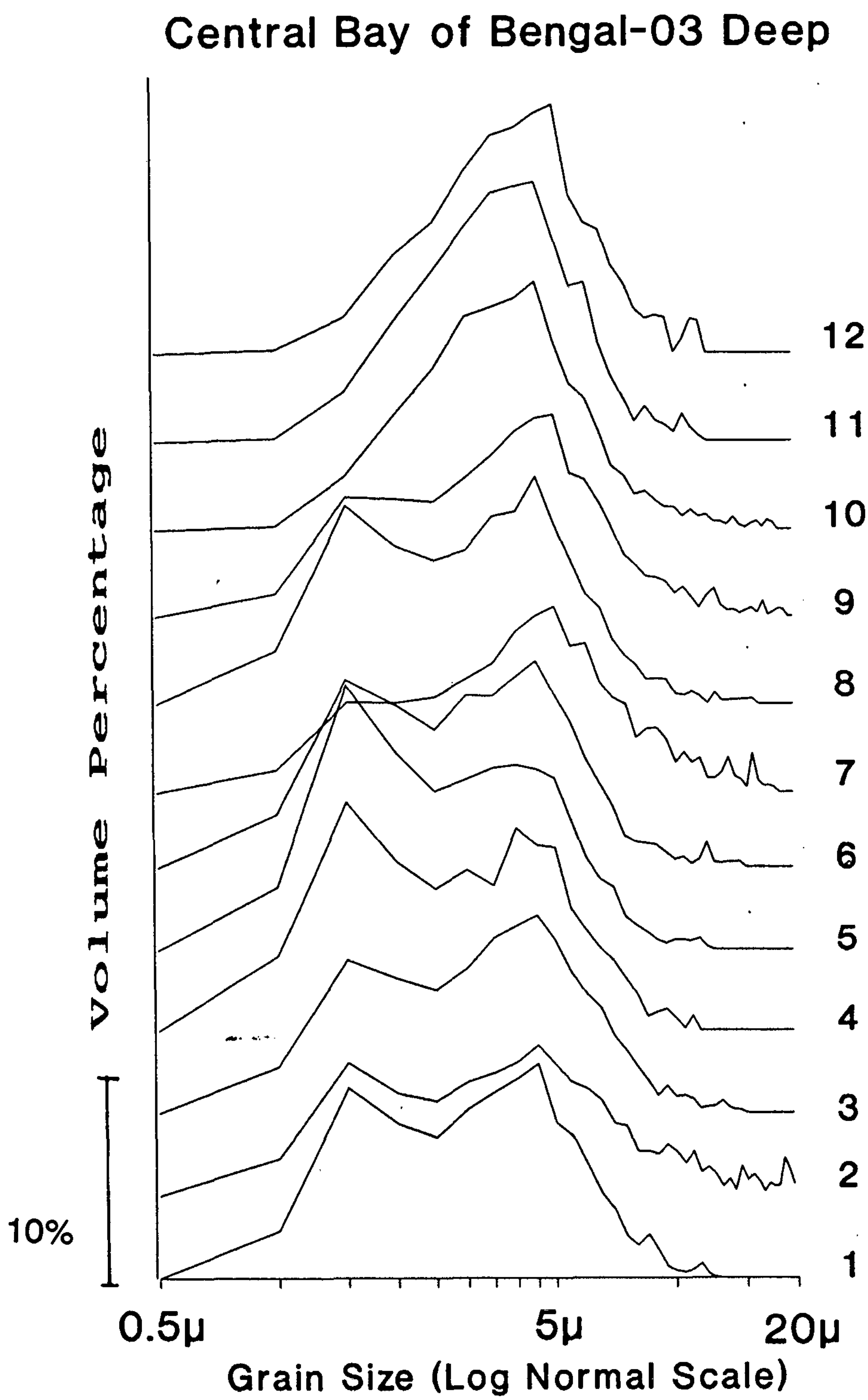


Fig. 7.9 Grain size distribution of the lithogenic fraction in the deep central Bay of Bengal trap. For sampling period see table 6.5. Y scale is indicated by a vertical bar.

The most characteristic feature of the grain size distribution are the modes at 1.5 and 4.5 micrometer. The bimodal distribution is seen at all sites in the northern Indian Ocean. They are probably related to clay mineral concentrations. It is known that in deep sea clays, particle diameter of kaolinite is the coarsest (up to 5 micrometer), followed by illite and chlorite and the finest sized clays are smectites. Electron micrographs show that in the Bay of Bengal, smectite and kaolinite are finer in size compared to illite and chlorite (*Bouquillon et al., 1990; Aoki et al., 1991*). In the Bay of Bengal smectite percentage increases and at the same time grain size becomes progressively finer with depth. This is particularly well demonstrated in the northern Bay of Bengal where smectite is introduced below 1500 m. In the deep traps over 10 percent of the total volume is composed of particles in the size range 1.5 to 2 micrometer whereas the shallow traps have only 5% of particles in this size range. The Arabian Sea lithogenic fraction also shows a bimodal peak as it contains a similar suite of minerals, but here the peak 1.5 micrometer peak is much lesser than the 4.5 micrometer peak due to lower concentration of smectites.

Significant intra-basin differences in grain size distribution are not seen but there are marked differences between the two basins. To bring out the difference more clearly two samples, one each from the central Arabian Sea and central Bay of Bengal are plotted on the same scale in Fig. 7.10. As mentioned before, particles in the Bay of Bengal are finer due to predominantly fluvial input with higher percentage of smectite whereas in the Arabian Sea sediments are dominated by eolian sediments with higher percentage of coarser grained illites and quartz. In the Bay of Bengal samples, particles in the size range 1.5 to 2 micrometer comprise over 8% of the total volume but for the Arabian Sea particle of the same size range

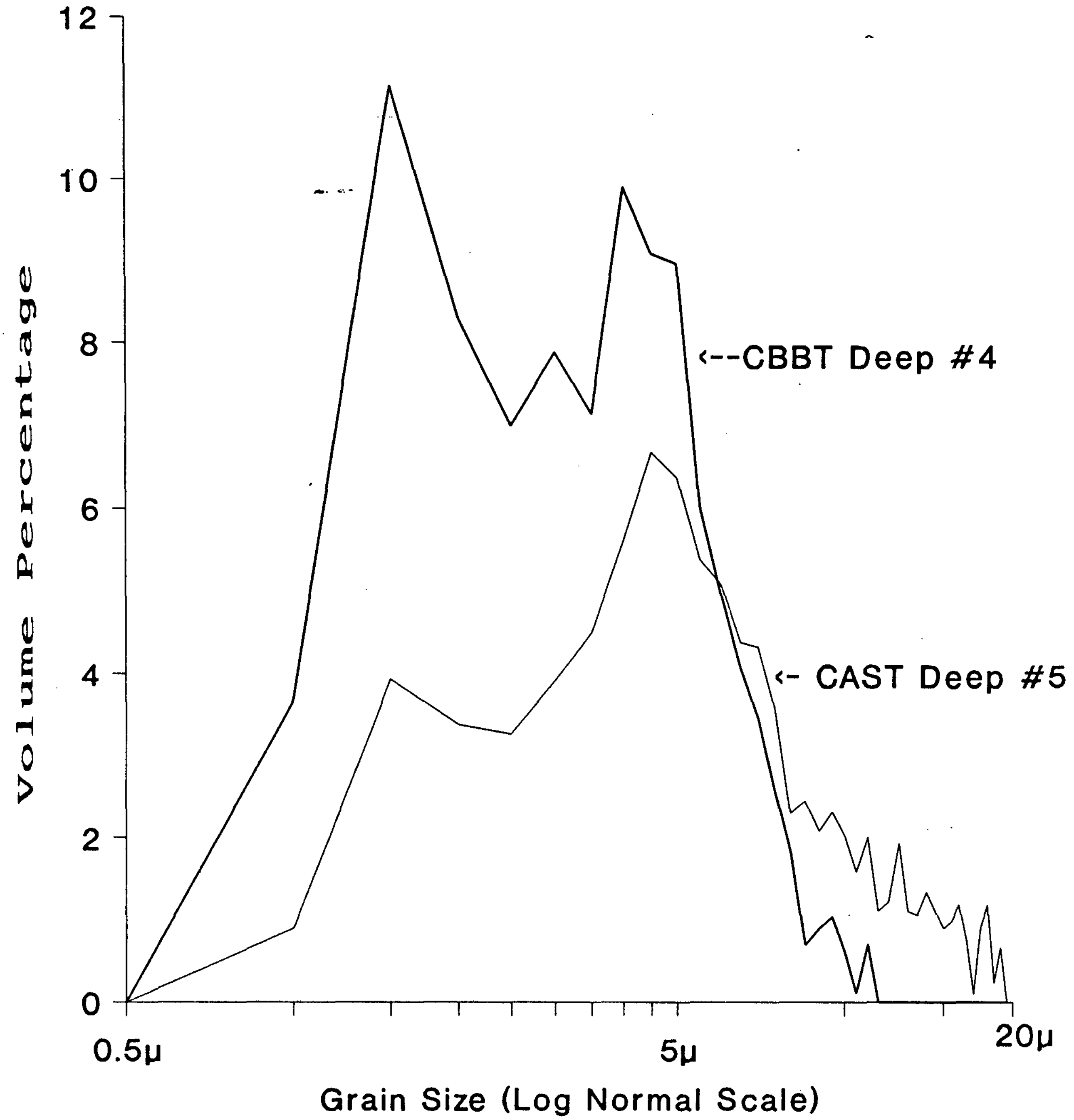


Fig. 7.10 Comparison of grain size distribution of the lithogenic fraction of settling particles from the Arabian Sea and Bay of Bengal. One representative sample from each area is taken. Compared to the Bay of Bengal, the volume percentage of fine particles (less than 5 µm) are lower in the Arabian Sea. Eolian particles, in the size range 14 to 20 µm, are distinctly higher in the Arabian Sea.

comprise less than 5.5% of the total volume. Similarly particles in the size range 4.5 to 5 micrometer comprise nearly 10% of the total volume in the Bay of Bengal but less than 6.5% in the Arabian Sea. In the central Bay of Bengal the 1.5 micron peak is not very prominent in some of the samples collected during the SW monsoon period. This peak appears to have shifted to 2 to 3 micrometer indicating slightly coarser material being deposited in this area, probably due to stronger currents along the western Bay of Bengal during this period (*Shetye et al., 1991*).

Particles in the size range 9 to 20 micrometer is much higher in the Arabian Sea. Particles coarser than 8 micrometer comprise 30 to 40 % of the total volume in the Arabian Sea whereas for the Bay of Bengal it is less than 15 per cent. These particles are most probably derived from eolian dust. *Clemens and Prell* (*1990*) reported that the median size of eolian sediments in the western Arabian Sea ranged between 14 to 17 micrometer. Grain size volume percentage distribution of the Arabian Sea shows erratic distribution in particles greater than 14 micrometer. Particles greater than 14 micrometer comprise less than 0.01% of the total number of particles but contribute to over 15% of the total volume due to their high specific volume. Because of the low number of particles in this size range it is not possible to get repeatable values but replicate analysis show that the scatter between 14 and 20 micrometer is always present in most of the samples. Also more coarse particles are present in the western Arabian Sea compared to the eastern Arabian Sea (Fig. 7.3 to 7.5).

Fig. 7.11 and 7.12 show the grain size volume distribution of the surface sediments of the two seas. All the Bay of Bengal sediments show a remarkable similarity in grain size distribution even though the distance between some of the samples is over a 1000 kilometers. There does not

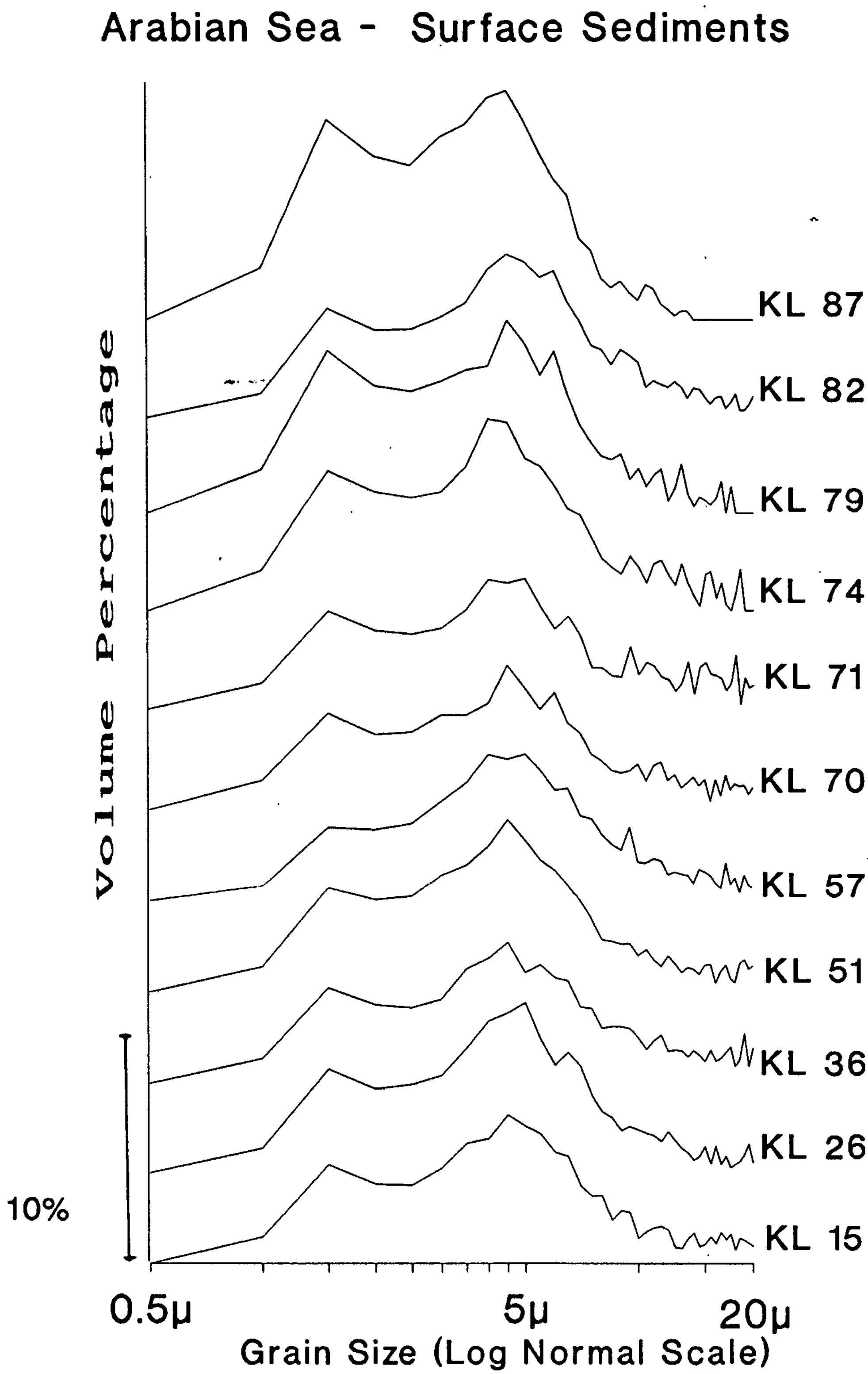


Fig. 7.11. Grain size distribution of the lithogenic fraction of core tops from the Arabian Sea. For sample locations see table 7.1. Y scale is indicated by a vertical bar. Eolian particle signature between 14 to 20 micrometer is seen in all the samples except KL 87.

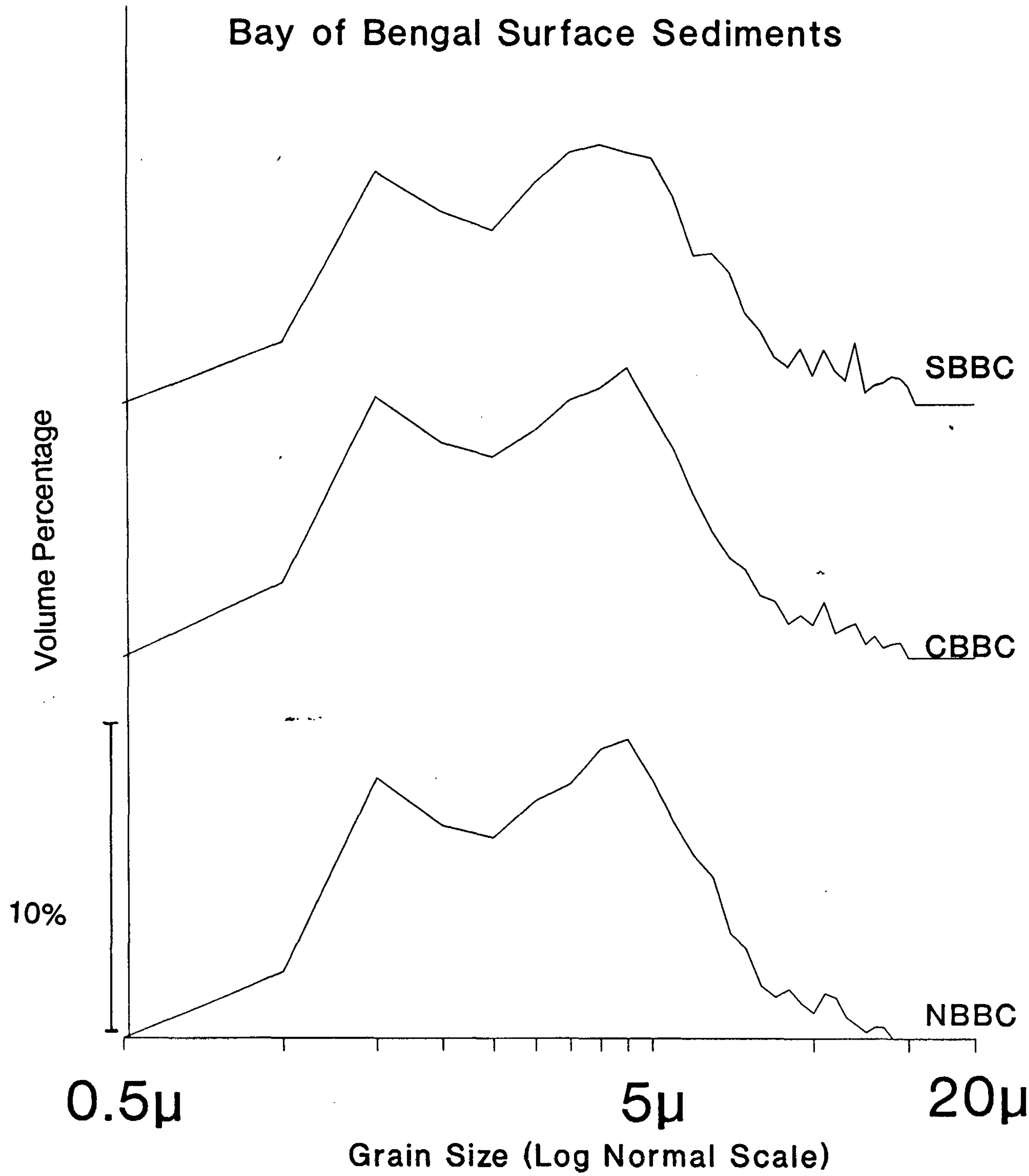


Fig. 7.12 Grain size distribution of the lithogenic fraction of core tops from the Bay of Bengal.

seem to be much variation from point of origin (near river mouths) to site of deposition. This shows that the Bay receives sediments predominantly form one source. In the Arabian Sea sediments there is considerable differences between the samples because the sediments here are composed of a mixture of fluvial and eolian sediments. The scatter between 14 to 20 micrometer is present in all the samples, particularly in the western Arabian Sea showing that the entire Arabian Sea receives eolian dust from the north African and Arabian Peninsula.

Chapter 8

Particle Fluxes, Mass Accumulation Rates and Benthic Fluxes in the Arabian Sea and Bay of Bengal.

8.1. Introduction

Particle fluxes in the oceans are dominated by biogenic material which undergoes extensive dissolution and degradation in the water column and at the sea floor. Variations in paleofluxes, as recorded in sediments, can be related either to productivity changes or degree of preservation. To distinguish between the two it is necessary to compare recent sedimentation rate with fluxes through the water column. Such comparisons are meaningful as no diagenetic or other post- depostional changes have occurred in sediment trap samples. As remineralization rates below 2000 m water depth are low it can be assumed that fluxes measured by the deep trap is similar to that reaching the sea floor. Reliable comparison can be made in areas where the rain rates and accumulation rates of the refractory elements are similar (*Dymond and Lyle, 1993*). Comparison of trap fluxes with accumulation rates in sediments can give us important information about biogeochemical cycling of elements. Of particular interest are the sites and rates of remineralization of particles (*Walsh et al, 1988*).

A limitation of this method is that deep sea sediments record fluxes averaged over the last 1000 years (due to bioturbation effects) while water column rain rates estimated from sediment traps are averaged at best over a few years. Such a comparison requires that no major changes in flux have occurred during the last few thousands of years. It is also assumed that sediment traps collect most of the particles settling through the water column and near-bottom transport of material is negligible.

8.2. Materials and methods

The present day mass accumulation rates of total particulate matter and first order components in the Arabian Sea were calculated by taking the weighted mean value of all measurements made in one year and multiplying by 365. Total and component fluxes in the Bay of Bengal were taken from *Ittekkot et al.* (*1991*) as data is available at three sites including the southern Bay of Bengal. Total mass accumulation rates in sediments during the Holocene have been calculated by multiplying sedimentation rates with the dry bulk density of core tops. Sedimentation rates and bulk dry density in the Arabian Sea and Bay of Bengal were taken from *Sirocko* (*1989*) and *Sarin et al.* (*1979*) respectively. Benthic fluxes were calculated as the difference between present day and Holocene mass accumulation rates.

8.3. Results and discussions

The present day rain rates of major components and their accumulation rates in sediments during the Holocene is shown in Table 8.1. Total rain rates measured by sediment traps were similar to mass accumulation rates in surface sediments at all three locations in the Arabian Sea and the northern Bay of Bengal and the offset between the two is less than a factor of 2. In the central and southern Bay of Bengal mass accumulation rates in surface sediments are less by a factor of 5. The reasons for this is not very clear. This cannot be explained by dissolution of carbonates as the sea floor lies well above the CCD (*Kolla, 1982*). Furthermore, ever the flux rates of lithogenic material alone is more than the total Holocene sedimentation rates. Another possibility is that sediments are getting resuspended and eroded. Though there are indication of strong bottom current activity in the Bay of Bengal (*Kolla, 1976b*) these currents are restricted only to the western part of the Bay. A more likely explanation is that the sedimentation rates reported by *Sarin et al.* (*1979*) are an underestimation. Sedimentation rates as low as 2

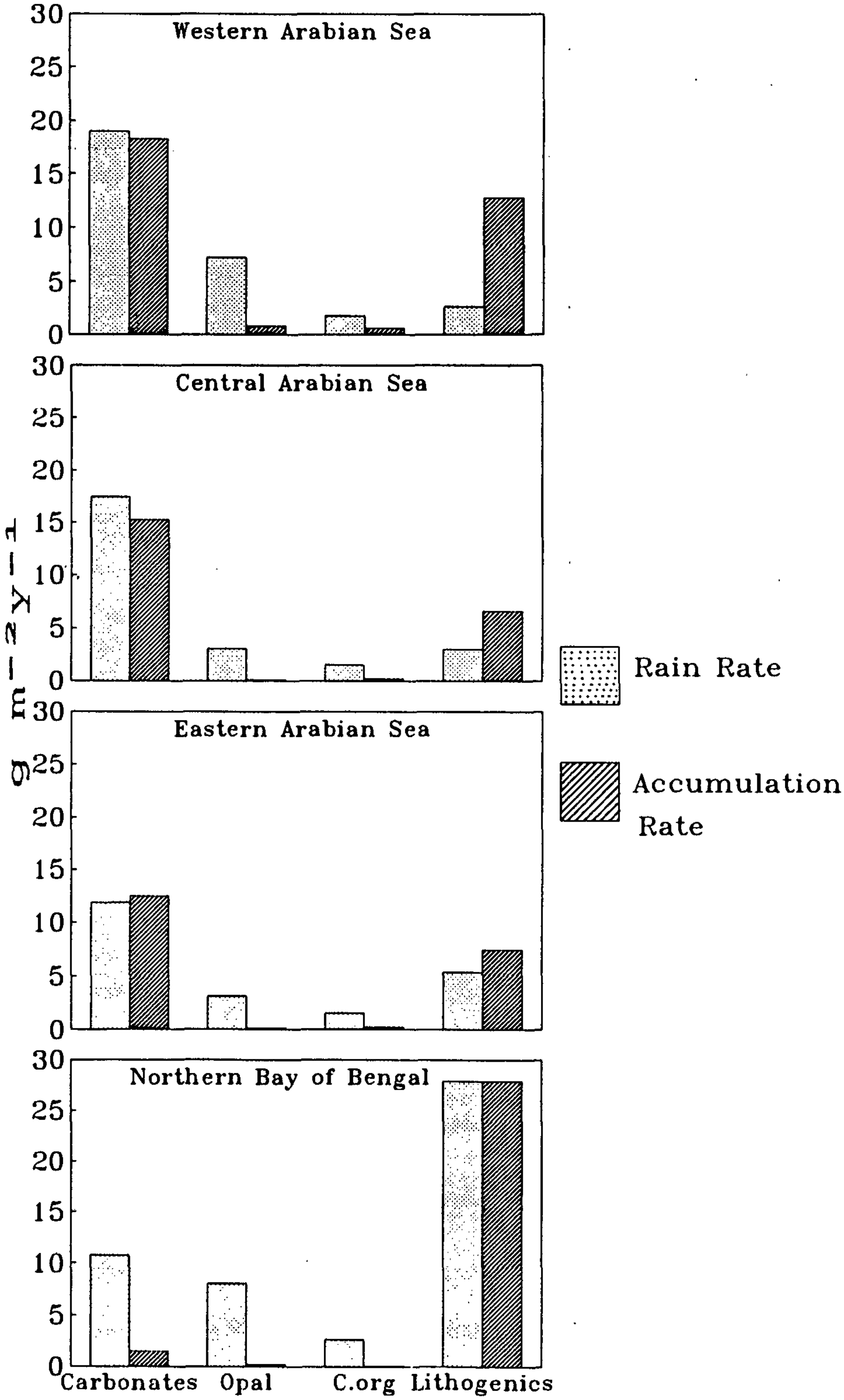


Fig. 8.1 Histogram showing the rain rate (measured by sediment traps) and accumulation rate in sediments of carbonates, opal, organic carbon and lithogenic material in the eastern, central and western Arabian Sea and northern Bay of Bengal. Most of the organic carbon and opal are remineralised at the sediment water interface.

Table 8.1. Comparison of total particulate and component fluxes measured by sediment traps with Holocene accumulation rates in surface sediments of the Arabian Sea and Bay of Bengal. Also shown are benthic fluxes for four sites in the northern Indian Ocean.

Station No	Lat	Long	Depth in m	Carb.	Opal	C.org	Lith.	Total
				All flux rates are in g m-2 y-1				
WESTERN ARABIAN SEA								
WAST	16. 20	60.30	3024	19.0	7.28	1.80	2.64	33.9
KL-71*	16. 14	60.15	4029	18.3	0.82	0.57	12.80	33.8
Benthic flux				0.7	6.36	1.23		8.3
CENTRAL ARABIAN SEA								
CAST	14. 29	64.46	2900	17.5	3.08	1.53	3.05	26.5
KL-15*	14. 53	64.45	3920	15.3	0.15	0.22	6.60	23.3
Benthic flux				2.2	2.93	1.31		6.4
EASTERN ARABIAN SEA								
EAST	15. 28	68.45	2770	11.9	3.13	1.56	5.40	23.6
KL-26*	15. 30	68.45	3776	12.5	0.10	0.21	7.50	21.1
KL-36*	17. 04	69.03	2055	12.9	0.07	0.19	5.50	18.5
Benthic flux				0.0	3.03	1.25		4.3
NORTHERN BAY OF BENGAL								
NBBT	17. 20	89.35	1727	10.7	8.06	2.65	27.96	51.6
30P#	17. 00	93.00	2403	1.5	0.20	0.02	27.90	30.0
Benthic flux				9.2	7.86	2.63		19.7
CENTRAL BAY OF BENGAL								
CBBT	13. 10	84.20	2282	16.92	8.42	2.61	14.70	44.8
38G#	13. 00	87.00	3076	n.d.	n.d.	n.d.	n.d	2.0
22PG#	11. 00	84.00	3456	n.d.	n.d.	n.d	n.d	5.6
SOUTHERN BAY OF BENGAL								
SBBT	06. 40	87.35	3006	18.5	7.27	2.04	8.56	38.1
44G#	04. 00	89.00	4057	n.d.	n.d.	n.d	n.d	5.1

Present day rain rates for Bay of Bengal from Ittekkot et al 1991.
*Holocene accumulation rates calculated from Sirokko 1989.
#Holocene accumulation rates calculated from Sarin et al 1979
n.d. = Not determined

mm per thousand years are generally reported only in areas far removed from continental influence (*Lisitzin, 1975*). Generally, accumulation rates of biogenic components were much lower in surface sediments showing that the sediment water interface is a site of intense remineralization, while lithogenic accumulation rates were similar or slightly higher.

8.3.1. Organic carbon fluxes and preservation

Organic carbon preservation in pelagic areas is low and average less than 5 percent of the rain rate to the sea floor (*Dymond & Lyle*, 1993). In sites less than 300 km from the coastline greater than 20% of the rain rate is preserved. Relatively high preservation is not uncommon near continental margins. For example, *Jhanke* (*1986*) observed approximately 50% preservation of organic carbon fluxes in the Santa Monica Basin.

Preservation of organic carbon or any other labile component can be calculated by the formula

Preservation = (C.org sed / C.org rain rate) X 100

Where C.org sed is the organic carbon accumulation rate in sediments and C.org rain rate is the annual flux.

The preservation of organic carbon in the eastern, central and western Arabian Sea is 11.4, 13.75 and 28.9 percent respectively. Among the three sites in the Arabian Sea organic preservation in the western Arabian Sea is almost double that of the central and eastern sites.

Dymond and Lyle, (*1993*) have proposed a model where preservation of organic carbon in the deep sea can also be a function of distance from shoreline . They have shown that at a distance of 400 km from the shore, organic carbon preservation is less than 10% and drops to less than 5% at a distance of 600 km. The present study shows that organic carbon preservation in the Arabian Sea is much higher compared to the open Atlantic or Pacific Ocean (*Dymond and Lyle, 1993*).

High apparent preservation of organic carbon in the Arabian Sea sediments could be due to i) deep lateral inputs of particulate organic matter ii) presence of refractory organic fraction near continents iii) or a coupling between sedimentary mass accumulation rates and the preservation of organic carbon fraction. High preservation rates may also be due to underestimation of organic carbon rain rates, due to dissolution of significant quantities of P, N and organic carbon in the collection cups. However, at depths greater than 2000 m the amount of soluble fraction is less than 20% and does not affect our results significantly. Lateral transport of organic matter from the high productive areas near continental margins can account for higher preservation in the deep sea. Since organic carbon fluxes decrease with depth in our samples lateral transport, if any, must be occurring below the deep traps.

Variation in sediment accumulation may affect the preservation of organic matter in sediments (*Heath et al., 1977; Muller and Suess, 1979*; and *Heinrichs and Reeburgh, 1987; Sirocko and Ittekkot, 1992*). the higher organic carbon preservation in the western Arabian Sea could be due to the higher mass accumulation rates. *Sirocko and Ittekkot* (*1992*) found a high correlation between organic carbon and bulk accumulation rate (r=0.91) in the sediments of the Arabian Sea. Emerson has argued that residence time of organic carbon at the sediment water interface (about 100 years) is too short for sediment accumulation rate to have any effect on the organic carbon preservation. However, if the organic matter fluxes to the sediment has a significant refractory component are is partly derived from land, it will be sensitive to sedimentation rate. *Reemtsma et al.* (*1989, 1992*) have found that refractory terrestrial organic matter is being transported to the northern Indian Ocean from the adjacent continental areas.

8.3.2. Calcium Carbonate fluxes and preservation

Calcite preservation in sediments is most strongly affected by depth dependant changes in calcite saturation in the oceans. In the Indian ocean the depth of the CCD is about 4200 m, hence calcite preservation in the sediments below the traps should be very high. In the Arabian Sea between over 80 % of the calcite flux is preserved in the sediments. However, in the northern Bay of Bengal only 14 percent of the carbon appears to be preserved. This may be related to the low C.carb./C.org ratios in settling particle (Table 4.5). The production of carbon dioxide from organic matter decomposition will enhance calcium carbonate dissolution within sediments (*Emerson and Bender, 1981*). The C.carb/C.org ratios in the deep traps of the northern Bay of Bengal is only 0.48 compared to 1.4 in the central Arabian Sea.

8.3.3. Opal fluxes and preservation

The percentage of opal preserved in sediments range between 2.4 to 10.3. *Brocker and Peng* (*1982*) have proposed a opal dissolution model in which rain rate of opal provides an important control on the degree of preservation. In this model sites with low opal rain rates experience a large fractional loss of the rain of opal as compared to sites with high rain rate of opal. According to this model primary rain rates of opal are enhanced by dissolution effects. This is most probably the reason why the gradient of opal percentage is so sharp at upwelling boundaries.

The Brocker and Peng model holds good for the Arabian Sea but not for the northern Indian Ocean as a whole. In the Arabian Sea highest percentage of opal preservation (10.3%) is in the western Arabian Sea where the opal rain

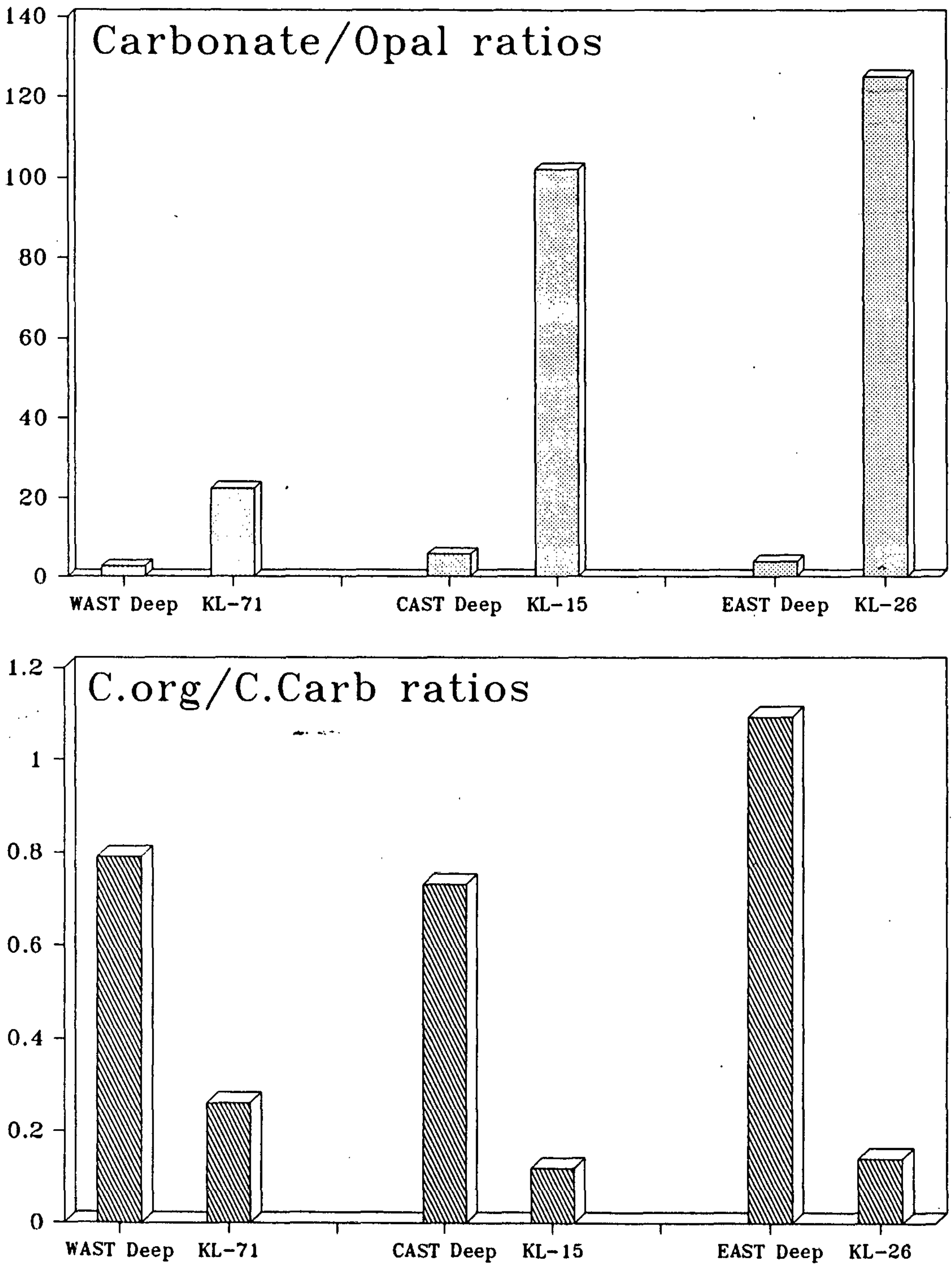


Fig. 8.2. Carbonate/Opal and Organic carbon/Carbonate carbon ratios in settling particles and core tops in the eastern, central and western Arabian Sea. Sharp changes in the ratios between the deep traps and surface sediments show that rapid remineralisation occurs at the sediment water interface.

rate (8.0 g m^{-2} y^{-1}) is the highest. However, preservation of opal in the northern Bay of Bengal is comparatively much less (2.4 %) even though it has a slightly higher opal rain (8.4 g m^{-2} y^{-1}). Variation in the rate of dissolution of opal is also significantly affected by the nature of opaline debris which range from robust radiolarian and diatom forms to easily soluble silicoflagellates. Visual examination shows a higher percentage of robust colonial radiolarians in the Arabian Sea samples.

8.3.4. Lithogenics fluxes and preservation

Lithogenic material undergoes little dissolution or transformation between their entry into the oceanic system and burial in sediments. For this component the burial rate should equal rain rate. However, the rain rate of lithogenic matter in the Arabian Sea is consistently less than sedimentation rates showing that additional material is transported and deposited through near bottom processes like, for example, nepheloid layer flows. Most of the sites in the northern Indian Ocean show increase in lithogenic flux with depth due to aggregation of fine clays by organic matter. Probably the rate of aggregation is much higher in the nepheloid layer. Rebound particles, i.e those particles which are part of the primary flux but get resuspended (*Walsh, 1985*) can get transported down slope and scavenge significant quantities of aluminosilicates especially in semi-enclosed basins like the Arabian Sea and Bay of Bengal. Some differences could also be due to inefficient collection of the sediment traps.

The eastern Arabian Sea receives sediments predominantly from the Indus. Construction of dams and barrages during the last 3 decades has reduced the suspended sediment discharge of this river significantly from about 400 million tonnes to less than 50 million tonnes per year (*Milliman et al., 1984*).

This may be responsible for the present low rain rates in the eastern Arabian Sea.

The largest differences in rain rates and burial rates are for the western Arabian Sea. This area receives terrigenous material predominantly from eolian dust. Even slight shift in wind intensity or direction can show significant interannual variability in eolian material rain rates. Moreover, the winds during the collection period were of lesser intensity due to the 1986-87 El Nino.

8.3.5. Benthic fluxes

Table 8.1 also shows the benthic fluxes i.e. amount of material which reaches the sea floor and is remineralised. It is seen that areas of high fluxes are also areas of high benthic fluxes. In the Arabian Sea about 15% of the carbonates, more than 87% of biogenic silica and 70% of organic carbon is remineralized at the sediment water interface. In the northern Bay of Bengal benthic fluxes are even higher with 85% of the carbonates and 95% of the biogenic silica and organic carbon is remineralised at the sediment water interface.

The sea floor is a site if intense remineralization of particles settling from above. Among the major components, dissolution rates of opal and organic carbon on the sea floor is much higher compared to carbonates or lithogenic matter. Fig. 8.2 shows that the carbonate/opal ratios in the sediments is much higher than that of settling particles. This means that opal is being remineralised at the sea floor 20 to 50 times more rapidly that carbonates. Similarly organic matter remineralization rates at the sea floor ar much higher than carbonates. Elemental carbon therefore has a much higher chance of being preserved in sediments (above the CCD) if it settles as carbonates than as organic matter.

Chapter 9

Major Conclusions

4.1. Total particle fluxes in the Arabian Sea and Bay of Bengal show a strong seasonality with high fluxes during the monsoon and low fluxes during the intermonsoon period. In the Arabian Sea, particle fluxes almost ceases at the end of the SW monsoon whereas in the Bay of Bengal moderate fluxes are maintained throughout the year.

4.2. Monsoon winds influence primary productivity and particle fluxes in the Arabian Sea by introducing nutrients in the photic zone from deeper waters. In the eastern and western Arabian Sea high particle fluxes are partly due to advection of nutrient rich waters together with lateral advection of particles from upwelling regions. In the central Arabian Sea high fluxes during the monsoon are due to increase in nutrient concentrations as a result of mixed layer deepening.

4.3. Nutrient inputs from adjacent rivers together with turbulent mixing at the base of the halocline is responsible for increase in the primary productivity and particle fluxes in the Bay of Bengal.

4.4. Component analysis show that particle fluxes in the Arabian Sea is carbonate dominated. During the early SW monsoon period there is a coccolith (carbonate) bloom followed by a diatom (opal) bloom during the later part of the monsoon.

4.5. Biogenic particle fluxes in the Bay of Bengal are also carbonate dominated but have low carbonate/opal ratios. Carbonate fluxes increase while opal and lithogenic fluxes decrease away from river mouths. Freshwater discharge alters the biological community structure in the photic zone with silica producers dominating during high river discharge periods.

4.6. Organic carbon fluxes do not match primary productivity patterns at all stations in the northern Indian Ocean. Organic carbon fluxes in the southern Bay of Bengal, a low productive area, are similar to that of the western Arabian Sea, an area of known high productivity.

5.1. The northern Indian Ocean is an area having very high lithogenic fluxes due to high suspended and eolian sediment input into the area.

5.2. In the Arabian Sea there is a decrease in lithogenic fluxes from east to west due to clockwise surface water circulation during the southwest monsoon period which diverts river discharge to the eastern part of the sea.

5.3. In the Bay of Bengal a north-south decrease in lithogenic flux is seen with fluxes decreasing away from river mouths.

5.4. In areas dominated by fluvial input lithogenic fluxes increase with depth due to scavenging of mineral matter by settling organic aggregates. The rate of scavenging is higher during high flux periods. In the Arabian Sea lithogenic fluxes does not seem to increase with depth as minerals are introduced mainly at the sea surface. More data is required to confirm this.

5.5. Biological control on lithogenic fluxes is well seen at all the sites in the northern Indian Ocean. Scavenging of abiogenic particles by organic aggregates is one of the important processes of lithogenic matter sedimentation. Up to 60 per cent of the lithogenic matter settling on the sea floor may be scavenged by organic aggregates.

5.6. Higher lithogenic percentages in marine snow or fecal pellets should enhance their sinking rates as the density of lithogenic particles is higher than organic matter. Lithogenic matter in organic aggregates may therefore serve as a ballast and help remove organic matter from the sea surface thereby increasing the drawdown of atmospheric carbon dioxide into the ocean.

6.1. Illite is the major mineral at all the trap locations, indicating that physical processes of weathering dominate in the source area of minerals to the northern Indian Ocean.

6.2. Smectites are relatively high during the southwest monsoon period in traps located near the Indian Peninsula. This is due to high rainfall in the Deccan Trap and adjacent areas during the this period.

6.3. Palygorskite is found in the traps at all three sites in the Arabian Sea indicating that eolian dust from Arabia and north Africa reaches even the eastern Arabian Sea.

6.4. In the Bay of Bengal, higher illite fluxes in the shallow traps indicate that the suspended load of the Ganges-Brahmaputra river discharge is transported towards the south in the surficial low saline layer. Higher smectite content in the deep traps indicate lateral advection at depth of material from the eastern continental margin of India.

6.5. In the Bay of Bengal there is a marked change in clay mineral composition of settling particles with depth. This shows that part of the suspended lithogenic matter advected by the prevailing currents is converted to vertical flux by organic aggregates.

7.1. Grain size distribution pattern of the lithogenic fraction of settling particles and surface sediment in the northern Indian Ocean show that most (over 99%) of the particles have a diameter less than 20 micrometer indicating deep water deposition in a quiescent environment.

7.2. Grain size distribution pattern of the Arabian Sea samples suggest a mixture of eolian and fluvial particles. Samples from this area have a higher percentage of eolian particles in the 14 to 20 micrometer range.

7.3. Grain size distribution pattern of the Bay of Bengal samples indicate a predominantly fluvial source as less than 1 % of the particles are coarser than 14 micrometer.

7.4. Most samples show a bimodal distribution with peaks at 1.5 and 4.5 micrometer. The modes are probably related to clay mineral composition of the lithogenic fraction. Samples having high smectite content have a higher percentage of fine particles in the 1 to 2 micrometer range.

8.1. The rain rate of particles are similar to accumulation rates in sediments at all three sites in the Arabian Sea and in the northern Bay of Bengal. In the central and southern Arabian Sea there are major discrepancies mostly due to underestimation of sedimentation rates.

8.2. Lithogenic particles enhance organic carbon fluxes but preservation of organic matter in sediments depends more on mass accumulation rates.

8.3. Calcium carbonate preservation in the Bay of Bengal seems to be low due to high organic matter fluxes.

8.4. Preservation of opal in the Arabian Sea is high due to episodic fluxes of opal and presence of more robust forms of diatoms and radiolarians.

8.5. Lithogenic rain rates in the Arabian Sea are lower than accumulation rates in sediments. This indicates i). Additional source of lithogenic particles below the deep traps, like for example the nepheloid layer ii). Reduced lithogenic input to the Arabian Sea due to construction of dams and barrages across rivers and iii) Reduced lithogenic input and fluxes during the year the measurements were made (1986-87) due to the El Nino effect.

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2. NAIR, R.R., ITEKKOT, V., MANGANINI, S.J., **RAMASWAMY, V**., HAAKE, B., DEGENS, E.T., DESAI, B.N., HONJO. S., 1989, Increased particle fluxes to the oceans related to the monsoons: **Nature**, V. 338, P. 749-751.
3. **RAMASWAMY, V**. and NAIR, R.R., 1989, Lack of cross shelf transport on the western margin of India: Evidence from clay mineralogy: Journal of Coastal Research. V.5, pp. 541-546
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5. ITTEKKOT, V., R.R.NAIR, S.HONJO, **V.RAMASWAMY**, M.BARTSCH, S.MANGANINI, and B.N.DESAI, 1991. Enhanced particle fluxes to the deep ocean induced by freshwater inputs. **Nature**, 351, 385-387.
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