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PREFACE

The purpose of the Intergovernmental Oceanographic Commission is "to promote scientific investigation with a view to learning more about the nature and resources of the oceans through the concerted action of its members" (Article 1, paragraph 2, of the Statutes).

To achieve this purpose, the Commission has developed, alone or jointly with other organizations of the UN system, scientific programmes in five main fields: Ocean Dynamics and Climate; Ocean Science in Relation to Living Resources; Ocean Science in Relation to Non-Living Resources; Ocean Mapping; and Marine Pollution Research and Monitoring.

To support these programmes, the Commission also co-ordinates four ocean services: the Integrated Global Ocean Services System; International Oceanographic Data and Information Exchange; the Global Sea-level Observing System; and the Tsunami Warning System in the Pacific.

The main mechanisms for planning, promoting and co-ordinating the implementation of these programmes and the related services are Scientific/Technical Committees, which deal with the global aspects, and Regional Sub-Commissions or Committees, which deal with the implementation of specific regional projects and of regional aspects of IOC's major global programmes.

Two such Regional Committees are concerned specifically with the Indian Ocean region: the Regional Committee for the Co-operative Investigation in the North and Central Western Indian Ocean (IOCINCWIO), and the Regional Committee for the Central Indian Ocean (IOCINDIO).

The Regional Committee (originally Programme Group) for the Central Indian Ocean was created by the IOC Assembly at its Twelfth Session (November 1982) through Resolution XII-14 which was sponsored by Bangladesh, India, Pakistan and Sri Lanka. The Assembly recognized that a better understanding of the major oceanographic processes, particularly as they relate to climate and living resources in the region, was needed. It recalled the active co-operation that had existed amongst the Member States of this region and other interested Member States from outside the region, during the International Indian Ocean Expedition (IIOE), and the importance of relating the major ocean processes in this part of the Indian Ocean with those in the western part, covered by IOCINCWIO, and with the Regional Committee for the Western Pacific (WESTPAC).

With a view to generating a preliminary basis for a future programme of work for IOCINDIO, an IOC-Unesco Workshop on Regional Co-operation in Marine Science in the Central Indian Ocean and Adjacent Seas and Gulfs was convened in Colombo, Sri Lanka, at the kind invitation of the Government of Sri Lanka, from 8 to 13 July 1985. Thirty-one scientists from 15 Member States participated and produced some fourteen proposals for regional co-operative marine scientific research, many in the form of fairly detailed project outlines. They were: (i) Coastal Water Dynamics; (ii) Water-mass Movements; (iii) Storm-surge Prediction for the Marginal Seas of the Northern Indian Ocean; (iv) Distribution of Oceanic Pelagic Resources (with Particular Reference to Tuna) in Relation to the Environment in the Indian Ocean Region; (v) Oceanic Environment, Pelagic Bioproductivity and Living Resources of the Northern Arabian Sea; (vi) Investigations of Exploitable Demersal Living Resources of the Deep Shelf and Bathyal Zones; (vii) Survey of Coral Resources and Strengthening of Regional Co-operation in Training and Data Exchange; (viii) Red Tides in the Central Indian Ocean and Adjacent Seas and Gulfs; (ix) Taxonomic Research on Marine Fauna and Flora of the Indian Ocean; (x) Survey of Continental Shelves; (xi) Coastal Processes; (xii) Effects of Damming Rivers; (xiii) Deep-water Sediments; (xiv) Studies of Crustal Structure, Tectonics and Geological Evolution. The Workshop also recommended the promotion of a regional component of the IOC Marine Pollution Monitoring (MARPOLMON) System and a study of the desirability of preparing detailed (large-scale) bathymetric charts.

Several of the Projects listed above were developed into Proposals and put to the IOC Regional Committee for the Central Indian Ocean at its First Session, held in Islamabad, Pakistan, 3-7 July 1988, at the kind invitation of the Government of Pakistan, with a view to their adoption as the major elements of the Regional Committee's Programme of Work.

The IOC particularly thanks the scientists who have been involved in marine research in the region, for their kind collaboration in developing these proposals at and since the Workshop in Colombo, and acknowledges with appreciation the editing of this Supplement by Dr. John Milliman of the Woods Hole Oceanographic Institution.

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INTRODUCTION

The Indian Ocean differs from the Pacific and Atlantic oceans by the presence of a continuous land mass (Asia) to the north and the resulting lack of communication with the Arctic. This has resulted in the Indian Ocean having a unique meteorological and oceanographic climate. The monsoon climate incites a dramatic seasonal shift in wind patterns and a corresponding change in oceanic circulation. The lack of formation of an Arctic deep water means that circulation in the deeper areas of the Indian Ocean is quite distinct from the deep-water circulation patterns in the Pacific or Atlantic.

Despite these differences, the central Indian Ocean, in many respects, remains perhaps the most unstudied large section of the world oceans. The first two major expeditions in the Indian Ocean (Challenger and John Murray expeditions in 1872-75 and 1932-33, respectively) were primarily the result of planning and participation by British scientists, and the International Indian Ocean Expedition (1959-65) was largely organized and carried out by developed countries from outside the region. To date, little regional research has resulted from the direct discussion and interaction by the countries bordering to the Indian Ocean.

Partly to encourage such local participation in Indian Ocean research, the Intergovernmental Oceanographic Commission (IOC) convened the IOC-Unesco Workshop on Regional Co-operation in Marine Science in the Central Indian Ocean and Adjacent Seas and Gulfs in Colombo, Sri Lanka, from 8 to 13 July 1985. The other important task of the Workshop, and one that is addressed in this supplement to IOC Workshop Report No. 37, was to provide a forum in which the oceanographic regime of the central Indian Ocean and important problems concerning this regime could be discussed. The main body of this supplement presents 15 full papers and three abstracts discussing various aspects of oceanography in the central Indian Ocean and adjacent seas and another four abstracts of talks presented at the Workshop.

In this introduction, topics that the participants considered to be the critical areas of future research in the central Indian Ocean and adjacent seas are briefly discussed. The topics covered are: Ocean Dynamics and Climate, Ocean Sciences and Living Resources, Ocean Sciences and Non-living Resources, Marine Pollution, and Future Co-operative Studies. In nearly all of these fields, it is obvious that co-operative research between two or more Indian Ocean nations would greatly increase the effectiveness of these studies.

Ocean Dynamics and Climate

1. Coastal water dynamics - National and international efforts are possible in this area, and well within the capabilities of most nations. Not only are studies of circulation and dynamics necessary to the understanding of regional processes, but they also are necessary for the effective management of coastal resources and of marine pollution. In many areas around the Indian Ocean there are serious data gaps, particularly in mean sea level, variation in the thermal field and currents, etc.

2. Water-mass movement - Very little is known about the mixing of Red Sea and Gulf waters with the deeper waters of the Arabian Sea, particularly in relation to monsoonal variations. Chemical parameters and the causes of their specific limits (e.g., oxygen minimum) are important in the identification of these water masses.

3. Storm surge prediction for the marginal seas of the northern Indian Ocean - Few natural disasters are as devastating as the storm surges that periodically strike the Bay of Bengal and the Arabian Sea. In the 110-year period from 1876 to 1985, at least 800000 people have been killed and countless more affected by injuries and the loss of personal property. Adequate storm-surge prediction requires not only local expertise but increased tidal and wind-field measurements, as well as the development of more sophisticated prediction models.

Ocean Sciences and Living Resources

1. Oceanic pelagic resources - Knowledge of the distribution of pelagic resources is needed, particularly with respect to tuna. This will require a better knowledge of the oceanographic regime, as well as more data on the spatial and temporal distribution of tuna eggs and larvae.

2. Environment and productivity of living resources in the northern Arabian Sea - Strongly influenced by the monsoonal climate, the northern Indian Ocean seasonally exhibits very high rates of biological productivity. The potential resources of epipelagic fish (such as sardines and mackerals), however, are largely underexploited. The seasonal productivity and associated upwelling also can produce locally almost anoxic intermediate waters which can severely impact the ambient environment.

3. Exploitable demersal resources in deep shelf and bathyal waters - On shelves, shrimp and coral reef fish provide potentially exploitable resources. The local environmental characteristics, such as oxygen levels, currents, etc., however, can affect these populations in ways still undocumented let alone understood.

4. Environment and living resources of the Red Sea - As noted in a number of papers in this volume, the hypersaline and often hyperthermal Red Sea has unique biological populations and biomasses about which better documentation is needed.

5. Coral-reef resources - Coral reefs and associated environments often are highly productive in terms of plants, invertebrates and fish. This has economic impact in terms of commercial fisheries (e.g., fish, shrimp and mollusks) and tourist attractions. Moreover, reefs and associated mangrove areas can serve as major barriers against the waves and surge created by major storms. Unfortunately, the impacts on the reef environment by present-day human activities (see the paper by Ormond) have led to the local demise of many reefs and threaten to curtail the survival of many more. Mapping reefs and studying the habitats and species could provide the basis on which the reefs and possible degradation can be documented.

6. Red tides - Toxic dinoflagellates and (to a lesser extent) blue-green algae produce neurotoxins that can find their way to man through infected fish and shellfish. At this point, too few data are available to know the actual impact of red tides on the human population of the Indian Ocean region, but as the consumption of fish increases, various diseases due to red tides probably will increase. Local authorities and public health authorities should be aware of the causes and symptoms of these various diseases.

7. Taxonomic research of marine flora and fauna - Some Indian Ocean scientists lack access to the taxonomic skills needed to recognize and classify tropical marine biota, but such skills clearly are required if marine biology is to be developed in these countries.

Ocean Science and Non-Living Resources

1. Survey of continental shelves - Because many of the potential non-living resources are found in water depths less than 200 metres (e.g., sand and gravel, heavy minerals, phosphate), this environment should be surveyed systematically by all coastal nations.

2. Coastal processes - Waves, tides and storms will play a major role in the physical stability of nearshore areas. Moreover, if these processes are understood, then the movement and deposition of pollutants can be predicted.

3. Effects of damming rivers - At present, rivers draining into the Indian Ocean carry large loads. These loads, however, will diminish as the rivers are dammed in response to increased economic development. Such decreases in sediment loads may result in marked increases in local shoreline erosion.

4. Deep-water sediments - Variation of sediments beneath the sea floor reflects the evolving nature of the Indian Ocean basin and its climate. By understanding these changes in time, possible future changes also can be predicted.

5. Crustal structure, tectonics and geological evolution - In contrast to the previous item, which can require cores and laboratory analyses, the structure and evolution of the ocean basin can only be defined by means of expensive geophysical equipment. Realistically, such studies are best carried out in concert with other neighboring countries.

Marine Pollution

At present, marine pollution is not perceived as a major problem in the Indian Ocean, certainly not as much as in impacted areas such as the Mediterranean and Baltic Seas. Yet the development of large industries (e.g., tin mining and petroleum exploration) can accelerate local pollution, and the occurrence of oil slicks related to ship traffic is probably increasing. Perhaps more disturbing is the potential loss of coral reefs and mangrove forests which still have considerable unexploited economic potential. Increased awareness and monitoring is required; coastal nations increasingly must co-operate in regional documentation of pollution and its effects.

Future Co-operative Studies

All of the foregoing critical fields of study in the northern Indian Ocean and adjacent marginal seas clearly show the need for interdisciplinary and international (at least inter-regional) studies. Understanding the living or non-living resources, for example, requires a knowledge of the physical and chemical oceanography as well as the biology and geology of the area. Successful studies will require co-operation on several levels: (1) Data exchange; (2)Training; (3) Monitoring; and (4) Interactions with international programmes such as the IOC Regional Committees for IOCINCWIO (Co-operative Investigation in the North and Central Western Indian Ocean) and for WESTPAC (Western Pacific).

For a more detailed discussion of the findings of the various working groups at the Colombo workshop, the reader is referred to IOC Workshop Report No. 37, IOC-Unesco Workshop on Regional Co-operation in Marine Science in the Central Indian Ocean and Adjacent Seas and Gulf.

MARINE ENVIRONMENT OF THE INDIAN OCEAN

THE INDIAN OCEAN --- AN ENVIRONMENTAL OVERVIEW[1]

R. SEN GUPTA AND S.Z. QASIM

National Institute of Oceanography
Dona Paula, Goa-403 004, India

Department of Ocean Development
Government of India Mahasagar Bhavan
New Delhi, India

INTRODUCTION

During the International Indian Ocean Expedition (IIOE), 1962-65, the Indian Ocean was defined as the area between 25°N and 40°S latitudes and 45°E and 115°E longitudes. This region included all the straits, adjacent seas and semi-enclosed basins within the littoral countries, which, because of their special characteristics, form a unique study area.

In the present review, the Indian Ocean we discuss is an area bounded by 25°N and 30°S and 35°E and 98°E. Geographically, it is the area from 30°S latitude to the Gulf of Oman and the head of the Bay of Bengal and from the East African coast to the coastlines of Burma, Thailand and Malaysia (excluding the Strait of Malacca) (Fig. 1). In this region of the Indian Ocean, there are 19 countries, comprising a total area of 9.6×10^6 km^2. They are inhabited by about 1220 million people, with an average population density of 127/km^2. Thus, an average 22.5% of the world population lives in 18.6% of the total land area. The 8 countries in the East African region (of which 4 are island countries), however, have a population density of only 16.5/km^2.

OCEANOGRAPHIC FEATURES

From the data collected during the IIOE, a comprehensive atlas of physical and chemical features was published by WYRTKI (1971). The various characteristics can be summarized briefly as follows:

PHYSICAL

The Indian Ocean occupies an area of 74.92×10^6 km^2 including the marginal seas (DIETRICH, 1963), representing 20.8% of the total world oceans. Average depth is 3873 m. According to BUDYKO (1972), the Indian Ocean has a negative water balance: Annually it receives 6000 km^3 of river runoff and 88,000 km^3 of precipitation while its evaporation is 103, 000 km^3.

However, there are some parts of the Indian Ocean in which the water balance is positive. One such area is the Bay of Bengal in the northeast, which receives an annual precipitation of 11,000 km^3 and a runoff of 1730 km^3. Assuming the limit of dilution at 25 m depth, these volumes or precipitation and runoff would dilute the surface water by about 5% (SEN GUPTA et al., 1978a). The Arabian Sea, in the northwestern part of the Indian Ocean, in contrast, is a region of negative water balance. It receives annually a precipitation of 10,000 km^3 and runoff of 525 km^3 from the rivers of India and Pakistan. Thus, these two areas, although covering only about 13.8% of the total area of the Indian Ocean, receive 37.6% and 23.9% of total precipitation and runoff, respectively. In fact, the Bay of Bengal and the Arabian Sea together occupy only 3% of the world oceanic area but receive 9% of the global river runoff.

[1] modified from a paper presented in The Oceans - Realities and Prospects, edited by R.C. Sharma, Rajesh Publ., New Delhi.

The northern boundary of the Indian Ocean receives waters of high salinity from the Arabian Gulf and the Red Sea. The Gulf water flows through the Hormuz Strait which has a maximum sill-depth of about 100 m (DUING and SCHWILL, 1967), while the Red Sea water flows through the Strait of Bab-el-Mandab with a maximum sill-depth of 125 m. These are the only two subsurface water masses in the Indian Ocean that originate from the north. The presence of the Gulf and the Red Sea water masses has been traced to the equator and into the western part of the Bay of Bengal. The Gulf water has a characteristic sigma-t of 26.5 that can be identified from a depth range 100 to 300 m in the Arabian Sea to about 500 m in the equatorial Indian Ocean (Fig. 2). The water from the Red Sea can be traced by its characteristic sigma-t of 27.1 between 400 and 900 m depth in the Arabian Sea and to about 1000 m depth in the equatorial Indian Ocean (ROCHFORD, 1964; WOOSTER et al., 1967).

Two other surface water masses are formed in the northern Indian Ocean. One is a high salinity surface water in the northern and central Arabian Sea formed due to excess of evaporation; it has a sigma-t of 25 (ROCHFORD, 1964) and occurs to a depth of about 125 m. Due to an excess of precipitation and a very large volume of runoff, a low salinity surface water mass forms in the Bay of Bengal. This water mass flows south of Sri Lanka to the west and can be traced over a considerable distance into the Arabian Sea (WYRTKI, 1973).

IVANENKOV and GUBIN (1960) termed the water mass between 100 and 300 m with a sigma-t of 27.6 as the North Indian Ocean Deep Water, and the water mass below 3000 m with a sigma-t 27.8 as the North Indian Ocean Bottom Water (Fig. 2). These two water masses are formed as a result of mixing of the Antarctic Deep and Circumpolar water masses. According to WOOSTER et al. (1967) water masses in the northern Indian Ocean below 1000 m depth originate in the Antarctic circumpolar region and enter the northern Indian Ocean from the southwest, spreading out in both the central and eastern basins. They can be identified by their characteristic sigma-t values. In addition to the water masses, broadly classified above, there are several other water masses formed due to mixing which are not described here due to limited space.

The Indian Ocean has an N-S asymmetric shape due to the presence of the Asian continent. The result is that this ocean is separated from the deep-reaching vertical convection areas of the northern hemisphere. Such an asymmetric configuration leads to a weak circulation and poor renewal in the deep northern Indian Ocean. In addition, it is strongly influenced up to 10°S by the characteristic atmospheric monsoon circulation (DIETRICH, 1973): Surface circulation in the north reverses every half year, from northeast during the winter monsoon to southwest during the summer monsoon. This phenomenon makes the Indian Ocean a very suitable area for the study of ocean-atmosphere interactions.

During the northeast monsoon (November-April) the oceanic circulation north of the equator remains moderately strong and is east to west. It penetrates beyond the thermocline. Due to shoaling of the thermocline, a weak upwelling has been identified in the northeastern part of the Indian Ocean. The northeast monsoon is quite strong in the Bay of Bengal, in which the enormous volume of river runoff probably compensates for the weak upwelling of the northeastern region.

During the southwest monsoon, the South Equatorial current becomes stronger and subsequently forms the East African Coastal Current. It flows northward, ending up as the Somali Current which frequently attains velocities as strong as those of other western boundary currents, e.g., the Gulf Stream or the Kuroshio (WARREN et al., 1966). The monsoon current, the South Equatorial Current and the Somali Current form a strong wind-driven gyre in the equatorial Indian Ocean (WYRTKI, 1973). This gyre then breaks down in a series of large gyres in the Arabian Sea (DUING, 1970). The resultant upwelling along the Somali coast is most intense between 5°N and 11°N, where the entire warm surface layer is removed and subsurface waters less than 20°C reach the sea surface (WARREN et al., 1966).

Strong winds also blow parallel to the Arabian Coast, east of 55°E, during the southwest monsoon and cause upwelling. This upwelling is different from the Somali upwelling because no strong current develops. During the southwest monsoon season weak upwelling has been reported from the east coast of India (VARADACHARI and SHARMA, 1967). Along the west coast of India, however, subsurface water comes very close to the surface, which has been interpreted as upwelled water (BANSE, 1968).

The subtropical anticyclonic gyre includes the South Equatorial Current, the Agulhas Current and those portions of the westwind drift, which lie north of the subtropical convergence, close to 40°S.

The hydrographic structure of the Antarctic waters flowing into the Indian Ocean is very similar to those in the Atlantic and the Pacific. It is governed by the surfacing of the main thermocline and by the strong deep-reaching Circumpolar Current forming the major bulk of the Indian Ocean deep and bottom water flowing northward.

CHEMICAL

The nutrient properties of the various water masses in the Indian Ocean aregoverned largely by the monsoon winds blowing over the northern Indian Ocean. A detailed review of all the data collected during the IIOE has been published by McGILL (1973). Some general features of the chemical characteristics are as follows:

Because of a sluggish circulation and weak renewal, a thick layer with very low oxygen concentration extends from about 100 m to 1000 m almost all over the northern Indian Ocean. The oxygen content of this water very often is less than 0.5 ml l^{-1}. Off the Indian coast values lower than even 0.05 ml l^{-1} have been reported. This feature of the Indian Ocean is very similar to that of the eastern tropical Pacific Ocean.

The monsoon gyre in the northern India Ocean is marked by high PO_4^{-3} - P contents, sometimes associated with low oxygen concentrations, due to upwelling. Phosphate concentrations in the region of Somali upwelling reach values greater than 1μ mol l^{-1} at the surface. For the southwestern Indian Ocean the range is from 0.25 to 0.54μ mol l^{-1}, increasing with depth to a maximum between 300 and 500 m. Concentrations remain more or less constant throughout the rest of the water column, decreasing slightly near the bottom.

Values for NO^{-3} - N indicate similar features, with higher values at the surface during the upwelling period. Generally, NO^{-3} - N concentrations range from 1μ mol l^{-1} at the surface to about 50μ mol l^{-1} between 3000 and 4000 m. But during the periods of high photosynthetic productivity, the surface values quite frequently reach zero.

Marked changes in nutrient concentrations occur at 10°S. The monsoon gyre in the northern Indian Ocean is marked by high PO_4^{-3} - P concentrations, and within this region the concentrations range from less than 0.4μ mol l^{-1} to more than 1μ mol l^{-1}. In the areas of the sub-tropical gyre nutrient values are very low.

RYTHER and MENZEL (1965) concluded that the general levels of all the nutrients in the western Arabian Sea are about twice as high as those of the corresponding depths of the North Atlantic.

Subsequent to the IIOE much work has been done to study the quantitative aspects of the nutrient chemistry in the northern Indian Ocean (SEN GUPTA et al., 1975, 1976a and b, 1977, 1979, 1980b; NAQVI et al., 1978, 1982; DEUSER et al., 1978). These studies have revealed that about one-third of the dissolved nitrate-nitrogen is lost during denitrification in the oxygen-poor layer of the northern Arabian Sea; this process can be traced to 2°N. Occasionally, two nitrite-nitrogen maxima and two oxygen minima are observed in this region. The depths of the maximum intensity of denitrification invariably coincide with the secondary nitrite maxima (Fig. 3). The rate of denitrification is 200 g m^{-2} in the waters of the northernmost part of the Arabian Sea, where renewal of deeper water is very slow due to the sluggish movement of water (QASIM, 1982). The rate is 70 g m^{-2} at 15°N (SEN GUPTA et al., 1980b), and 45 g m^{-2} at 12°N (NAQVI et al., 1982). This is equivalent to 5-10% of the global rate of denitrification required to maintain a steady state in the distribution of nitrogen on the surface of the earth (EMERY et al., 1955).

Assuming that only about 10% of the photosynthetic productivity reaches below the euphotic zone (DEUSER, 1971) and about 1% of the total production results in denitrified nitrogen (DEUSER et al., 1978), the mean 'residence time' of water masses between 75 and 1200 m in the northern Arabian Sea can be expected to lie in the range 43-54 years (SEN GUPTA et al., 1980b). A 'residence time' of 30 years has been suggested by HARTMANN et al. (1971) based on the rate of outflow from the Gulf of about 3 x 10^3 km^3 yr^{-1}.

Based on the oxidative ratios of the nutrients, the ratios of change of carbon: silicon: nitrogen: phosphorus in the waters of the northern Indian Ocean have been calculated as 108: 40: 16: 1 atoms. These agree fairly well with the atomic ratio of concentration of nitrogen: phosphorus in plankton of the Indian Ocean which is 15.4: 1.

Silicate concentrations in the deeper waters of the Indian Ocean are higher than those at the corresponding depths in the other two oceans. Silicate values can be used to identify different water masses by plotting them against the sigma-t values. This also has been confirmed by examining the relationship between silicate and oxygen consumption. The relationship shows that beyond a certain concentration of silicate, the relation becomes asymptotic, indicating that the water masses having such silicate and oxygen concentrations originate from a similar source (SEN GUPTA et al., 1976b).

BIOLOGICAL

Several atlases on the phytoplankton and zooplankton production in the Indian Ocean were published based on IIOE data (KREY and BABENERD, 1976; IOBC, 1968a and b). More recently both primary and secondary production rates of the Indian Ocean in relation to various relevant factors were estimated by QASIM (1977). The gist of the information is presented below.

The compensation depth of the euphotic zone (1% of the surface illumination) in the Indian Ocean varies between 40 and 120 m, with increasing depth from coastal to offshore regions.

The range in column production varies from 0.001 to 6.5 g C m^{-2} day^{-1}. Areas of high productivity lie mostly in the coastal and upwelling zones. These include the Gulf of Oman, the mouth of the Indus River, Saudi Arabian waters, Somalian and Mozambique coasts, southwest coast of India, the mouth of the River Ganga, the Burma coast, and off northern Sumatra Java (Fig. 4). The average production in the coastal areas works out to be 485 g C m^{-2} yr^{-1} as compared to the offshore average of 229 g C m^{-2} yr^{-1}. Total photosynthetic productivity in the Indian Ocean has been found to range from 3 to 6 x 10^9 tonnes C yr^{-1} with an average of 4.5 x 10^9 tonnes C yr^{-1} or 258 mg C m^{-2} day^{-1}.

The Arabian Sea and the Bay of Bengal are the two most productive regions of the Indian Ocean. The total column production in the Arabian Sea is 1.1 x 10^9 tonnes C yr^{-1}, while in the Bay of Bengal it is 0.4 x 10^9 tonnes C yr^{-1}. This means that these two areas account for about one-third of the total photosynthetic productivity of the Indian Ocean although together they occupy only about 14% of the total area. There is, however, a qualitative difference in the productivity of these two areas. The surface productivity (1 m depth) in the Bay of Bengal is 4.9 tonnes C km^{-2} yr^{-1} while in the Arabian Sea it is 3.9 tonnes C km^{-2} yr^{-1}. This anomalous situation between column and surface productivity has been attributed to greater cloud cover and higher surface nutrient load, particularly nitrate-nitrogen, in the Bay of Bengal (QASIM, 1977).

Zooplankton biomass in the Indian Ocean varies from 15 to 50 ml m^{-2}. The total zooplankton biomass in the Indian Ocean has been estimated to be 5.2 x 10^8 tonnes yr^{-1} (PRASAD et al., 1970), and secondary production works out to 69.3 x 10^6 tonnes C yr^{-1}. Areas of high abundance mostly coincide with high photosynthetic productivity.

Benthic production in the Indian Ocean predictably decreases with water depth. The density of meiofauna, however, is greater than that of the macrofauna in deeper waters. Benthic population is larger in the areas of high productivity. Average benthic productivity in the Arabian Sea to a depth of 2100 m is 1.8 g C m^{-2} yr^{-1}, while in the Bay of Bengal (to a depth of 1870 m) and the Andaman Sea (to a depth of 1820 m) productivities are 0.5 and 1.2 g C m^{-2} yr^{-1}, respectively (HARAKANTRA et al., 1980; ANSARI et al., 1977; PARULEKAR and ANSARI, 1981). Extrapolating these values to the whole of the Indian Ocean, the annual benthic production works out to be 89.9 x 10^6 tonnes in terms of carbon.

Recent figures of pelagic, demersal and crustacean resources from the different regions indicate that the annual catch for the Indian Ocean is about 4 million tonnes (FAO, 1978). Thus, the Indian Ocean, occupying 20.8% of the world oceans area, yields only 2.8% of the global fish catch. The potential yield of the Indian Ocean has been estimated as 14.2 x 10^6 tonnes yr^{-1} (QASIM, 1977). Therefore, there is much scope for increasing fish production of the Indian Ocean by increasing the effort.

GEOLOGICAL

Because of its asymmetric shape, the Indian Ocean is separated from the deep-reaching vertical convection areas of the northern hemisphere (DIETRICH and ULRICH, 1968). Recent geophysical investigations indicate that the Indian Ocean is the most complex of the three major oceans in terms of its history of formation, and hence its physiography, probably because of this asymmetric configuration (McKENZIE and SCLATER, 1971). Measurements of its magnetic field show that the floor of this ocean is spreading at the rate of 2 cm/year and implies that the Indian Ocean is about 10^8 years old (DIETRICH, 1973).

One of the most prominent topographic features of the Indian Ocean is its seismically active, rugged inverted Y-shaped Mid-Indian ridge, which is cut by numerous north-northeast trending fracture zones (Fig. 5) shown by HEEZEN and THARP (1964). Microcontinents and meridional aseismic ridges, e.g., the Ninety-East Ridge and the Chagos-Laccadive Plateau, are unique to the Indian Ocean (HEEZEN and THARP, 1964). Although massive, these features are relatively smooth-surfaced blocks and stand in marked contrast to the rugged features of the Mid-Indian ridge.

Like in the other oceans, there are a good number of deep basins between the topographic highs. Sediment cones and abyssal plains, with sediment thicknesses exceeding 1.5 to 2.5 km are present in the Bay of Bengal, Arabian Sea, Somali basin and Mozambique basin (EWING et al., 1969). In parts of the Ganges cone, sediment thickness exceeds 12 km (CURRY and MOORE, 1971). In the high southern latitudes near the polar front, smooth, 'swale' topography with moderate sediment thickness appears as a result of high productivity of the Antarctic Ocean, while in other areas, and especially on topographic highs, the sediment cover is very thin (VENKATARATHNAM and HAYES, 1974).

The complex physiography, the asymmetrical distribution of the surrounding continents, their varying geology and climate, and the northward closure of the Indian Ocean (preventing the influence of colder northern latitudes) have a considerable effect on the fauna and the sediment character in the northern part of the Indian Ocean. Moreover, the Indian Ocean receives about 34 x 10^8 tonnes of suspended sediment per year from the rivers draining into it (HOLEMAN, 1968). Almost half of this, about 16 x 10^8 tonnes, is added by the river systems draining the Indian sub-continent.

In terms of areal extent, the most extensive sedimentary facies in the Indian Ocean is calcareous sediment which dominates the Mid-Indian Ridge and other more shallower areas not influenced by terrigenous influx or the influx of siliceous skeletal remains (VENKATARATHNAM and HAYES, 1974). The critical depth at which calcium carbonate becomes a minor component varies with latitude (LISITZIN, 1971). Other sedimentary facies include: terrigenous clay adjacent to the major river basins; siliceous clay and ooze in the equatorial and southern latitudes influenced by high biological productivity; brown and red clays in deep areas outside the productivity belts and the areas of terrigenous influx (VENKATARATHNAM and HAYES, 1974; SIDDIQUIE et al., 1976a and b). Although the areal extent of terrigenous sediment is relatively small, because of its enormous thickness its volume exceeds 70% of the total sediments of the Indian Ocean (EWING et al., 1969). Within the limits of the broad features noted above, VENKATARATHNAM and HAYES (1974) mapped the following materials in the Indian Ocean: (i) hard rocks (volcanic and limestones) on topographically high areas, especially on the Mid-Indian Ridge; (ii) Mn-nodules in the deep basins and Agulhas Plateau, far removed from terrigenous sediment supply; and (iii) silicic volcanic ash in the eastern Indian Ocean adjacent to the Indonesian archipelago. In addition, deposits of ilmenite placers have been mapped in the coastal regions of several of the countries bordering the Indian Ocean, (e.g., India, Australia and Sri Lanka).

NATURE OF ENVIRONMENTAL PROBLEMS

Almost all the countries bordering the Indian Ocean are developing countries. Their major source of revenue is agriculture and, in some countries, also mining. The effects of pollution in the marine environment, largely arising out of the economic activities, began to be felt only recently although the practices have continued for a long period of time. These problems, however, are largely confined to the coastal areas of most countries, although because of the prevailing wind system, the patterns of water circulation and the nature of bottom topography, the effects can have far-reaching implications.

While some of the problems are chronic in some countries, they are relatively simple in others. In this review, however, no attempt has been made to differentiate them.

DOMESTIC SEWAGE AND INDUSTRIAL EFFLUENTS

Due to increasing urbanization and industrialization throughout the region, the volume of sewage and industrial waste is increasing constantly, and, as a result, both contribute directly and indirectly to the degradation of the adjoining seas. Sewage and industrial wastes in these countries, untreated or partially treated, are discharged into the rivers and seas. Many countries have large rivers, but because of increased human activities, many of these rivers have become badly polluted. In India, for example, only 42 cities with populations over 100,000 have arrangements for sewage treatment. Hardly 50% of the total population in the countries bordering the Indian Ocean is provided with proper sanitation arrangements. The result is that high counts of coliform bacteria often are found on the beaches and in coastal waters.

It has been calculated that 27 million people in East Africa live along the coastline and are directly or indirectly dependent on the sea for their livelihood. Calculating the consumption of water at 60 litres/capita/day, this population can be expected to generate $585 \times 10^6 m^3$ sewage and effluents per year. The same population will generate 7.8×10^6 tonnes of solid waste (at 0.8 kg capita/day).

Only 17% of a population of 3.8 million living in the coastal cities in eastern Africa are sewered (UNEP, 1982a, b). Representative values of this sewage before disposal indicate high BOD and COD loads, along with a high concentration of suspended solids. The solid wastes generated are used mostly for landfill. Consequently, the risk of contamination of ground water from the landfill leachate is fairly high. Representative analysis of leachates from Kenya indicate high values for BOD, alkalinity and total hardness. Assuming that the industrial effluents would be about 10% of the sewage, the total quantity which is discharged in the coastal marine environment, in mostly pre-treated form, would come to $59 \times 10^6 m^3 yr^{-1}$. The sources of these effluents are slaughter houses, breweries, distilleries, textile mills, saw mills, refineries, fertilizer plants, pulp and paper mills, etc. Representative analysis of the industrial effluents from Tanzania indicate high BOD and COD loads.

Concentrations of toxic metals, such as Hg, Cd and Pb, in plankton and fish are still much lower than the acceptable maxima in many industrialized countries.

AGRICULTURAL WASTES

Fertilisers, pesticides and insecticides are abundantly used in developing countries for agriculture, pest control and vector control. The quantities of pesticides and insecticides used every year vary from 45,000 tonnes in India to 3.5 tonnes in Bangladesh (NIO, 1976). In many countries, however, organochlorine pesticides are either prohibited or gradually are being replaced by organophosphorus and carbamate pesticides. Very little study on their accumulation and harmful effects has been carried out. However, a recent survey has shown that plankton in the Arabian Sea, off the west coast of India, has DDT concentrations ranging from 0.05—3.21 ppm wet weight (KUREISHY et al., 1978).

Use of these harmful organochlorine and organomercurial pesticides has been totally banned in all the industrial countries of the world. But their total production has not been reduced; rather their manufacture has increased, suggesting that soon there will be a 'southward' tilt of the concentration of organochlorine and organomercurial pesticides in the marine environment.

RADIOACTIVE AND THERMAL WASTES

In countries in the Indian Ocean region, power generation is mostly thermal. But in some countries, nuclear power is also being generated. So far no serious harm has been reported from these sources, although fly ash from thermal power plants invariably creates environmental problems.

Radioactive wastes from nuclear power plants are normally discarded according to strict international conventions. However, their heat generation can pose several problems. Nuclear power plants normally release 50% of the generated heat to the coastal marine environment. Flora and fauna in the warm tropical waters live dangerously close to their upper lethal limits of temperature, particularly during the warm summer months, and as such they cannot withstand a large increase in ambient temperature. Release of hot salty water, when coupled with the wind system, can alter the current and mixing patterns in local areas. This is more probable in tropical areas where the range of the semi-diurnal tides is quite high.

OIL SPILLS

In 1982 the global marine transport of oil was 1270 million tonnes (mt), of which 579 mt or 43% of the total was shipped from the Gulf countries (BP, 1982). The main routes of marine transport of oil from the Gulf countries are across the Arabian Sea. One of these is through the Mozambique Channel round South Africa to the Western Hemisphere, while the other one is around Sri Lanka across the southern Bay of Bengal and through the Malacca Strait to Far East and Japan (Fig. 6). In 1982, 331 mt of oil was shipped to the Western Hemisphere and 248 mt to the Far East and Japan from the Gulf countries. This, coupled with the increasing emphasis on offshore oil exploration in many countries of the region, makes the northern Indian Ocean very vulnerable to oil pollution.

Sources of oil pollution include tanker disasters, ballast water and bilge washings. Fortunately, only a few tanker disasters have occurred so far along these tanker routes. However, the effect of the oil pollution can be seen on the beaches in the form of deposits of tar-like residue. The frequency and intensity of this residue depend on the current direction along the coastal regions. Because of the monsoon winds, the surface currents change the direction every six months. Whenever a shoreward component of current develops, heavy deposition of tar balls occurs on the beaches.

It has been calculated that at any time the amount of floating tar in the surface layers of the Arabian Sea is about 3700 tonnes, while along the tanker route across the southern Bay of Bengal the tar amounts to 1100 tonnes (SEN GUPTA and KUREISHY, 1981). This agrees well with the intensity of tanker traffic and the volume of oil transported through the two areas.

The East African countries together import 6.55 mt of oil per year, and oil pollution is a chronic problem in countries of this region (NIO, 1980) (Fig. 7). It has been estimated that the total oil spill from tanker traffic, harbour operations, coastal industries, etc., comes to about 33,000 tonnes per year in East Africa (UNEP, 1982c). Figure 6 suggests that tanker traffic, from the Gulf to South Africa, represents the single largest source of oil pollution.

TOURISM

In developing countries more and more emphasis is being given towards the promotion of tourism. The result is that too many large hotels are being constructed along the beaches. Wastes from these hotels very often pose serious problems in the adjacent marine area. In some countries garbage and other wastes from the hotels have spoiled the beauty of the beaches, leading to contamination of the environment and generation of hydrogen sulphide in the water.

CORAL REEFS AND MANGROVES

Coral reefs and mangroves occur widely in most all Indian Ocean coastal areas. Damages that can occur to these two ecosystems as a result of over-exploitation of the reefs or mangroves or because of pollution are quite extensive. These can be summarized as follows:

Coral Reefs

Coral reefs of the tropical Indian Ocean include fringing and barrier reefs; sea-level atolls; and elevated reefs (STODDART, 1972). They occur along the NW and SE coasts of India, the Laccadive Islands, Andaman and Nicobar Islands, Maldive Islands, Chagos Archipelago; Seychelles, Mauritius, Aldabra and Comores Islands, coasts of Kenya, Tanzania and West Malagasy. Radiocarbon dating of some of the coral reefs has indicated ages of more than 2000 years (STODDART, 1972). The growth rate of corals in the Atlantic has been found to be 0.15—0.5 cm yr^{-1} by ^{228}Ra (MOORE and KRISHNASWAMI, 1972). The growth rate of the corals in the Indian Ocean can be assumed to be of a similar order.

Most of the coral reefs in this region have been declared as endangered ecosystems. Several coral reefs have almost disappeared because of the withdrawal of coral debris and live corals as a raw material for cement industry, while others have become dead due to their constant exposure to pollutants, particularly oil. Examples of this are the Kavaratti reef in the Laccadives; reefs in the southern part of the Great Nicobar Island in the Andaman group; and SW part of Malagasy. Dredging for the construction of jetties and works connected with the development of harbours has destroyed some of the coral reefs on Mahe Island in the Seychelles.

Mangroves

Mangroves, which occur along NE coast of India, Bangladesh, Malaysia, Sri Lanka, Mauritius, Seychelles and Indonesia, constitute an important resource in the region as they serve as spawning grounds, nurseries and feeding grounds for economically important fishes and crustaceans. They act as a buffer zone and offer protection to vulnerable communities like the coral reefs. They also stabilise the bottom sediments, control the local mean water levels and the direction of flow. Mangroves constitute a significant portion of the coastal wetland in many countries and a fairly large percentage of human population is dependent on them. For example, about one-eighth of the Bangladesh wetland is mangrove and one-third of the total population of that country is either directly or indirectly dependent on the mangrove ecosystem (UNESCO, 1979).

Due to the ever increasing demand for land and fuel, many mangrove areas of the Indian Ocean region have been destroyed. This has led to heavy siltation in the nearshore region and, with no protective cover of mangrove, the devastation of men and material in coastal areas during cyclones.

SILTATION

As noted above, the Indian Ocean receives annually 34 x 10^8 tonnes of suspended sediment (HOLEMAN, 1968), half of which comes from the rivers draining the Indian sub-continent. This quantity is increasing due to human activities, such as mining, clearance of land for agriculture, lumbering, urbanization and industrialization, and dredging to deepen harbour channels and estuaries.

Most of the silt settles near the river mouths and in coastal areas. This probably decreases the productivity of water and depletes fishery resources. Although these effects have been suggested in many countries, not much direct evidence has been obtained on the influence of siltation on marine and estuarine fisheries.

RESEARCH AND MONITORING ACTIVITIES

Monitoring and research on marine pollution are fast developing activities in many countries of the Indian Ocean region. Baseline surveys on almost all potential pollutants in the marine environment are underway, and a fairly good data base has been built up. These data mostly concern environmental conditions in the coastal and nearshore regions and also on some parts of the open areas of the Arabian Sea, the Bay of Bengal, and adjacent areas of the Laccadive and the Andaman Islands. Several reviews have been published in recent years (QASIM and SEN GUPTA, 1980, 1983a and b). The following details have been extracted from these publications combined with data collected subsequently.

PETROLEUM HYDROCARBONS

In all, 6689 observations were made on oil slicks and other floating pollutants along the tanker and trade routes across the northern Indian Ocean. Of these, oil was sighted on 5582 occasions or 83.5% of the total number of observations (SEN GUPTA and KUREISHY, 1981). Plotted in 5°-squares (Fig. 6), the percentage of oil sightings ranged from 51 to 96. Even in areas away from tanker and trade routes, the percentage of oil sightings was almost of a similar order. Fortunately, the occurrence of tanker disasters has not been very significant along these routes so far, and most of the oil released in this region comes mainly from the ballast and bilge washings.

Observations on the floating petroleum residues from the Indian Ocean region are presented in Table 1. As expected the concentrations and variations in time and are fairly high occassionally along the tanker routes. In the Arabian Sea concentrations range from 0 to 6.0 mg m^{-2}, with a mean of 0.59 mg m^{-2}. The range along the Bay of Bengal tanker route varies from 0 to 69.75 mg m^{-2}, with a mean of 1.52 mg m^{-2}. This suggests that the tanker route in the Bay of Bengal is relatively more polluted than that in the Arabian Sea.

Observations in March 1981 along the eastern oil tanker route across the Arabian Sea, between 12° and 19°N and 70° and 74°30'E, indicated a range of 0.3—112.2 mg m^{-2} with an average of 12.7 mg m^{-2} for the floating tar balls. This may, perhaps, indicate that oil pollution in the Arabian Sea is on the increase.

A number of observations were taken along the eastern tanker route in the Arabian Sea during June-September 1983 (Table 1). Absence of floating tar balls in this region during this period was expected, as the surface currents on the Arabian Sea normally flows directed towards the Indian west coast during the SW monsoon months.

Applying the average concentrations of floating tar balls to the areas of the Arabian Sea (6.225×10^6 km^2 from 0°—25°N and 50°—80°E) and the tanker route across the southern Bay of Bengal (0.73×10^6 km^2 from 5°—7°N and 80°—95°E) the total quantity present comes to the 3700 and 1100 tonnes quoted previously.

Accumulation of floating tar is entirely dependent on surface currents. In areas of strong surface currents, e.g. the Gulf Stream and Kuroshio, heavy accumulation of tar particles has been observed. The observed concentration in the North Atlantic ranges from 0.12—0.64 mg m^{-2}, with a maximum of 91.8 mg m^{-2} (LEVY and WALTON, 1976). Accumulations in the northwestern Pacific range from 0.02—13.3 mg m^{-2} (SANO et al., 1979). Thus, the oil tanker routes in northern Indian Ocean appear to be as much polluted as those in the other oceans.

If we assume that the abundance of oil entering any marine area is proportional to the volume transported, the average amount of oil present in the seas around India is 0.36% of the total flow, 1% of the total input ends up as floating tar, and the volume transported across the Bay of Bengal tanker route is 35 % of the total transport across the Arabian Sea, then the 'residence time' of floating tar in the northern Indian Ocean ranges from 33 to 38 days (SEN GUPTA and QASIM, 1982). The average tar concentration on the surface of the North Atlantic has been calculated as 5200 tonnes and its 'residence time' is about 58 days (LEVY and WALTON, 1976). Thus, it can be concluded that tar found in the sea remains unchanged from 30 to 60 days.

As can be seen from Table 1, the concentrations of dissolved and dispersed petroleum hydrocarbons from the surface to 20 m are almost uniform. However, some seasonal variations may be due to differences in the intensity of tanker traffic from month to month and because of changes in meteorological conditions.

The average concentrations of dissolved petroleum hydrocarbons in the upper 20 m of the tanker routes in the Arabian Sea and the Bay of Bengal are 42.8 and 28.2$\mu g\ kg^{-1}$ respectively. Calculating the volumes in the upper 20 m from the total areas mentioned earlier, the quantity of petroleum hydrocarbons in the uppermost 20 m of the water column would be about 5 x 10^6 tonnes in the entire Arabian Sea and 0.4 x 10^6 tonnes along the Bay of Bengal tanker route. Summing up of all the values in Table 1 results in Figure 6, depicting the dissolved petroleum residues in the northern Indian Ocean.

All available data on dissolved/dispersed petroleum hydrocarbons in the upper 20 m of the Northern Indian ocean (Table 1A) indicate a general decreasing trend of oil transport from the Middle East countries, particularly along the west going tanker route. Consequently, the volume of tanker traffic has also reduced considerably. Due to the reduction in traffic, an apparent reduction in the oil pollution can be deduced from the data for the tanker routes across the Arabian Sea and the Bay of Bengal. The reduction was sharp from 1982 onwards in the Arabian Sea and from 1981 onwards in the Bay of Bengal. Thereafter, the conditions appear to have stabilized in both the areas. However, the net reduction in transport is more significant for the west-going tanker route than that for the east-going one. In fact, more oil was transported to the eastern hemisphere from 1984 as compared to the western hemisphere.

Almost all of the countries of this region are signatories to the 1973 MARPOL Convention of IMO and the 1978 protocol. Many have already ratified the accords. These have resulted in the establishment of reception facilities for oily bilges, ballasts and sludges at almost all the major ports. With the establishment of such facilities at all the ports, oil pollution in the Northern Indian Ocean can be expected to be reduced in the future.

A few observations on the concentration of petroleum hydrocarbons in zooplankton and in the sediments of the Arabian Sea off the west coast of India are also available (FONDEKAR et al, 1980). In zooplankton concentrations were found to range from 19.5—83.3 $\mu g\ g^{-1}$, while in the sediments the range was 4.8—8.5 $\mu g\ g^{-1}$, both on dry weight basis.

Deposition of tar-like residues on the beaches is more or less a chronic problem in almost all countries bordering the Indian Ocean. However, this is a seasonal phenomenon, regulated by the monsoons. Records from the west coast of India during the years 1975 and 1976 indicate a range of 22 to 448 $g\ m^{-2}$ (DHARGALKAR et al., 1977). Observations on the beaches in Kenya give a range of values of 20 to 2400 g 100 m^{-2}, with a peak value of 1386 $g\ m^{-2}$ (EAMFRO, 1973). Tar lumps as heavy as 16 kg have been observed on Kenyan beaches.

HEAVY METALS OTHER THAN MERCURY

In Water

Some data on the dissolved concentrations of heavy metals are available from the Indian Ocean region (Table 2). The large ranges may be due to different analytical techniques used as well as horizontal and vertical variations. The ranges are: Cu 0.08—49.1, Cd 0.01—1.88, Fe 0.10—96, Mn 0.07—80, Zn 0.3—42.4, Pb 0.02—7.5, Ni 0—16.3, and Co 0—7.9, all in $\mu g\ l^{-1}$. Except for copper, all the higher values are either from coastal India or off the coasts of oceanic islands. The higher ranges of iron, manganese and nickel are from around the Laccadive Islands; high cadmium and lead are from around Andaman Islands; high zinc occurs off Bombay, and high cobalt off the river mouths in the Bay of Bengal. An exceptionally high value of cadmium, 80 $\mu g\ l^{-1}$, was recorded in the polluted coastal waters off the city of Bombay (GANESAN et al., 1980).

In Plankton and Fishes

Available values (KUREISHY et al., 1981) of some of the metals in zooplankton from the Arabian Sea and Bay of Bengal give the following ranges in ppm fresh weight: Cd 0.7—6.0, Cu

2—5, Mn 3—7, Zn 8—31, Fe 35—94, Pb 1—13, Ni 0.2—3, and Co 0—4 (Table 3). In all these analysis, the recovery was calculated with reference to the standard fishmeal obtained from International Laboratory of Marine Radioactivity, Monaco. Concentrations of all these metals have been analysed in 26 species of fish and crustaceans from both inshore and offshore regions of the northern Indian Ocean and in different parts of their tissues, such as, muscle, liver, gills and heart (KUREISHY et al., 1979, 1981). It can be seen from Table 3 that concentrations of almost all of the metals, particularly the toxic metals such as Cd, Pb and Hg, are within the permissible limits for human consumption. It has been observed, in the same study, that the concentrations of all these metals in the livers of the dolphin fish (*Corryphaena hippurus* Linnaeus), barracuda (*Sphyraena picuda* Bloch), sharks (*Eulamia ellioti* Day), skipjack tuna (*Katsuwonus pelamis* Linnaeus), and yellowfin tuna (*Neothunus macropterus* Schlegel) were significantly higher than in their muscles. This indicates that most of the metals are assimilated by these fishes in a fat soluble form. Acceptable correlation was, however, observed in the different fish tissues with their sexes, sizes and stages of maturity.

MERCURY

Because of its extreme toxicity and very harmful nature to human beings, mercury warrants special discussion. Unfortunately, the study of mercury in the marine environment is relatively of a recent origin in India and hence data on this metal are very sparse.

In Water

Mercury concentrations in dissolved form are normally found at the nannogram levels. Their sources are mostly industrial effluents, but some contribution also can be expected from weathering, leaching and atmospheric flux from deposits of mercury in the crustal rocks of the hinterland. High concentrations of mercury in the nearshore waters of the SW coast of India are probably due to deposits of mercury in the adjoining land. In the absence of any industrial source, such high concentrations, ranging from 204 to 407 $\mu g\ l^{-1}$ in the surface waters, cannot be due to any other reason (SINGBAL et al., 1978).

Ranges of all the observed values are presented in Table 2. Without taking into account the variation with depth, the range is from 0 to 222 $\mu g\ l^{-1}$. However, a decrease occurs in the concentration of mercury with depth, as is the case with most of the other metals (SINGBAL et al., 1978; SANZGIRI and BRAGANCA, 1981).

In Plankton and Fishes

An estimation of the total mercury concentration in plankton and fishes was carried out without fractionating the mercury into methyl and inorganic forms. In organic matter, however, it can be assumed that mercury is present mostly in the methyl form and very little in the inorganic form.

The absence of mercury in zooplankton (Table 3) indicates that the metal is probably assimilated by fishes through other pathways. Mercury concentrations in the muscles of sharks and skipjack tuna are particularly high (Table 3). Concentrations of non-essential heavy metals (Hg, Cd and Pb) in different tissues of the fishes from the northern Indian Ocean (Table 4) indicate highest occurrences in the liver and kidney. However, the highest concentrations of mercury, found in muscles, are still much lower than the internationally permissible maximum of 0.5 ppm of mercury for human consumption. Mercury contents in the muscles of several commercially important fishes from inshore regions and from a polluted creek in and around the city of Bombay range from 0.04—0.57 ppm on a fresh weight basis (TEJAM and HALDAR, 1975; RAMA and SOMAYAJULU, 1972). However, the concentration of the same metal in crab muscles and sediments analysed in 1980, showed an increase by several fold (GANESAN et al., 1980). These values (Table 5) also indicate a decrease in the mercury concentration away from the shore. A few fishes around Seychelles have been observed to contain 1-2 mg kg^{-1} mercury, much above the maximum permissible limit of 0.5—1 mg kg^{-1} in many countries.

PESTICIDES

Some analyses of pesticide residues in zooplankton in the Arabian Sea and in sediments along the east coast of India have been carried out (KUREISHY et al., 1978; KANNAN and SEN GUPTA, 1987; SARKAR and SEN GUPTA, unpublished). Concentrations in zooplankton only for DDT and its isomers, represented as t-DDT, (Figure 7) decrease from nearshore to offshore regions indicating the land origin of the pesticides. Measurements of t-DDT in the surface waters of the Arabian Sea and in the air over it indicate a northward increase from the equator (TANABE and TATSUKAWA, 1980), confirming the land origin of these residues.

A recent study of pesticides residues in the sediments along the east coast of India (SARKAR and SEN GUPTA, unpublished) indicates that, apart from DDT and its isomers, residues of Gamma BHC, Aldrine and Dieldrine also are present at a number of places. Their individual concentrations are higher than t-DDT at a few stations. Diagrammatic representation of these values indicates higher concentrations mainly at the river mouths (Fig. 8). Once again, this would indicate the land origin of the pesticides.

The estimated annual amount of pesticides and insecticides used in India is about 55,000 tonnes (Table 6). Recent estimates indicate that about 3,000 tonnes of pesticides are imported into Bangladesh every year (UNEP, 1986). Most of the countries in this region are mainly agricultural and considerable quantities of pesticides and insecticides also can be expected to be used by them. It is commonly believed that about 25% of these ultimately end up in the sea. The cumulative effect of the non-degradable pesticides over the years can be considerable in the coastal marine sediments.

INPUT STUDIES

Data on the input of different cations and anions to the Indian Ocean are very scanty. Background data on sources of input, e.g., river runoff, load of silt, volume of domestic sewage and industrial effluents, rainfall, and aerosol, are not available from all parts of the region. However, some data are available from India (Table 6). A rough approximation for all the countries perhaps can be obtained by extrapolating the data given in Table 6 to population density.

Most of cations and anions added to the oceans are adsorbed on to clay and silt particles. Considering the enormous amount of sediment added annually to the Indian Ocean, i.e., 34×10^8 tonnes (HOLEMAN, 1968), river-borne sediment may be assumed as one of the principal sources of input.

The mean concentrations of some elements in the rain water from the coastal region of India are: Na 4.6, Cl 6.8, Ca 1.34, Mg 0.51, K 1.02 all in mg/l; Mn 23.4, Br 33.6, I 13.3 all in $\mu g\ l^{-1}$ (SADASIVAN, 1979); Fe 4.3, Co 0.1, Ni 0.1, Cu 6.8, Zn 13.3, Pb 21.5 all in $\mu g\ l^{-1}$ (FONDEKAR and TOPGI, 1979). Aerosol contents over the Indian Ocean varies from 1.4 to 68 $\mu g\ m^{-3}$ air. It decreases southwards with the latitude and becomes zero close to Antarctica (SEN GUPTA and QASIM, 1983a). From these values it can be inferred that the atmospheric flux of the elements is not very significant and the major transport route of pollutants in the Indian Ocean is through the riverine source.

Another important source of input is from mining activities. Rejects and tailings from the mining operations carried out in coastal and nearshore regions add substantial quantities of heavy metals to the coastal environment.

ASSESSMENT OF POSSIBLE IMPACTS OF POLLUTANTS ON THE MARINE ENVIRONMENT

From the information summarized above it is clear that the data collected so far are too fragmentary to allow a clear understanding of the total pollution problem in the Indian Ocean. However, the information suggests that this region is not yet seriously affected by pollution.

Localized problems, both short-term and long-term, do appear from time to time, and their overall importance varies from country to country.

However, the problem common to almost all the countries is oil pollution. Because of the transportation of a large volume of oil through the Indian Ocean, many areas are being damaged. The worst affected ecosystems probably will be coral reefs and sandy beaches. Significant damage already has been noted on some of the atolls of Laccadives, coral reefs of the Andaman and Seychelles islands, Madagascar, and also from coastal regions of Kenya and Tanzania.

Other possible impacts can be expected to come from the disposal of untreated or little treated domestic sewage and industrial effluents. Continued washings of fertilizers, pesticides, herbicides and insecticides due to widespread agricultural practices in all the Indian Ocean countries also can pose significant problems.

But in the tropical and equatorial regions of the Indian Ocean, the tides are mostly semi-diurnal, with ranges varying from less than 1 m to more than 8 m. Tidal flushings twice a day, associated with biannual reversal of the direction of monsoon winds and the surface currents associated with smooth bottom topography, help in dispersing and diluting the pollutants and reducing the magnitude of their impact on the marine environment considerably.

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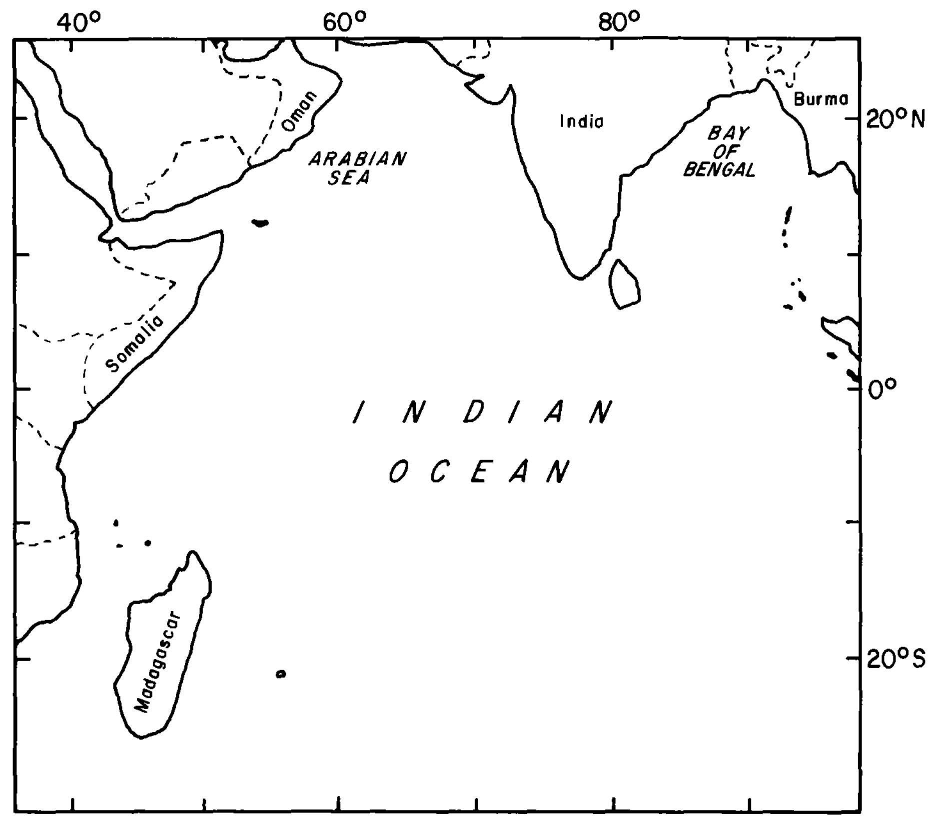

Figure 1. The Indian Ocean as defined in this paper.

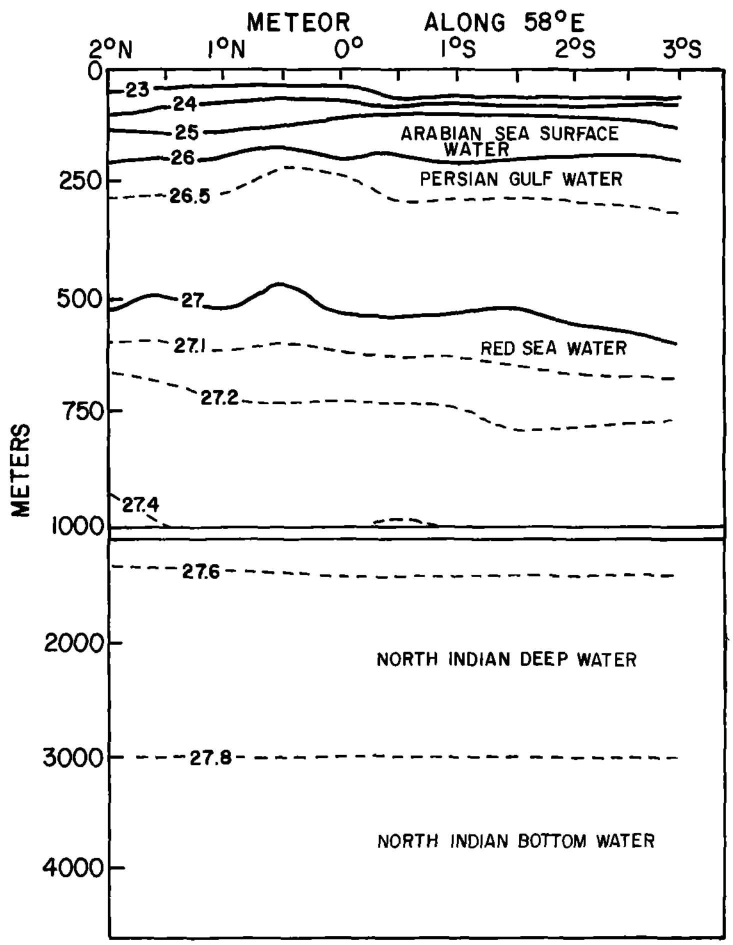

Figure 2. Water masses in a cross-section at 58°E longitude in the Indian Ocean (From METEOR data 1964/65).

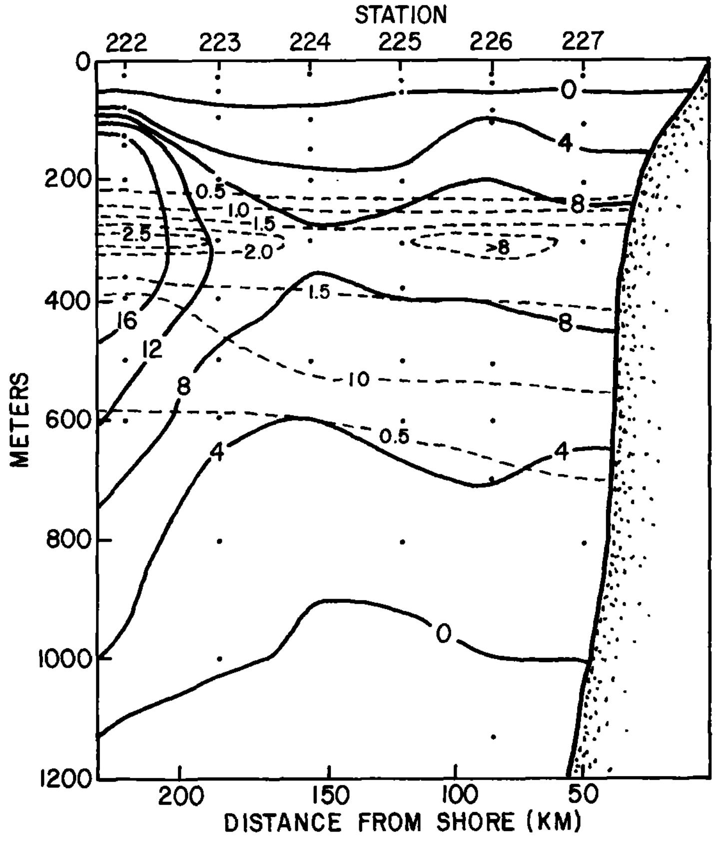

Figure 3. Calculated nitrate anomaly and nitrite distributions for METEOR (1964/65) stations 222-227 in the northeastern Indian Ocean. Nitrate anomaly (continuous lines) and dissolved nitrite (broken lines). Concentrations are in ug-at l^{-1}

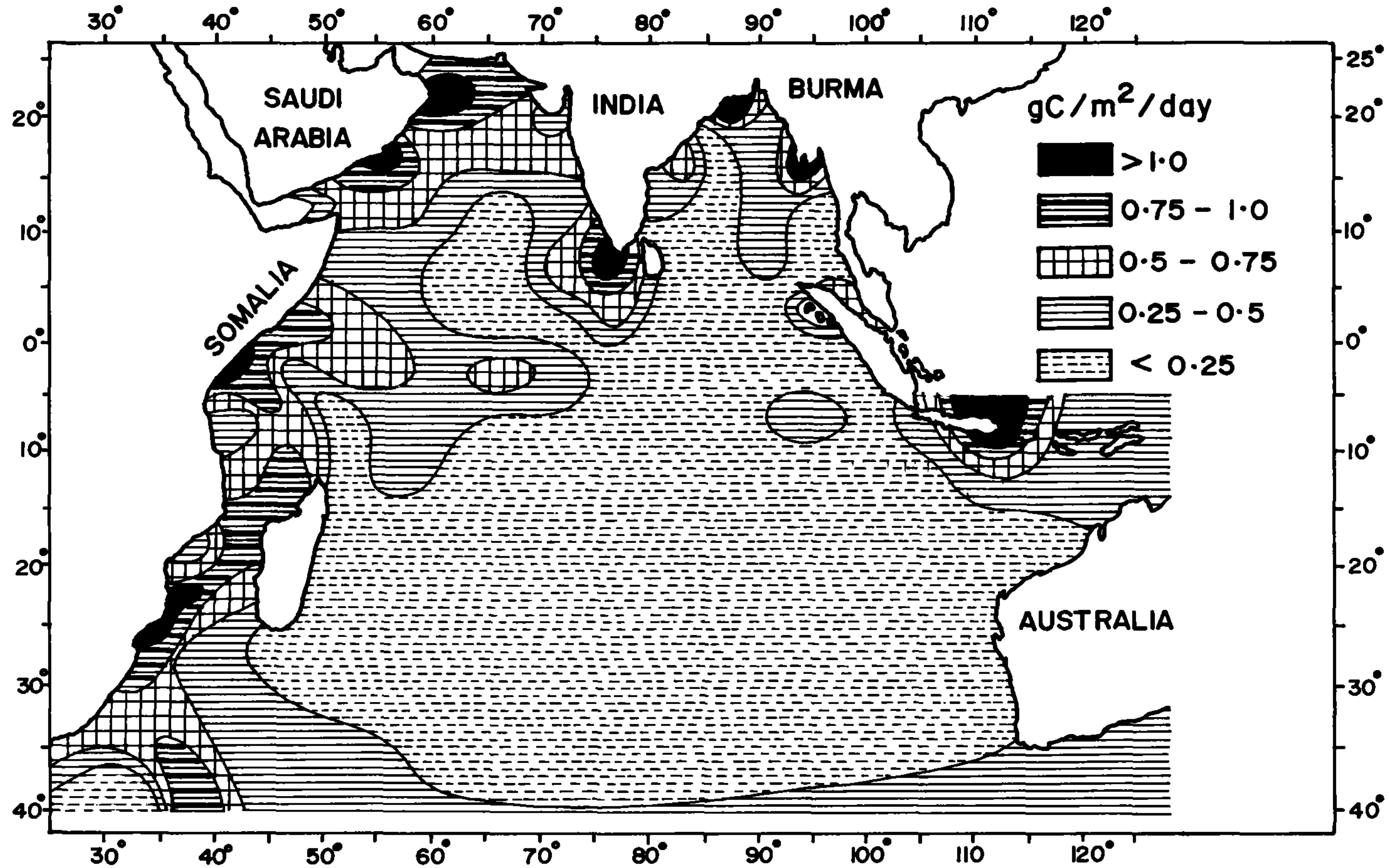

Figure 4. Primary productivity in the Indian Ocean (From QASIM, 1977).

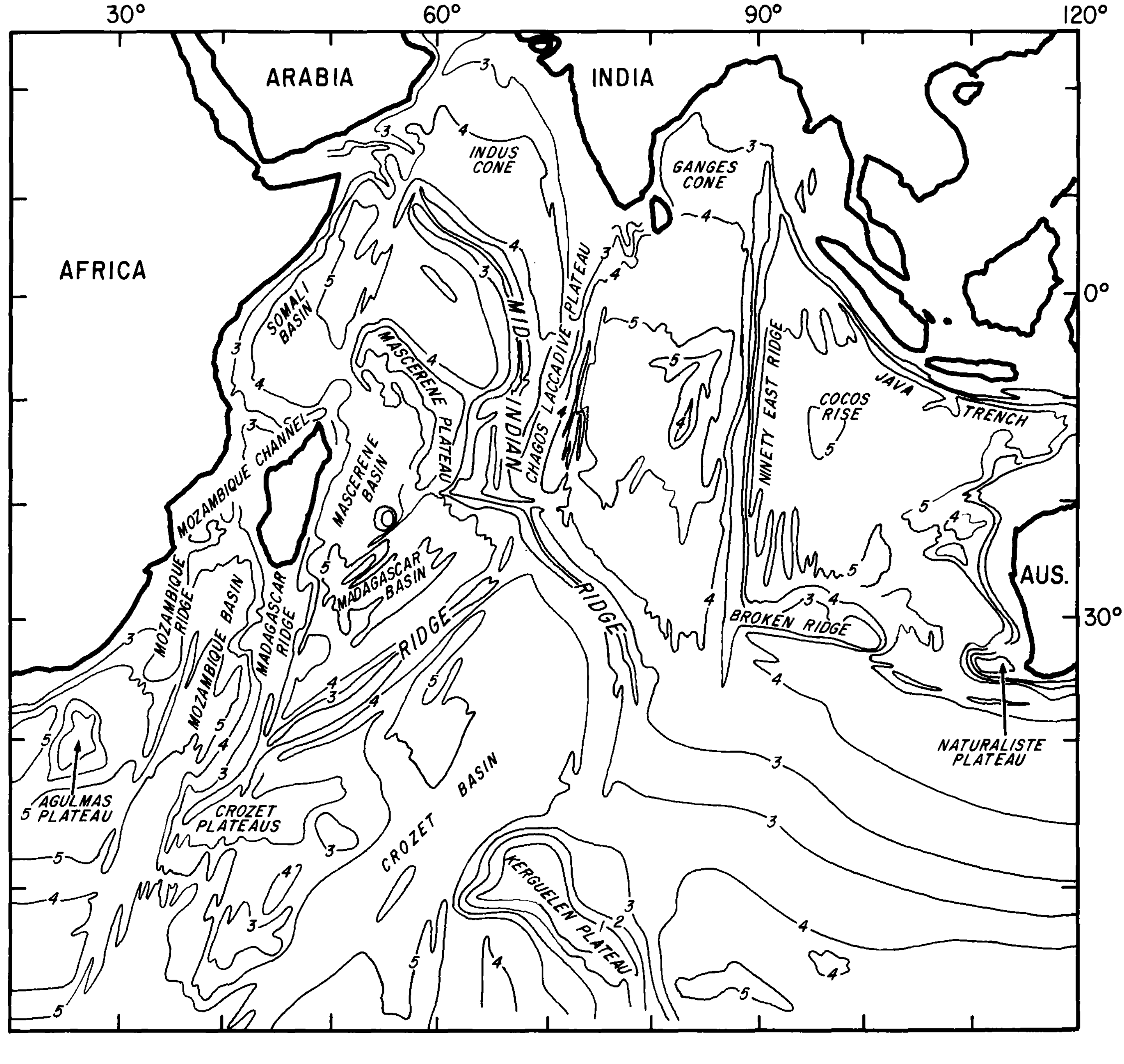

Figure 5. The Indian Ocean floor (from HEEZEN and THARP, 1964).

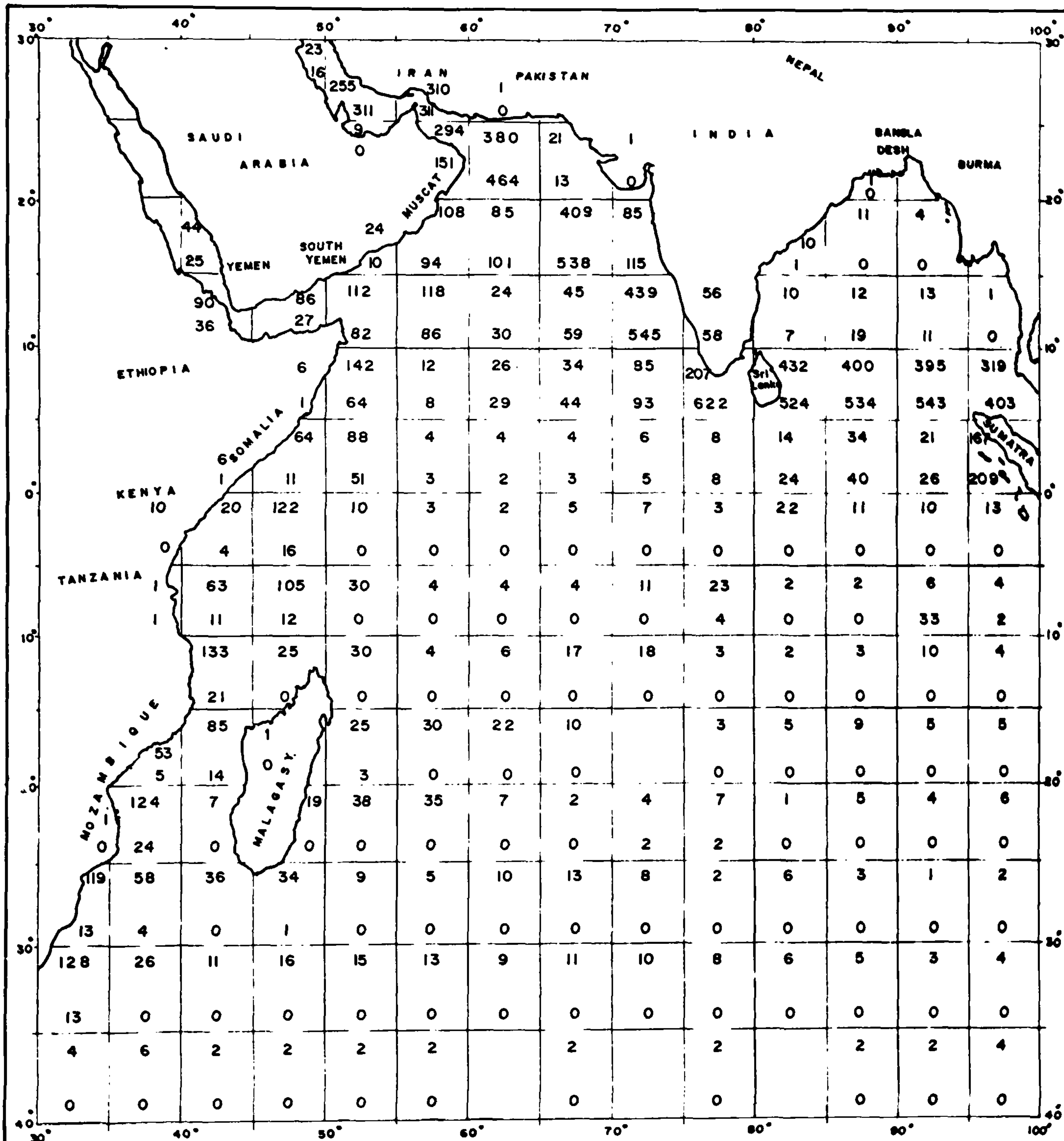

Figure 6. Observations of oil slicks and other floating pollutants, every 50-square in the Indian Ocean. Upper numbers indicate number of sightings; lower numbers indicate occasions of the absence of oil slicks.

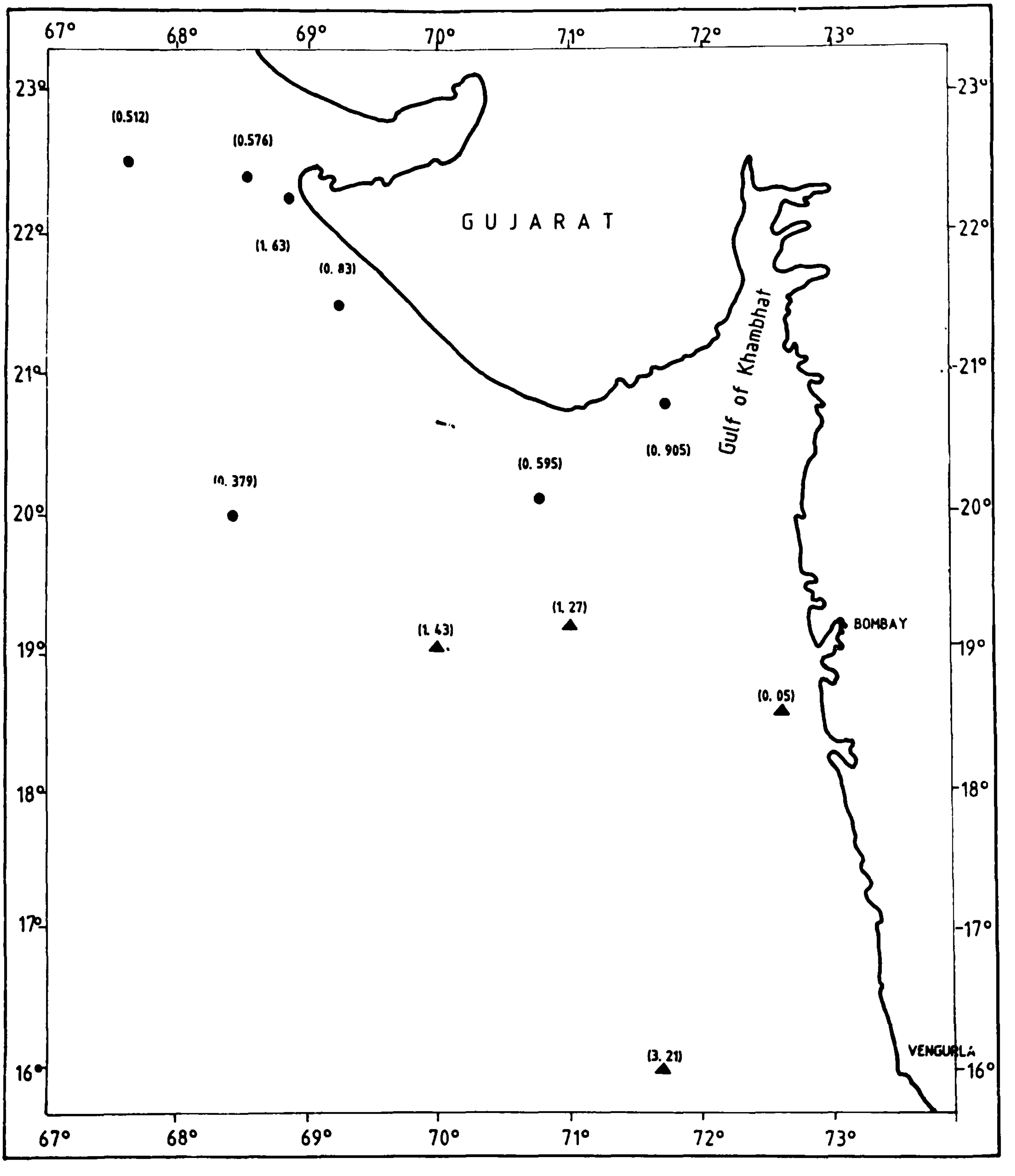

Figure 7. t-DDT in zooplankton of the eastern Arabian Sea. Closed triangles are observations in 1978. Closed circles indicate results of observations in 1985 (from KANNAN and SEN GUPTA, 1987).

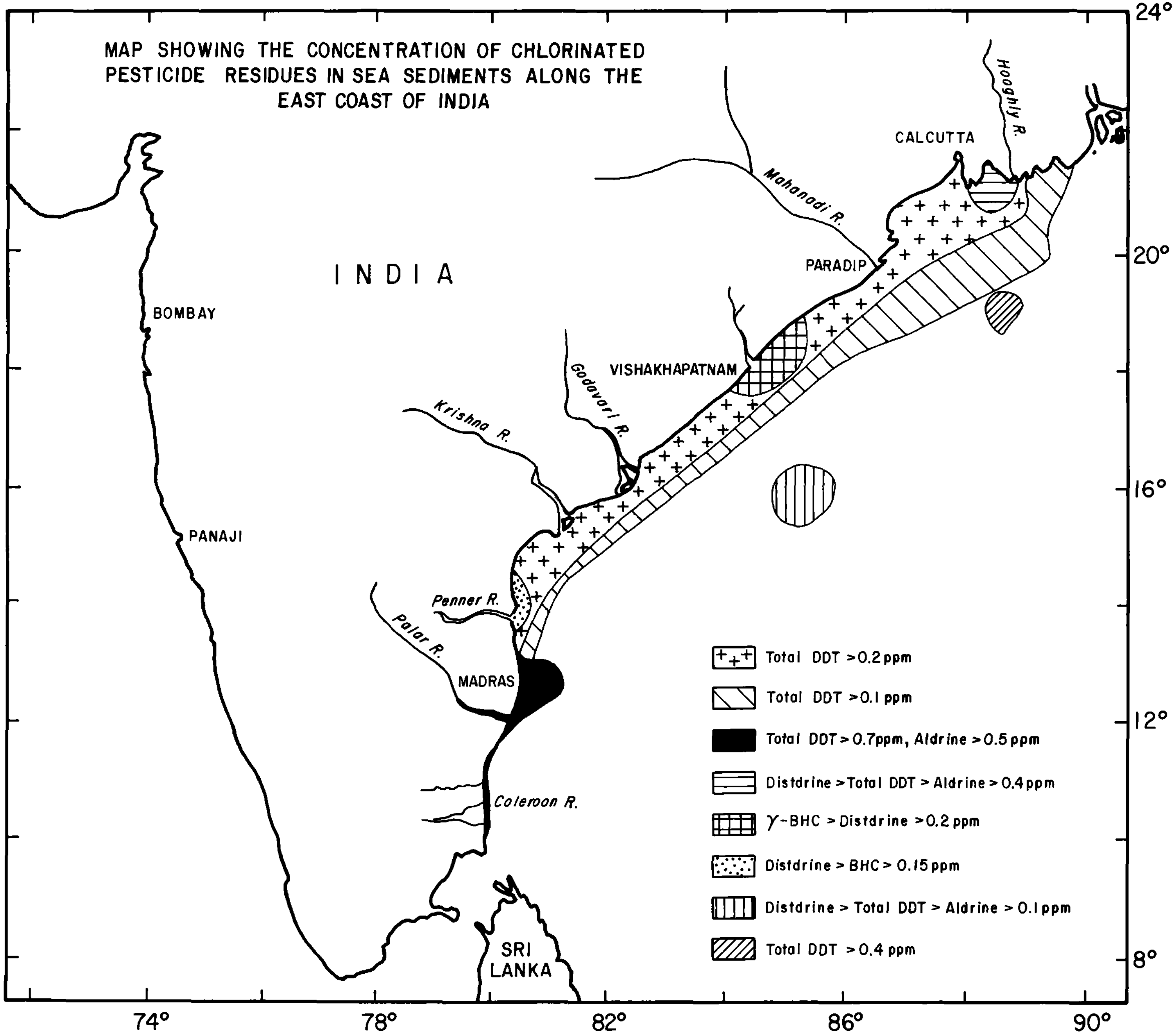

Figure 8. Chlorinated pesticide residues in sediments along the east coast of India (from SARKAR and SEN GUPTA, unpublished).

Table 1. Dispersed hydrocarbons and particulate petroleum residues (tar balls) in the Arabian Sea and Bay of Bengal (SEN GUPTA et al., 1980a; FONDEKAR et al., 1980; SEN GUPTA and KUREISHY, 1981; TOPGI et al., 1981).

Dissolved/Dispersed hydrocarbons ($\mu g\ kg^{-1}$)

	Arabian Sea			Bay of Bengal		
Coastal area	Surface	8.6—31.0	(March 1978)	Surface	0— 2.3	(June 1978)
	"	10.6—17.7	(Oct. 1983)	"	0— 4.5	(Aug. 1981)
					0— 2.8	(June 1982)
					0— 3.4	(Jan. 1984)
Tanker route	Surface	17.1± 4.1	(March 1978)	Surface	21.7± 1.0	(June 1978)
	"	21.5—42.8	(Dec. 1978)	"	75.2± 10.0	(Feb. 1979)
	"	18.6—41.6	(Dec. 1978)	"	5.3—27.4	(Jan. 1980)
	"	6.4— 9.0	(Feb. 1980)	"	6.4—11.2	(Feb. 1980)
	"	47—230	(Oct. 1980)	10 m	23.2—26.9	(June 1978)
	"	200—305	(March 1981)	"	22—94	(Feb. 1979)
	"	3.0— 8.9	(July 1983)	"	2.2—20.0	(Jan. 1980)
	"	7.4—16.2	(Sept. 1983)	"	3.1—13.6	(Feb. 1980)
	"	5.7—10.6	(Oct. 1983)	20 m	13.1—28.2	(June 1978)
	"	5.1—12.4	(Nov. 1983)	"	24—73	(Feb. 1979)
	10 m	0.9—22.9	(March 1978)	"	1.6—27.1	(Jan. 1980)
	"	16.8—42.5	(Dec. 1978)	"	1.2— 9.0	(Feb. 1980)
	"	16.9—34.5	(Dec. 1979)			
	"	3.1— 6.3	(Feb. 1980)	Range		
	"	77—210	(Oct. 1980)	Arabian Sea	Bay of Bengal	
	"	130—267	(March 1981)	0—42.8	0—28.2	
	"	0— 1.2	(July 1983)	Average Value		
	"	3.4— 8.5	(Sept. 1983)	15.8	4.6	
	"	2.1— 5.4	(Oct. 1983)			
	"	1.4— 6.1	(Nov. 1983)			
	20 m	25—28	(June 1978)			
	"	28.6—37.5	(Dec. 1978)			
	"	10.4—24.9	(Dec. 1979)			
	"	2.4— 4.5	(Feb. 1980)			
	"	0— 1.0	(July 1983)			
	"	0— 1.2	(Sept. 1983)			
	"	0— 1.1	(Oct. 1983)			
	"	0— 1.6	(Nov. 1983)			

Particulate Petroleum Residues (tar balls) ($mg\ m^{-2}$)

	Arabian Sea		Bay of Bengal	
Tanker route	0.02— 0.32	(March 1978)	0.42— 3.45	(June 1978)
	1.26— 3.46	(June 1978)	0— 1.6	(Jan. 1980)
	0— 0.54	(Dec. 1978)	0—69.8	(Feb. 1980)
	0.09— 6.0	(May 1979)	0— PI	(Jan. 1984)
	0— 0.53	(Feb. 1980)		
	0.30—112.2	(March 1981)		
	0— 0.06	(Feb. 1982)		
	0— PI	(June 1983)		
	0— —	(July 1983)		
	0— —	(Sept. 1983)		

Table 2. Ranges of Dissolved heavy metal concentrations ($\mu g\ l^{-1}$) in the Indian Ocean.

Source	Cu	Cd	Fe	Mn	Zn	Pb	Ni	Co	Hg(ng l^{-1})
Topping (1969)	0.5—49.1	—	0.1—61.8	0.1—4.6	3.9—19.5	—	—	—	—
Chester & Stoner (1974) Surface only	0.2— 1.2	0.02—0.14	0.5—3.1	0.07—0.37	0.3—3.0	—	0.3—2.6	—	—
Sen Gupta et al. (1978b)	1.7—7.9	—	7.2—66.9	—	0.5—42.4	—	0—11.5	0—6.6	—
Singbal et al. (1978)	—	—	—	—	—	—	—	—	13—187
Sanzgiri & Moraes (1979)	1.9—19.9	—	8.5—96	1.8—80	1.2—29.7	—	0—16.3	0—6.7	—
Danielsson (1980)	0.08—0.48	0.01—0.16	0.15—10	—	0.6—13.8	0.02—0.18	0.18—0.95	0—0.02	—
Sanzgiri et al. (1979)	—	—	—	—	—	—	—	—	0—204
Braganca & Sanzgiri (1980)	22—37.2	—	6.2—131.5	1.8—40.8	2.4—20	—	0—12.2	0—7.9	—
Sanzgiri & Braganca (1981)	1—5	0.15—1.9	2—21.7	1.5—24.7	1.2—12.8	0.25—7.5	0—1	0—1	—

Table 3. Ranges of concentration of a few essential and non-essential heavy metals (PPM) wet weight) in zooplankton, crustaceans, bivalve and muscles of certain fishes from the Northern Indian Ocean.

Fish	Essential Heavy Metals						Non Essential Heavy Metals		
	Cu	Fe	Mn	Zn	Ni	Co	Pb	Cd	Hg
Zooplankton	2.0-5.0	35.0-94.0	3.0-7.0	8.0-31.0	0.2-3.0	ND-4.0	1.0-12.6	0.02-5.9	ND
Prawns (6 Spp)	3.5-24.0	—	—	—	—	—	<1.0	0.2-2.5	ND-0.17
Crabs	0.7-13.5	—	—	—	—	—	<1.0-7.88	0.61-1.12	0.004-0.01
Clams	—	—	—	—	—	—	1.28	1.66	0.06
Oysters	45.0	—	—	—	—	—	<1.0	1.36	0.02
Mussels	—	—	—	—	—	—	1.31	1.38	0.09
Flying Fish	0.1-0.7	4.0-62.0	ND-3.7	4.0-21.0	ND-0.9	0.2-1.3	1.08-5.76	ND-0.65	ND-0.07
Silver Bellies	1.0-1.6	—	—	—	—	—	<1-3.21	0.58-2.11	0.001-0.01
Malabar Anchovies	4.4	—	—	—	—	—	<1	0.7	0.01
Sardines (2 Spp)	0.03	8.0-10.0	0.2	4.5-6.3	—	0.7-1.1	<1	ND-0.62	ND-0.01
Mackarel (2 Spp)	1.0-1.3	12.0	0.01	6.0	ND	1.8	<1	0.22-1.62	0.01-0.02
Jew Fish (2 Spp)	ND-0.8	6.0-8.0	0.3-10.0	4.0-4.8	ND	0.7-1.1	<1-1.14	0.19-0.42	0.006-0.01
Perch (3 Spp)	0.2-0.7	6.0-29.0	ND-0.1	3.4-6.1	0.3-0.5	ND	<1	ND-1.47	0.007-0.1
Pilot Fish	0.1-4.9	—	—	—	—	—	<1-2.95	ND-0.83	ND-0.02
Scianid (2 Spp)	0.1-0.3	—	—	—	—	—	<1	0.86-1.36	ND-0.02
Sole	—	—	—	—	—	—	<1	0.35	0.01
Pomfret	—	—	—	—	—	—	<1	0.73	0.01
Cat Fish	—	—	—	—	—	—	1.02	0.92	0.06
Trevally (2 Spp)	ND-0.7	5.0-11.0	0.1-9.0	2.0-5.0	ND-0.6	ND-1.2	<1	ND-0.62	0.018-0.08

Table 3. (continued)

Fish	Essential Heavy Metals						Non Essential Heavy Metals		
	Cu	Fe	Mn	Zn	Ni	Co	Pb	Cd	Hg
Grunter	0.36	—	—	—	—	—	2.7	ND	0.24
Talang	0.4	—	—	—	—	—	<1	ND	0.36
Tuna (4 Spp)	0.3-3.0	7.0-164.0	0.1-7.5	4.0-12.0	ND-4.0	ND-3.2	<1-3.3	ND-2.00	0.004-0.22
Dolphin Fish	0.2-1.7	13.0-39.0	ND-3.1	5.0-9.0	0.1-1.2	ND-1.9	<1-2.95	ND-0.95	0.01-0.14
Seer Fish	0.4	—	—	—	—	—	<1-1.5	0.25-0.66	0.09-0.11
Barracuda	0.1-0.5	4.0-17.0	0.2-3.1	3.3-5.8	0.1-0.3	0.6-1.9	<1	ND-0.28	0.06-0.2
Sea Pike	—	—	—	—	—	—	1.46	ND	0.11
Sharks (4 Spp)	0.14-1.1	10.0-57.0	ND-2.0	4.5-12.0	ND-0.3	ND-3.8	<1-6.02	ND-0.81	0.02-0.21

ND—Not Detectable

Sources: KUREISHY et al., 1979, 1980, 1983, Unpublished.

Table 4. Range and average concentrations of a few non-essential heavy metals (ppm wet weight) in different parts of fishes from the Northern Indian Ocean.

Body Parts	Mercury		Cadmium		Lead	
	Range	Average	Range	Average	Range	Average
Muscle	N.D.—0.36	0.07	N.D.— 3.24	0.59	<1— 3.43	1.11
Liver	N.D.—0.04	0.01	1.2—87.3	20.18	<1—17.62	3.8
Gill	N.D.—0.03	0.016	N.D.— 0.76	0.42	<1— 7.0	3.14
Heart	N.D.—0.08	0.026	N.D.— 1.91	0.54	<1— 3.4	1.36
Kidney	N.D.—0.04	0.015	0.38—36.69	9.02	<1—69.46	8.61
Gonads	N.D.—0.03	0.015	N.D.— 8.06	1.25	<1— 4.76	1.36

Sources: KUREISHY et al., 1979, 1981, 1983 (unpublished).

Table 5. Mercury pollution in and around Bombay City.

Components	Coast	Distance from shore				Thana Creek
		1 km	1.5 km	2 km	3 km	
Water ($\mu g\ l^{-1}$)	1.4—23.0	12—29	10—35	9—30	14	42
Crab muscles (ppm wet weight)	0.2— 4.1	0.1—1.5	0.2—1.0	0.3—1.3	1.8	7.3
Sediment (ppm dry weight)	0.1—27.0	—	—	—	—	38

Source: GANESAN et al. (1980).

Table 6. Population and related data and some estimates of pollutants entering the Sea around India (as of 1984).

Population	720 million
Coastal population (25% of total)	180 million
Area of the Country	3.276×10^6 km^2
Agricultural area	1.65×10^6 km^2
Exclusive economic zone	2.015×10^6 km^2
River runoff (annual mean)	1645 km^3
Rainfall per year (on land)	3.5×10^{12} m^3
Rainfall per year (on Bay of Bengal)	6.5×10^{12} m^3
Rainfall per year (on Arabian Sea)	6.1×10^{12} m^3
Domestic sewage added to the sea by coastal population per year (@ 60 1 per head per day)	3.9×10^9 m^3
Industrial effluents added to the sea by coastal industries per year	0.39×10^9 m^3
Sewage and effluents added by the river to the sea per year	50×10^6 m^3
Solid waste and garbage generated by coastal population per year (@ 0.8 kg per head per day)	53×10^6 tonnes
Fertiliser used per year (@ 30.5 kg ha^{-1} yr^{-1})	5×10^6 tonnes
Pesticides used per year (@ 336 g ha^{-1} yr^{-1})	55 000 tonnes
Synthetic detergents used per year	125 000 tonnes
Oil transported across the Arabian Sea in 1982	579×10^6 tonnes
Oil transported to Western Hemisphere in 1982	331×10^6 tonnes
Oil transported to Far East and Japan in 1982	248×10^6 tonnes
Tar deposition on beaches along the West Coast of India per year	750—1000 tonnes

Table 1A. Relation between the dissolved/dispersed petroleum hydrocarbons in the upper 30 m of the Northern Indian Ocean and oil transport along the tanker routes.

Year	Transport (mt) To west	To east	Concentrations Range (ug/kg) Arabian Sea	Bay of Bengal
1978	652	323	0.9 — 42.5	0 — 28.2
1979	659	351	10.4 — 41.6	24.0 — 75.2
1980	561	308	6.4 — 230.0	3.1 — 27.4
1981	455	270	130.0 — 305.0	0 — 4.5
1982	332	247	—	0 — 2.8
1983	291	222	0 — 16.2	—
1984	237	252	1.6 — 22.9	0 — 3.4
1985	215	232	0.7 — 31.0	5.2 — 29.5
1986	—	—	1.0 — 23.5	—
Net % decrease from 1978 to 1985	67	28		

PROBLEMS IN THE PHYSICAL OCEANOGRAPHY OF THE INDIAN OCEAN

ERIC B. KRAUS
University of Colorado/CIRES
Campus Box 409
Boulder, Colorado 80309, U.S.A.

ABSTRACT

The singular features of the large-scale water circulation in the Indian Ocean are briefly described, followed by a list of research problems that can be expected to be of regional or global interest.

INTRODUCTION

This note was presented in the framework of a workshop to provide the scientific background for the subsequent First Conference on Economic, Scientific and Technical Cooperation in the Indian Ocean. In this context it seemed appropriate to describe briefly the peculiar features of the large scale Indian Ocean dynamics. The singular nature of the water circulation there determines some of the more important problems that need to be elucidated before changes in the physical environment can be understood or predicted. The actual selection of particular research projects, which was one of the indicated objectives of the workshop, depends, of course, not only upon the physical problems, but also upon the specific interests of the scientific community, upon personnel, finance and material resources, and upon political or administrative considerations.

The Indian Ocean represents a rather unique dynamic environment because of the following features:

(1) The Indian Monsoon produces a stronger annual variation in wind forcing than is experienced by any of the other main ocean basins. The semi-annual forcing is also very pronounced. This causes seasonal variations of the near-surface zonal currents along the equator (Fig. 1).

(2) The mean westerly wind component at the 500 mb level in winter is stronger over the Indian Ocean than over the Atlantic or the Pacific (Fig. 2). The core of maximum wind speed is also somewhat further south in the same region. This suggests strong cyclonic activity at relatively high latitudes.

(3) The annual variation in wind forcing causes exceptionally large annual sea surface temperature changes near the Somali coast. Large surface temperature changes also occur in the 30°-40°S latitude belt (Fig. 3).

(4) Being closed in the north, the Indian Ocean experiences a large net southward transport of heat across the equator.

(5) The surface current along the equator tends to be convergent and not divergent as in the Atlantic or Pacific.

(6) An unusual thermohaline structure results from the intrusion of dense, highly saline water from the Red Sea and from the Gulf of Oman on the one hand and from the inflow of warm low salinity water through the Indonesian Archipelago on the other hand.

(7) The mid-latitude circulation is divided into two separate parts: a subtropical, anti-cyclonic gyre east of Madagascar and the Agulhas current system in the southwestern area of the basin.

(8) The Leuwin current along the coast of Australia represents the only known poleward eastern boundary current in the subtropics.

(9) PIOLA and GORDON (1984) estimate that the Agulhas and Leuwin currents together transport about 19 sverdrups (1 sv = 10^6 m^3/s) southward in the upper layers of the Indian Ocean. The maximum estimate of the inflow into the Indian Ocean through the Indonesian passages is 14 sv. The balance must be supplied, therefore, by upwelling fed by deep-water currents from the Southern Ocean.

(10) The subdivision of the basin by meridional ridges results in three separate deep circulation gyres with separate deep western boundary currents (See Fig. 4).

RESEARCH PROBLEMS

Some of the features listed above are not well understood quantitatively. They all involve secondary processes and phenomena that can affect economic relations. Among them are regional and global changes of the climate, the dispersion of pollutants, fluctuations in biological activity and variations in the location, and the rate of sedimentation. Most of these processes require further research for their elucidation, and some of this research depends on large-scale international cooperation.

The oceanographic problems and research topics of concern can be listed conveniently under two geographical headings:

A — The Monsoon-driven regime of the northern Indian Ocean

B — The Subtropical gyre and the Subantarctic region

The following problems in area A have attracted a good deal of scientific attention:

A1: It is not fully known to what extent the dynamics of the reversing Somali Current and the cross-equatorial western boundary flow are affected by local wind stress changes and to what extent they are determined by teleconnections with the central Indian Ocean. The phenomenon may affect the time of onset and cessation of the Indian Monsoon.

A2: The pattern and dynamics of the southward heat transport across the equator in the Indian Ocean is not well know. Quantitative information and a better understanding is needed for a better assessment of the planetary heat balance and for an improved understanding of the global climate.

A3: The onset of the Monsoon can produce very rapid changes in the wind stress pattern over the Arabian Sea as indicated by Fig. 5. During summer, the region is exposed to very large gradients of the wind stress curl (Fig. 6), juxtaposing upwelling along the Arabian and Iranian coast with strong downwelling a relatively short distance further south. This type of wind forcing must affect the oceanic circulation and turbulence. Its influence upon productivity, surface temperature and ultimately variations in the Indian Monsoon has not been fully investigated so far.

A4: Surface currents tend to reverse also in the region south of India/Sri Lanka. It appears that these currents carry about three times more water eastward in summer than westward in winter. It is not known where and when the resulting net inflow of water into the Bay of Bengal is compensated.

A5: The semi-annual variation of the equatorial current system has been the subject of investigations by McPHADDEN (1982), GENT et al. (1983) and REVERDIN and CANE (1984), but the subject deserves further observational and theoretical study.

A6: Systematic observations of the water transport through the Indonesian passages are now being planned under a joint U.S./Indonesian program. However, the projected duration of this program may be too short to resolve the dependence of this transport upon interannual sea-level changes in the Pacific. (Southern Oscillation/El Niño).

A7: Precursors of the Southern Oscillation/El Niño phenomenon in the Indian Ocean have been stipulated or derived from statistical evidence by several investigators. If the existence of these precursors could be confirmed, this could have considerable predictive value.

The southern area of the Indian Ocean has not been explored intensively. Among the open research problems there one might list the following topics:

B1: What is the fate of the water transported by the Agulhas Current and what is the nature of the Atlantic/Indian Ocean water exchange? (See e.g., DE RUIJTER, 1982; OU and DE RUIJTER, 1986.)

B2: How does the Agulhas Current interact with the anticyclonic gyre east of Madagascar?
B3: Where and how are water masses (mode waters) formed in the Indian Ocean?
B4: What is the origin of the water transported by the Leuwin Current off western Australia? (See e.g., GODFREY and RIDGWAY, 1985.)
B5: The circulation in the Subantarctic region is only known qualitatively. The heat exchange between the Indian Ocean and the Circumpolar Current system has apparently never been assessed by physical methods.
B6: Figure 4 illustrates the existing uncertainties in our knowledge about deep meridional currents in the Indian Ocean.

REGIONAL AND INTERNATIONAL STUDIES

All the research problems listed above do affect also various current or planned international programs such as the Tropical Ocean and Global Atmosphere (TOGA) project, the World Ocean Circulation Experiment (WOCE) and the International Geosphere/Biosphere Program (IGBP). As such they are of some concern to all the participating countries and to the global scientific community. Regional studies within the framework of these international programs should provide scientists from the Indian Ocean littoral states with opportunities for cooperation and international support.

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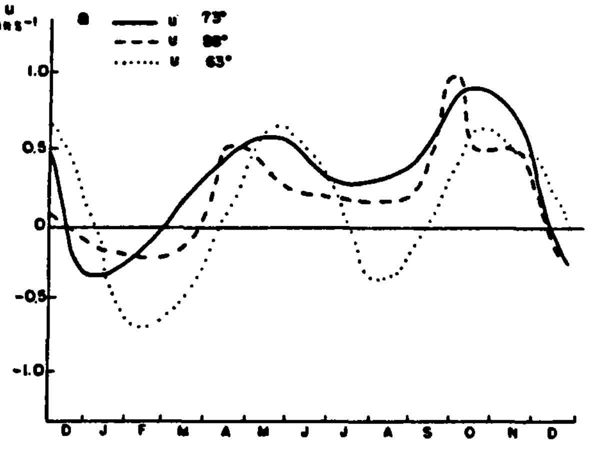

Figure 1. The seasonal variation of near-surface zonal current speeds along the equator at different longitudes (after McPHADDEN, 1982)

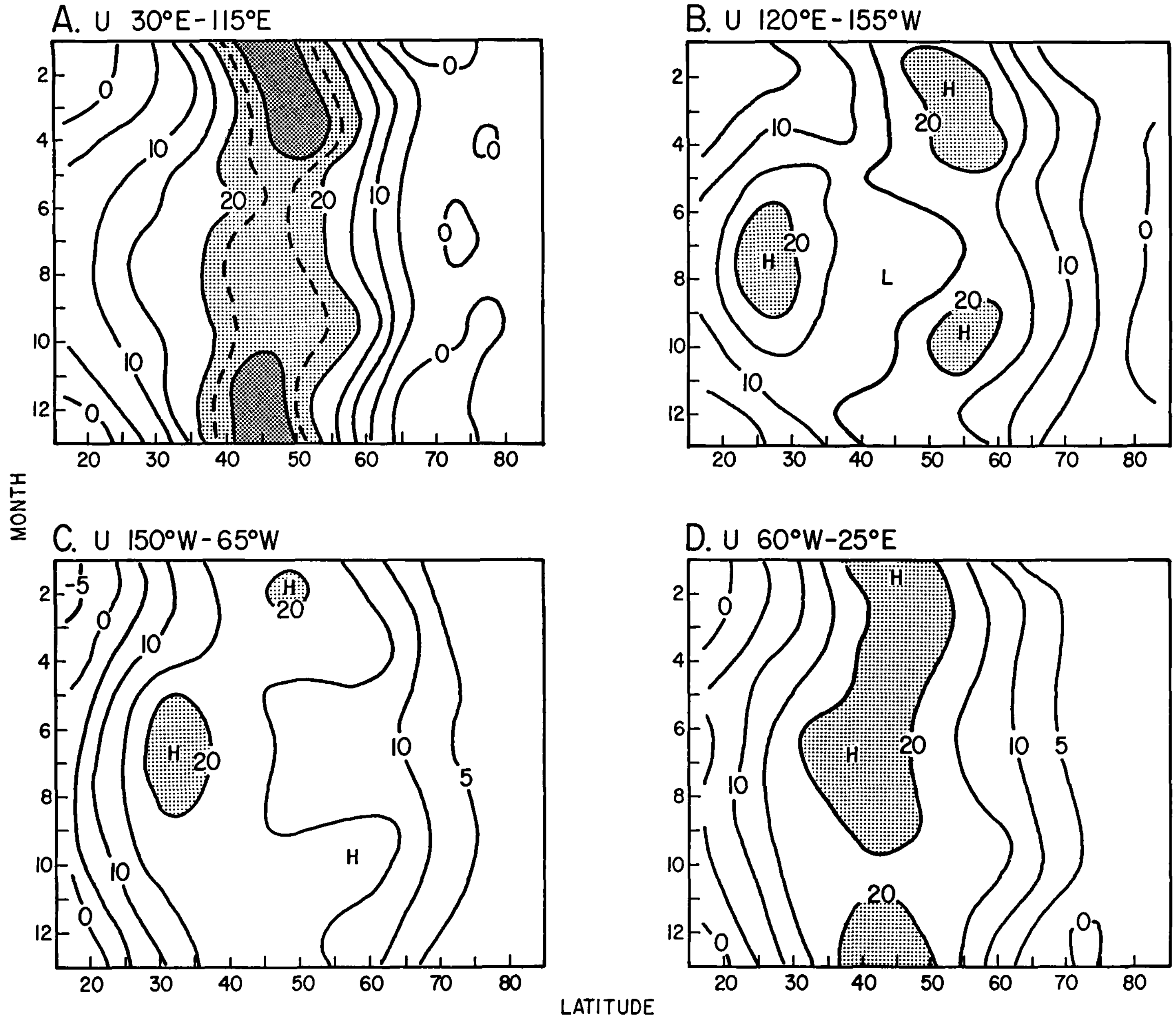

Figure 2. The mean annual cycle of the westerly wind component in m/s averaged over 90° sectors at 500 mb as a function of month and latitude: (a) Indian Ocean, 30-115°E, (b) Australasia, 120°E-155W, (c) Pacific, 150-65°W, (d) Atlantic, 60°W-25°E (from TRENBERTH, 1979).

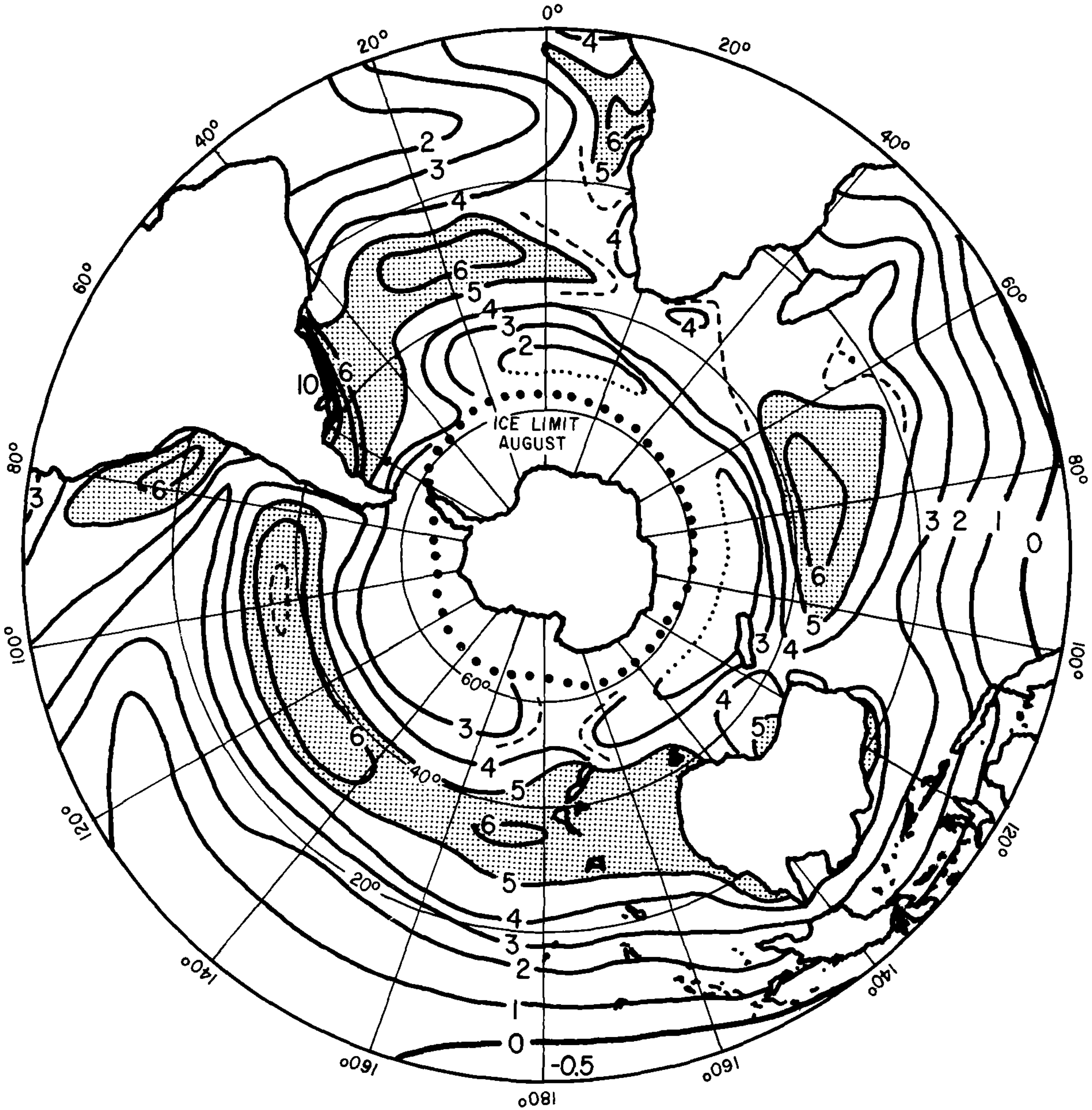

Figure 3. Sea surface temperature range (February-August) in the southern hemisphere (from VANLOON, 1972).

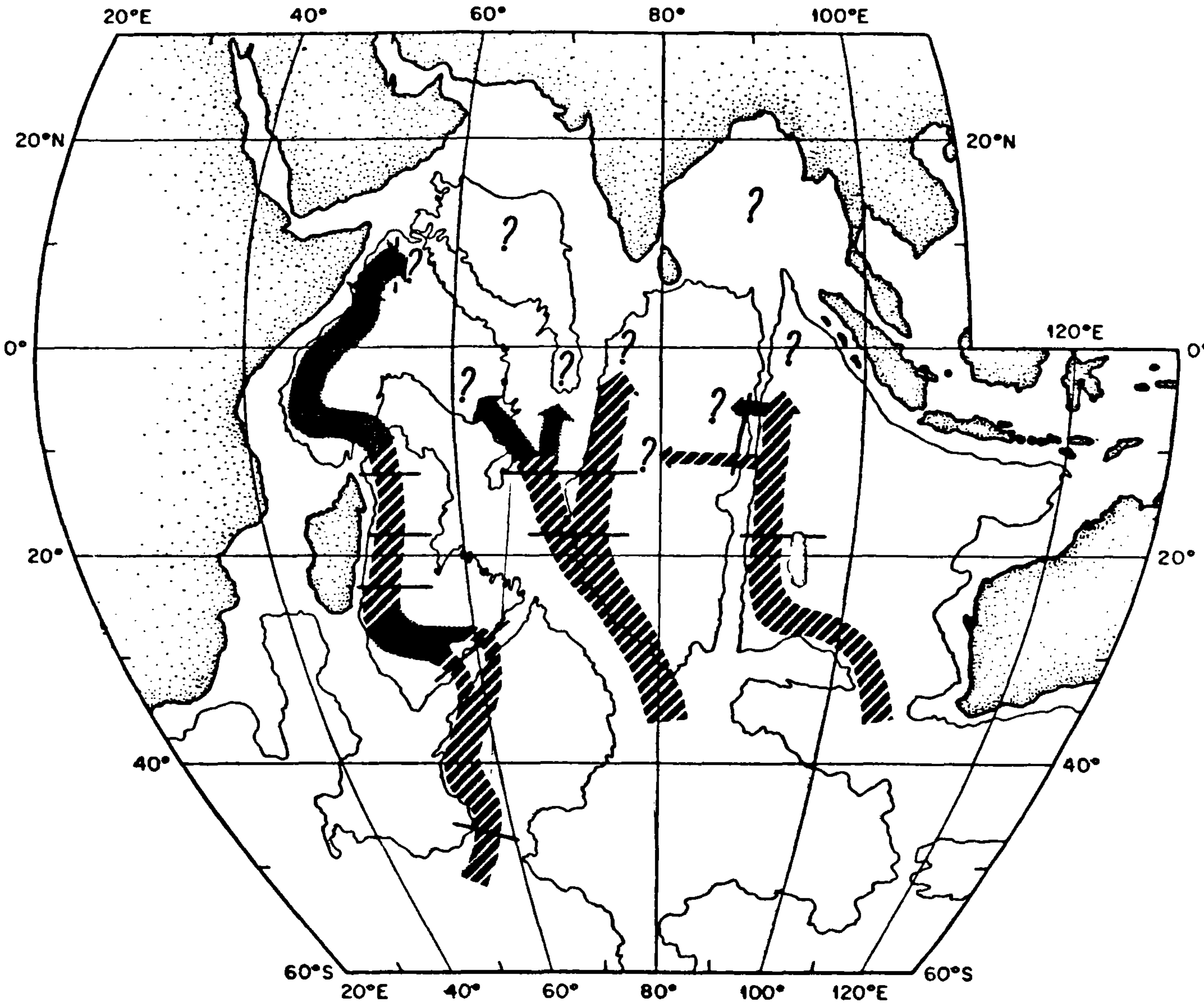

Figure 4. Meridional deep currents. The existence of currents represented by striped bands has been established; the existence of those represented by dark bands is conjectural. (From B. WARREN).

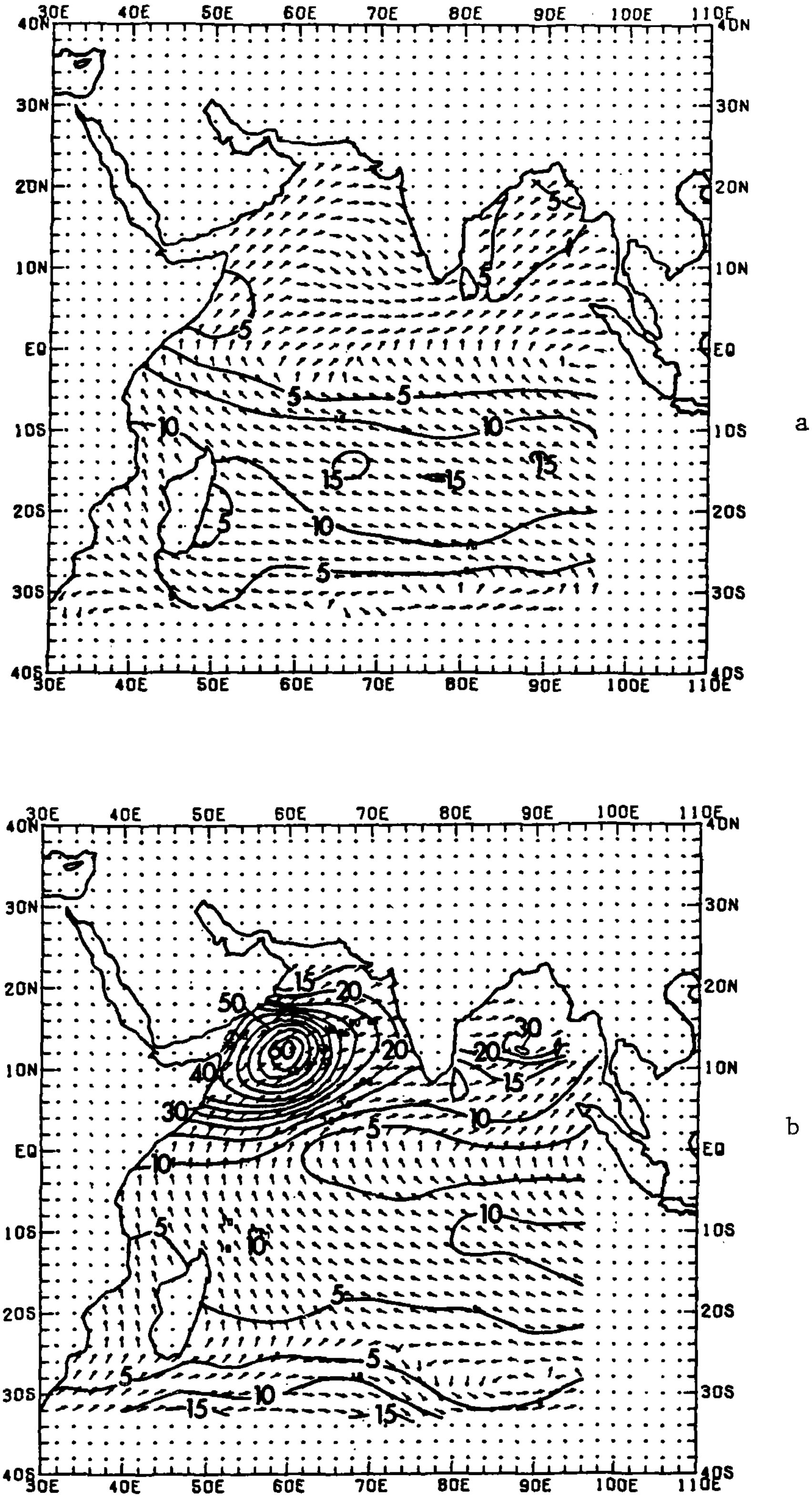

Figure 5. Wind stress in 0.01 Newton/m² or 0.1 dyne/cm² averaged over ten days periods: (a) 31 May-10 June, 1979 and (b) 20-29 June, 1979. (From WYLIE and HINTON, 1982).

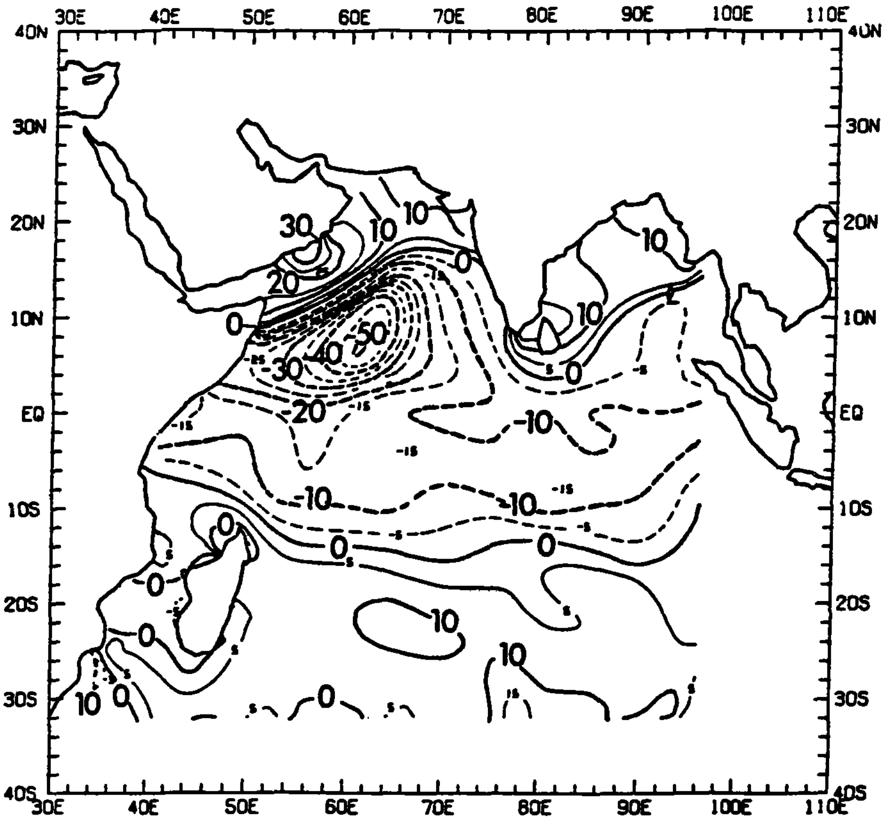

Figure 6 Average wind stress curl for July 1979 in 10^{-8} N/m³. Negative values are dashed. (From WYLIE and HINTON, 1982).

PHYSICAL OCEANOGRAPHY OF THE NORTHERN INDIAN OCEAN — A PERSPECTIVE

V.V. VARADACHARI

National Institute of Oceanography
Dona Paula, 403 004 Goa
India

ABSTRACT

The northern Indian Ocean, including the Bay of Bengal, the Arabian Sea and the adjacent gulfs, is unique in respect of its physical oceanography and meteorology because of the monsoons. Nowhere else does the ocean undergo such regular, widespread and large-scale seasonal changes in wind forcing and consequent changes in the characteristics of the ocean, especially the circulation. The uniqueness of the northern Indian Ocean and the monsoons led to several international expeditions such as IIOE, ISMEX, MONEX etc., which provided basic data on the oceanography and meteorology of the area. In spite of all these expeditions, our knowledge of the area is still quite meagre.

In order to understand the interannual variability of the monsoons and the mechanisms that determine their variability and predictability, several components have been identified in the TOGA (Tropical Ocean and Global Atmosphere) and WOCE (World Ocean Circulation Experiment) programme defined by CCCO (the Joint SCOR-IOC Committee on Climatic Changes and the Ocean) for the northern Indian Ocean. Some of the important elements of the programmes are the seasonal cycles of the Somali current system and the equatorial current system, the Arabian Sea surface temperature variations, cross-equatorial fluxes, the surface-layer dynamics, the exchange of energy between the ocean and the atmosphere etc. The sea-level variations, the dynamics of eddies and the transport processes, vertical and horizontal mixing, temperature and salinity distribution and the upwelling dynamics of the area are also of significance in our understanding of the interannual variability of the monsoons. The monsoon-dependent ocean circulation and other physical characteristics of the waters of the area have important impacts on other characteristics, such as biological productivity, distribution of nutrients, etc.

In addition to the climatic variability of the area, studies on problems such as the genesis of cyclones in the Bay of Bengal, storm surges on the east coast of India, etc., will be of great interest and importance.

THE DESTRUCTION OF THE TETHYS AND PALEOCEANOGRAPHIC DEVELOPMENT OF THE INDIAN OCEAN

BILAL U. HAQ
Exxon Production Research Company
Houston, Texas 77001

ABSTRACT

The floor of the Indian Ocean was formed at the expense of the ancient circum-global Tethys Seaway. The destruction of the eastern limb of the ancestral Tethys and the evolution of the modern Indian Ocean were simultaneous and intimately related. The western and eastern sides of the ancient Tethys were first connected some 175 million years (m.y.) ago when the Central American passage first opened. But the development of the Indian Ocean can be said to have begun some 210 m.y. ago, with the initial breakup of the Gondwanaland supercontinent. Paleogeographic events such as the separation of Madagascar from Africa, the rifting between Australia-Antarctica and Africa, and the split between Antarctica and Australia were important thresholds in the development of the Indian Ocean. However, the most consequential event that help form this new basin and that led to the demise of the Tethys Seaway was the rapid movement of the Indian Plate towards Eurasia that began about 80 m.y. ago. The circulatory and climatic patterns during these developments differed significantly from their modern counterparts and were strongly influenced by the continental positions and opening and closing of surface and deep water connections. The patterns of sedimentation in turn were influenced by the changing hydrographic and climatic patterns.

The collision of India with Eurasia, some 50 m.y. ago, sealed the fate of the ancestral Tethys and the eventual closure of the oceanic pathways north and east of the Indian Plate. The eastern Tethys was essentially closed by 30 m.y. B.P., and the development of the present-day climatic and circulation patterns, and their attendant coastal regimes, drainage and sedimentary patterns, were initiated by this time.

INTRODUCTION

Prior to the evolution of the modern day Indian Ocean and the establishment of circum-Antarctic circulation in southern high latitudes, a circum-global seaway — the Tethys — existed in the low and temperate latitudes. The Tethys extended westward, from the space now occupied by northern Indian Ocean, to the Pacific through Eurasia and the nascent gap between North and South America. The westward flowing current through the Tethys was the main source of hydrographic, climatic, and biotic dispersals for some 140 million years, from the late Jurassic through early Tertiary. The events that led to the formation of the Indian and Southern Oceans were also responsible for the destruction of the ancestral Tethys and the shift of circum-global hydrographic dispersal from the low latitudes to southern high latitudes around Antarctica. This shift in the locus of circum-global circulation caused major changes in the nature of surface and deep water circulation and global climates.

Inspite of its relative youth and smaller size compared to the other major ocean basins, the tectonic history of the Indian Ocean is quite intricate. Its floor has resulted from a complex spreading history, involving numerous rotations of its component plates, and several episodes of creation, relocation and/or destruction of spreading centers. But since the basin is bounded by a single major trench to its northeast, the subduction in the remainder of the basin has been minimal and the seafloor readily yields its secrets. Magnetic anomaly data, diligently gathered over the past decade and a half, have led to the deciphering of the major events that shaped the present basin.

In addition, a large body of sedimentological and paleontological data was collected by the Deep Sea Drilling Project (DSDP) in the Indian Ocean. During 1972-73, DSDP drilled 58 sites in the Indian Ocean and the part of Southern Ocean to its south. The stratigraphic information from these drill sites has added valuable information about their antiquity and sedimentation patterns. These have helped in the piecing-together of the sedimentary, hydrographic and climatic histories of this basin.

The purpose of this paper is to outline the geological and oceanographic-climatic evolution of the Indian Ocean since 150 Ma (Ma = m. yrs. B.P.), from the time of the breakup of eastern Gondwanaland to the development of modern day patterns. Sedimentary and climatic history of the basin are outlined and inferred global paleoceanographic patterns, in which the Indian and Southern Ocean play a major role, are also presented. In this paper the paleogeographic events that led to the destruction of the Tethys Seaway and the development of the Indian Ocean, and the oceanographic-climatic evolution of the region will be reviewed. The sources for this review are varied and based on geophysical, sedimentological and paleontological evidence from the deep sea, as well as stratigraphic data from the continents surrounding the Indian Ocean.

In order to discern the sequence of events that led to the development of this complex basin, it may be helpful to first examine its present-day geologic setting and sedimentary patterns.

MODERN GEOLOGIC SETTING

The ocean floor in the Indian Ocean is dominated by the mid-ocean ridge system where the seafloor is actively spreading at the present time (Fig. 1). This spreading ridge system forms an inverted Y, whose three limbs are offset by extensive fracture zones (see paper in this volume by SIDDIQUE et al.). The limbs extend: 1) first north and then northwest towards Gulf of Aden, forming the Central Indian and Carlsberg Ridges, 2) southwest between Africa and Antarctica, extending into South Atlantic and forming the Southwest Indian Ridge, and 3) southeast between Australia and Antarctica forming the Southeast Indian Ridge. The present-day half spreading rates range from about 1 cm/yr on the Southwest Ridge, and between Africa and Antarctica, to about 2 cm/yr on Central and Southeast Ridges. These rates are comparable to rates in other ocean basins.

Also prominent on the Indian Ocean floor are two aseismic ridges and several topographic highs or plateaus with little or no seismic activity at the present time. The Ninetyeast Ridge (90E) is the most prominent of these features (Fig. 1). This presently "quiet" ridge is the longest aseismic feature in the world ocean and extends from the Bay of Bengal to 32°S along 90°E meridian. It has been shown to have a mixed tectonic and volcanic origin, being a paleo-transform fault with volcanic leakage, and forming a trace of the northward movement of Indian Plate. The age of 90E Ridge decreases progressively away from its northern end, ranging from late Cretaceous (ca 85-65 Ma) in the north to Oligocene (ca 30-35 Ma) in the south.

The second aseismic ridge — the Chagos-Laccadive Ridge — extends from the western margin of India to about 10°S, where it is truncated by the Central Indian Ridge. Like the 90E Ridge, the Chagos-Laccadive Ridge is also a former transform fault and a leaky western trace of the Indian Plate's northward movement (McKENZIE and SCLATER, 1971).

The more prominent oceanic highs in the eastern and southern Indian Ocean include the Broken Ridge, and the Naturaliste and Exmouth Plateaus in the east, off Australia, and the Kerguelen Plateau south of the inverted Y-shaped spreading ridge. Naturaliste Plateau has been in existence at least since the mid Cretaceous (ca 100 Ma) and was a shallow platform receiving terrigenous sediments from surrounding continents in its early history. Later on pelagic carbonate sedimentation replaced terrigenous input. Much of the early Tertiary record, however, has been removed by subsequent erosion (LUYENDYK and DAVIES, 1974).

Broken Ridge maybe as old as the Naturaliste Plateau. In the late Cretaceous it was a shallow carbonate platform that probably split from the Kerguelen Plateau during Eocene when the Southeast Indian Ridge originated. Since then subsidence and pelagic carbonate sedimentation have characterized the Broken Ridge. The Kerguelen Plateau most probably had an oceanic origin as the rocks from Kerguelen Island show similarities to other oceanic islands (UPTON, 1982). Seismic data from the Exmouth Plateau suggests that its basement is at least as old as Triassic. Although some authors have argued in favor of an oceanic origin for this plateau (see e.g., VEEVERS and COTTERILL, 1978), the question whether it is of oceanic or continental origin has not been fully resolved.

In the western Indian Ocean the major topographic features include the Mascarene (or Seychelles-Mascarene) Plateau and the Madagascar Ridge, northeast and south of Madagascar, respectively; and the Mozambique Ridge, off the eastern coast of South Africa. At least the northern part of the Seychelles-Mascarene block is of great antiquity, as much as 600 m. yrs. old (FISHER and BUNCE, 1974). It separated from the Indian Plate in early Paleocene and has subsided steadily since then (NORTON and SCLATER, 1979). Drilling in the middle of Mascarene Plateau has shown that since its separation it has been a site of pelagic carbonate sedimentation.

The Madagascar Ridge has long been considered to be a submerged continuation of Madagascar, but DSDP drilling on the ridge did not resolve the issue. The sedimentary record shows a drape of pelagic carbonates over early Eocene basaltic pyroclastic sediments, the latter pointing to volcanic activity in the earliest Tertiary (SIMPSON et al., 1974).

Drilling on the Mozambique Ridge revealed that it has an oceanic-type basaltic basement, somewhat older than Valanginian (ca 130-135 Ma). In the late Cretaceous it was a site of extensive volcanism, but since then the ridge has been inactive (VALLIER, 1974). It was first the site of deposition of carbonate sands, and later pelagic carbonate oozes.

The deeper basins of the Indian Ocean contain fairly well delineated seafloor magnetic anomaly patterns that serve as important clues to the antiquity and spreading history of these basins. Major basins include, the Wharton Basin, east of the 90E Ridge; the Central Indian Basin, sandwiched between the 90E and Chagos-Laccadive Ridges; the Crozet Basin, south of the inverted Y-shaped spreading center; Somali and Madagascar Basins, north and east of Madagascar, respectively; and the Mozambique Basin, between the Madagascar and Mozambique Ridges. The oldest identifiable magnetic anomalies on the Indian Ocean floor are of late Jurassic to early Cretaceous age.

Other important features of the Indian Ocean include the Indus and Bengal Fans (Fig. 1), both of which comprise thick piles of terrigenous sediments, deposited since the collision of India with Eurasia. The latter event is discussed in more detail in the next section.

MODERN SEDIMENTARY PATTERNS

The sedimentary patterns of the Indian Ocean have been summarized on a regional basis in Initial Report volumes of the DSDP Legs 22 through 27, and on an oceanwide basis by KOLLA and KIDD (1982), KIDD and DAVIES (1978) and DAVIES and KIDD (1977).

In general, land-derived (terrigenous) sediments (mainly muds and sands) accumulate along the continental margins, brought down by the rivers. Major areas of terrigenous input are the Indus and Bengal submarine fans, and to a lesser degree the Zambesi Fan, off central Africa coast. By far the thickest and the most extensive pile of sediments is that of the Bengal Fan. Seismic data shows this fan to be as much as 12 km thick (CURRAY and MOORE, 1974), extending from the mouth of Ganges River to the Ceylon Abyssal Plain in the northern part of the Central Indian Basin. The Indus Fan is also of considerable extent, reaching up to the Carlsberg Ridge, with a maximum thickness of over 10 km (NAINI and KOLLA, 1982). Both of these fans receive their great load of sediments eroded from the relatively youthful mountain ranges of the Himalayas. The main locus of volcanogenic sediments is along the southern margin of the Indonesian Island Arc, which contains some 14% of the worlds active volcanos. The material from this island arc is silicic in composition (KIDD and DAVIES, 1978).

The Indian Ocean carbonate compensation depth (CCD — below which all carbonate is dissolved and which separates the calcareous oozes from clays) lies at an average depth of about 4000 meters. Thus carbonate accumulation is restricted to the relatively shallower parts of the deep basins and on ridges and plateaus. Below the CCD, in deep basins, siliceous sediments and clays accumulate. The biogenic siliceous sediments are, however, more characteristic of the equatorial and subpolar high productivity areas. The isolated deeper basins, away from the continental influences or high productivity belts, receive negligible amounts of sediment. Examples of such areas are, the deeper Wharton Basin, Crozet Basin and the southern Mascarene Basin. Where deep currents are active and sedimentation is minimal, manganese nodule pavements may develop, e.g., between the Broken and Southeast Ridges (DAVIES and KIDD, 1977; KIDD and DAVIES, 1978).

In the northern Indian Ocean the carbonate content of the sediments increases gradually towards the south with the decreasing dilution by terrigenous input. In the Arabian Sea biogenic productivity is highest along the Arabian Peninsula and in the Gulf of Aden, which is reflected in the high opal and biogenic siliceous content of the sediments. Organic matter values in the sediments are highest, however, along the western margin of India, facilitated by better preservation due to oxygen-poor environments (KOLLA et al., 1982).

GEOLOGICAL HISTORY OF THE INDIAN OCEAN

Seafloor magnetic anomalies and paleomagnetic data from the surrounding continents provide important information about the age of the seafloor and the paleoposition of the continents at different times. Sedimentary patterns furnish additional clues about circulation, climates, productivity and the continental influences on the ocean and its contained biota. Both these lines of evidence have been valuable in deciphering the broad picture of the geological history of the Indian Ocean.

The general outline of the evolution of the Indian Ocean basin from magnetic evidence was first presented by McKENZIE and SCLATER (1971). Later studies, including those by SCLATER and FISHER (1974), PIMM et al. (1974), VEEVERS and McELHINNY (1976), LARSON (1977), SCHLICH et al. (1977), LUYENDYK et al. (1980), CANDE and MUTTER (1982), CURRAY et al. (1982) and RABINOWITZ et al. (1983) have all added details on a regional or oceanwide basis. Drilling and geophysical data collected during DSDP Legs 22 through 27 have further enlarged the database and a considerably detailed picture of the development of Indian Ocean has emerged. The history of this youngest of the major oceans begins with the last phase of the breaking up of the supercontinent of Gondwanaland.

THE BREAKUP OF GONDWANALAND AND THE MESOZOIC DEVELOPMENT OF THE INDIAN OCEAN

The Great Slip

The major Mesozoic events in the Indian Ocean are summarized in Figure 2. The oldest identifiable magnetic anomaly in the Indian Ocean (in the Mozambique Channel) is of late Jurassic age. Thus the actual separation of eastern Gondwanaland (Madagascar-India-Antarctica-Australia) from the western Gondwanaland (Africa-South America) predates anomaly M22 (ca 140 Ma), and probably occurred some 150 m.y. ago. The initial rifting between the two halves of the supercontinent, however, may have occurred as early as early Jurassic (ca 210 to 200 Ma) when first open marine conditions are encountered in eastern Africa (NORTON and SCLATER, 1979). The Karroo volcanics in South Africa and the Dufek intrusion on Antarctica are both of early Jurassic age and most likely mark the timing of the initial rifting phase between East and West Gondwanaland (NORTON, per. comm., 1985). Early motion between East and West Gondwanaland was predominantly strike-slip. The large intercontinental strike-slip faults probably restricted oceanic circulation until well into the Cretaceous.

In the Mesozoic the area north of eastern Gondwanaland was occupied by the Tethys Sea, which extended up to the Asian mainland just north of the tropical latitudes. During the Jurassic this seaway extended westwards into Europe, but in the early and most of the middle Jurassic, the circum-global Tethys Current had not yet initiated as the Central American passage between North and South America was still blocked. This passage was breached sometime in the later part of middle Jurassic, thereby connecting the eastern Tethys with the Pacific, via the proto-North Atlantic. An epicontinental seaway existed prior to the opening of the Central American passage, but communication with the open ocean was restricted, except during times of high seastands. Biogeographic data suggest the establishment of a permanent connection with the Pacific in Bathonian time (ca 175 to 170 Ma). The westward flowing Tethys Current may have been developed soon afterwards (HAQ, 1984).

Much of the old Tethys seafloor has since been lost to subduction. The only retrievable record comes from the margins of this seaway, from southern Eurasia and northern Africa. Evidence from these shores indicates that great volumes of carbonate sediments were deposited in this tropical-temperate seaway (McGOWRAN, 1978) where open marine conditions existed. Clays accumulated in the more restricted basins (KIDD and DAVIES, 1978).

The Great Split

The late Jurassic seems to have been the time of the great split when South America separated from Africa, beginning the opening of the South Atlantic. As mentioned earlier, East and West Gondwanaland had already separated by this time (NORTON and SCLATER, 1979). Madagascar, to which India was still firmly attached, had begun moving away from Africa in the late Jurassic (RABINOWITZ et al., 1983), some 140 m.y. ago. The spatial constraints suggest that Madagascar could not have reached its present position relative to Africa until anomaly MO time (ca 110 Ma) as suggested by NORTON and SCLATER (1979) and SIGOUFIN and PATRIAT (1981).

The breaking-up of eastern Gondwanaland predates anomaly M4 (ca 115 Ma), which is the oldest anomaly identified off western Australia (LARSON, 1977; markl, 1978). JOHNSON ET AL. (1980) date the inception of the spreading between greater India and Australia somewhat earlier, at anomaly M10N (ca 102 Ma).

Prior to the breakup, "greater" India occupied the space just west and northwest of Australia. LARSON (1977) included Tibet as a part of greater India, which abutted western Australia as far north as the Exmouth Plateau. Tarim may also have been located north of Australia and east of Tibet (LARSON, 1975). Recent Chinese paleomagnetic studies in southern Tibet show, however, that Tibet lay about 8° north of the equator and was not a part of India since at least the late Cretaceous time, when India was still attached to Madagascar (MOLNAR and CHEN, 1978).

The separation between Australia and Antarctica had been thought to have begun in the late Paleocene (ca 55-53 Ma). A careful reinterpretation of the magnetic anomalies between the two continents by CANDE and MUTTER (1982) doubled the age of this event, which may have occurred some 50 to 55 m.y. earlier, within the magnetic quiet zone preceding anomaly 34. A more up-to-date time scale used here (Fig. 2) places the magnetic interval in question somewhere between 105 and 84 Ma. It is quite likely that this event was a part of the same major breakup event that started the split of India from Madagascar some 90 m.y. ago. The latter event is associated with a suite of late Cretaceous lava flows along Madagascar's eastern coast (NORTON and SCLATER, 1979).

According to CANDE and MUTTER (1982) the first phase of spreading between Australia and Antarctica that lasted from just before anomaly 34 to anomaly 19 (ca 43 Ma) was very slow. The breakup was accompanied by rapid subsidence of sedimentary basins along the southern margins of Australia. This breakup and spreading initiation between India and Madagascar (and between Australia and Antarctica) marks a major reorganization of plate boundaries and a reorientation of the spreading direction of some 60°. At this time, or somewhat later (around 80 Ma), the spreading in the Tasman Sea and between New Zealand and Antarctica was also initiated (WEISSEL et al., 1977; NORTON and SCLATER, 1979).

As the pieces of the old East Gondwanaland started to drift apart, the spreading created new seafloor that would become the modern Indian Ocean. This spreading progressively constricted the Tethys, and the eventual demise of the seaway had begun. Each separation event (rifting and spreading) during the breakup of East Gondwanaland was preceded by periods of uplift and erosion, and deposition of thick synrift terrigenous sediments in the rift valleys. These coarse non-marine sediments of the early phases, overlain eventually by finer, organic-rich, marine sediments, form excellent oil reservoirs, e.g., along the southern and western margins of Australia (VEVEERS and McELHINNY, 1976).

The early sedimentation within the newly created Indian Ocean basins were characterized by accumulation of clays in the restricted basins and the deposition of calcareous sediments in the areas higher than the CCD. Terrigenous sedimentation was confined along the margins of the continents (KIDD and DAVIES, 1978).

Models of oceanic circulation can be constructed by employing both the epirical data and analogies from the present-day hydrographic and meteorological principles. The reconstructions can be reconciled iteratively with the known stratigraphic data and modified, if necessary. Climatic scenaria of the distant geological past can be constructed indirectly from the sedimentary and fossil record and from the paleogeographic and oceanographic reconstructions. When available, the oxygen-isotopic record, pieced together from marine microplankton and benthos, can yield direct paleotemperature data.

The oxygen-isotopic ($^{18}O/^{16}O$) ratios of calcareous plankton and benthos can yield information about the surface, deep and bottom water temperatures at the time the organisms were living. From these data the temperature structure of the water column can be constructed. Carbon isotopes, on the other hand, supply information about the marine fertility and productivity, as their ratios ($^{13}O/^{12}O$) are dependent on the supply and storage of organic carbon.

Oxygen-isotopic data from the Jurassic are lacking, but both the sedimentary and fossil evidence point to generally warm global climates, with relatively dry conditions in the early and middle Jurassic. Well-developed reefs along the Tethyan margins indicate warm climates in this embayment (FRAKES, 1979).

The sparse oxygen-isotopic data from the Cretaceous suggest that the global climates remained warm but became somewhat more equable than the Jurassic. Humidity may have increased due to the generally high seastands. A climatic optimum occurred from the Albian to Santonian times (110-80 Ma). The vertical thermal structure of the water column was relatively homogeneous, with a low thermal gradient (DOUGLAS and SAVIN, 1978). Atmospheric modeling of the mid Cretaceous predicts polar surface temperatures from about 15 to 19°C, with tropical temperature only a few degrees above the present (BARRON and WASHINGTON, 1982). This simulation also shows a belt of variable high and low pressure zones in the mid latitudes along the edge of the continents bordering the Tethys, related to land-sea thermal contrasts. The northern margins of Tethys are believed to have received extensive rainfall during this time. Such conditions would have produced strong monsoonal-type circulation and excessive evaporation over the mid and low latitude epicontinental seas. Thus these sites over tropical-temperate latitudes may have been the source of Warm Saline Bottom Waters (WSBW) that characterized the Cretaceous deep water regime (BRASS et al., 1981).

To appreciate the circulation within the ancestral Tethys and in the developing Indian Ocean, it is necessary to see it in a global context. The mid-Jurassic and mid-Cretaceous circulation models before and after the opening of the Central American passage and the initiation of the Tethys Current have been presented by HAQ (1984) and need not be repeated here (Figs. 3 and 4). These patterns show the eastern Tethys to be dominated by an anti-clockwise gyre with an extension into the western Tethys that may have reached the proto-North Atlantic during high seastands. After the Central American opening, the Tethys Current became the dominant feature of the world ocean and the main source of hydrographic, climatic and biotic dispersals (Fig. 4). Circulation in the newly opened Indian Ocean south of India was still restricted to some surface communication with the Tethys.

The Rapid Flight of India

India broke loose from Madagascar about the same time as Australia separated from Antarctica, around 90 m.y. ago. Initial rate of the northward flight of India was slow, but rapid spreading began around 80 Ma. Half spreading rates between magnetic anomalies 33 and 28 averaged about 6 cm/yr. At anomaly 28 time (64 Ma) the rates increased to between 7 and 8 cm/yr, and around anomaly 24 time (ca 53 Ma) they accelerated further to about 18-19 cm/yr. The rate slowed down dramatically after anomaly 23.

Figure 5 shows a comparison between the half spreading rates in the three major ocean basins. The Indian Ocean rates stand out from the other basins in showing the greatest variation in the spreading rates, being exceptionally high almost throughout the latest Cretaceous and earliest Tertiary. Somewhere between anomaly 23 and 21 (52-48 Ma) the rates decelerated and a dramatic reduction occurred after anomaly 21, when India's northward motion nearly came to a standstill. This is the time when there was little or no movement along the 90E Ridge (McKENZIE and SCLATER, 1971). The event most certainly marks the first encounter of the Indian Plate with Eurasia. Up to this time the Asian mainland's marginal subduction zone may have been actively consuming oceanic lithosphere, until the more dense continental crust of the northern Indian Plate was encountered. Magnetic data in Figure 5 narrow the date of this event to around anomaly 22 time (ca 50 Ma). This still ongoing collision of India with Asia is a natural laboratory in which the implications of a continent- continent type conjunction can be studied.

As mentioned earlier, the 90E Ridge is a long (4500 km) and narrow (50-100 km) aseismic feature with an average height of about 2 km above the seafloor. Along its eastern edge is a sharp escarpment. The Ninetyeast Ridge in the east and Chagos-Laccadive Ridge in the west were both formed during the rapid movement of India northwards. MORGAN (1972) was first to propose that Ninetyeast Ridge was formed as the Indian Plate moved over a fixed mantle plume-hotspot of the Kerguelen Islands. SCLATER and FISHER (1974), on the other hand, preferred to think of it as originating from excessive volcanism at a migrating spreading center-transform fault junction. They modeled the Ninetyeast and Chagos-Laccadive Ridges as transform faults, marking a paired offset of an active spreading center. The transform faults terminated as active faults when the spreading direction changed from N-S to NE-SW, around anomaly 11 time (SCLATER and FISHER, 1974).

LUYENDYK and RENNICK (1977) suggested that Ninetyeast Ridge was laid down mainly from the Amsterdam - St. Paul hotspot. According to these authors the Broken Ridge and its Naturaliste Plateau extension, as well as the Kerguelen Plateau, were once joined and were formed from the Kerguelen hotspot. Kerguelen and Broken Ridge rifted and separated once spreading began between Australia and Antarctica.

PIERCE's (1979) study of the paleolatitudes of basement rocks from the Ninetyeast Ridge and from the Deccan Traps from India suggests that the source of volcanic rocks on the Ridge had remained constant for some 60 m.y. near 50° S latitude, the present location of the Kerguelen hotspot. Thus the genesis of the Ninetyeast Ridge seems to be intimately linked to the Kerguelen hotspot. The Amsterdam - St. Paul hotspot may, in fact, also be a younger and a northward leaking expression of the Kerguelen, emanating from the same magma chamber.

CENOZOIC EVOLUTION OF THE INDIAN OCEAN

The chronology of major Cenozoic events in the Indian Ocean has been summarized in Figure 6. The Cenozoic development of this ocean is characterized by the continued rapid northward flight of India and its eventual collision with Eurasia. In the early Paleocene (some 60 m.y. ago) the spreading ridge between Madagascar and India jumped northeast towards India. This initiated the spreading between India and the Seychelles Platform and the formation of the Chagos-Laccadive transform (NORTON and SCLATER, 1979).

The most significant event of the Cenozoic was the collision between India and Eurasia. The timing of this event also marks the acceleration of the motion between Africa and Antarctica (NORTON and SCLATER, op. cit.). This event and its repercussions for the Asian mainland are discussed below.

The Grand Collision

The first encounter of the northward migrating Indian plate with the Asian mainland occurred between magnetic anomaly 23 and 22 time, around 50 m.y. ago in the early Eocene (Fig. 6). This event caused a major reorganization of the geohedron that initiated spreading on the Central Indian Ridge at this time, (NORTON and SCLATER, 1979). The average half-spreading rates changed from about 18-19 cm/yr before to less than 2 cm/yr after the collision, literally stopping the Indian Plate in its tracks (Fig. 5). A period of about 10 m.y. of stasis followed in which little or no motion took place. The spreading pace quickened somewhat between anomalies 17 and 19 (about 40 m.y. ago), but then slowed to about 2 cm/yr after anomaly 11 time (ca 32 Ma). The present convergence rate between India and Asia is calculated at about 1/2 cm/yr.

The ophiolite suite along the Indus Suture Zone and in Tsang Po Valley marks the former subduction zone along which the Indian Plate was being consumed under the Eurasian Plate. Major thrusting in the Himalayas began at the close of Eocene epoch (36-37 Ma), and this time may also mark the more intimate contact between the two plates. The large scale vertical movements on Eurasia, which was previously a stable platform, began at this time and many of the eastern China's basins and grabens have developed since then (MOLNAR and TAPPONNIER, 1975).

The junction of India with Eurasia had serious repercussions for the Asian mainland. Over 1500 km of crustal shortening and deformation of the continental lithosphere has taken place since the beginning of the collision. The deformation zone extends up to 3000 km northeast of the Himalayas. The absence of a long and shallow-dipping fault zone between India and the overlying Tibetan crust precludes underthrusting of India along the Indus Suture Zone at present (MOLNAR and TAPPONNIER, 1975). According to these writers, crustal shortening by underthrusting of the Indian Plate beneath Himalayas and Tibet in the geologic past accounts for only 300 to 700 km. Another 200-300 km can be accounted for by crustal thickening and thrusting in the Pamirs, Tien Shan, Atai and Nan Shan areas. A major fraction of this convergence occurred along major east-west trending strike-slip faults in China and Mangolia and may account for up to a 1000 km of crustal shortening. Such large strike-slip faulting may be a characteristic feature of collisions between continents (MOLNAR and TAPPONNIER, 1975).

Early Cenozoic Oceanography and Climates

During the late Cretaceous and early Cenozoic, the northward migration of India progressively created more open ocean conditions south of the Indian Plate and between other pieces of the ancient East Gondwanaland. The tectonic activity of this period and changing bottom water circulation, however, caused a very discontinuous or patchy stratigraphic record (KIDD and DAVIES, 1978). Pelagic clays continued to be deposited in the deeper parts of the basins and calcareous sediments in the shallower areas. In the Tethyan Seaway thick carbonates still accumulated in Paleocene and most of the Eocene. However, as the seaway became narrower and shallower the nature of carbonates changed accordingly.

First major influx of sediments in the northern Arabian Sea occurred in the middle and late Eocene, after the initial uplift of the Himalayas. First Indus-derived sediments in this area and the major buildup of the Bengal Fan began in the Oligocene (KIDD and DAVIES, 1978). CURRAY and MOORE (1974) suggested that the Bengal Fan was already building up, from a western river source, as far back as the early and late Cretaceous.

The mid Paleocene paleoceanographic patterns (Fig. 7) show that the source of at least some deeper waters was probably still in the mid to low latitudes, although by this time Pacific deep waters may have been of mixed high and low latitudes origin. The Tethys Current dominated the circum-global circulation. In the Indian Ocean south of the drifting Indian Plate the cell was most likely deflected southwestward, hugging the western Australian margin during the return flow eastward.

The position of India in the earlier Eocene did not severely restrict the flow of the Tethys Current north of the Indian Plate. During the generally high seastand times of the early middle Eocene this passage may have remained open to the westward Tethyan flow. In late middle Eocene the general drop in sea level and the further convergence of India and Asia, reduced the passage more severely and the main flow shifted to the west of the Indian Plate (Fig. 8). This is indicated by the continued presence of neritic marine and marginal sediments of middle and late Eocene age in Pakistan (McGOWRAN, 1978) and their general absence further northeast. The Tethys Current became sharply restricted to a narrow southwestern passage by early Oligocene (ca 35 Ma).

Near the Eocene/Oligocene boundary (ca 37-36 Ma) there is widespread evidence of the development or intensification of bottom water activity and a general climatic cooling in all ocean basins. Oxygen-isotopic data from the Indian Ocean (Fig. 9) show generally high temperatures in the late Cretaceous and Paleocene, reaching a peak in the late Paleocene-early Eocene interval. These patterns follow global trends, and this interval may have been the warmest in the Cenozoic (HAQ, 1982). First temperature drop occurred in the middle Eocene, followed by another period of relative stability, until the Eocene/Oligocene transition, when a dramatic drop both in the surface and bottom water temperatures took place. This event manifests itself in the presence of widespread erosional hiatuses in the sedimentary record throughout the Indian and southwestern Pacific Oceans. The scouring of the sediments by cold bottom waters may have caused these hiatuses. This bottom water activity can be ascribed to the partial thermal isolation of Antarctica to higher latitudes and the beginning of large scale freezing at sea level that spurred the initiation of cold Antarctic Bottom Water (AABW). Thus, a drop of about 5°C in the bottom water temperature at the Eocene/Oligocene boundary reflects a fundamental change in the nature of the bottom waters and signals the development of the psychrosphere (BENSON, 1975; SHACKLETON and KENNETT, 1975).

By the middle to late Oligocene time (30-24 Ma) the global surface circulation had begun to evolve the essential present-day patterns (Fig. 10). An event of major significance at this time, that changed the global oceanographic patterns forever, was opening of the Drake Passage in the mid Oligocene, between anomalies 10 and 8 (ca 30 Ma or somewhat earlier). A recent tectonic map of the Scotia Arc region (BRITISH ANTARCTIC SURVEY, 1985) suggests that the deep connection at Drake Passage began at the time of initiation of the spreading in the region, around anomaly 10 time and was well established by anomaly 8 time. The spreading in the area continued until anomaly 5 time. This event led to the enhanced thermal isolation of Antarctica and the initiation of the circum-Antarctic circulation.

The Oligocene was probably the most influential epoch in the history of the world ocean when the oceanographic-climatic features were dramatically changed. The most important change was the shift of circum-global circulation from the tropical-temperate latitudes to the southern high latitudes around Antarctica. This is the time of the fundamental change in the bottom water regime and the entry of the earth from a non-glacial to a predominantly glacial mode.

The Neogene Indian Ocean

The Neogene period (25-1.7 Ma) was characterized by the further accentuation of oceanographic-climatic conditions that had been initiated in the Oligocene. During the early Miocene the convergence of Eurasia and Africa finally interrupted much of the periodic influence of the Indian Ocean on the Mediterranean that had continued intermittently through much of the Oligocene. By this time the major geographic features of the Indian Ocean were also in place and the sedimentary patterns begin to resemble the present-day patterns (Figs. 11 and 12).

During the Neogene, terrigenous sedimentation in the Indian Ocean increased in importance. In the east the Bengal Fan continued to accumulate. The Miocene subduction at the Sunda-Java Trench with its island-arc volcanic activity supplied important volumes of volcanogenic sediments in that region (KIDD and DAVIES, 1978). In the west, the Indus Fan continued to grow. The sediments on this fan seem to have been transported both by the action of turbidity currents as well as the slower-speed suspension currents (WHITMARSH, 1974). In the south, the Zambesi Fan began to develop sometime in the middle Miocene (KIDD and DAVIES, 1978).

A major middle Miocene phase of Himalayan uplift may have initiated upwelling off the Arabian Peninsula by establishment of the cyclic monsoonal system (WHITMARSH, 1974). In other areas, such as the Somali Basin, there was a general increase in the biogenic siliceous input and organic matter in the sediments throughout the Miocene, the result of increased surface productivity of the water masses in the area. Intensification of the western boundary undercurrents supplied with corrosive AABW in the late Oligocene-early Miocene have eroded large surfaces of the Somali, Mascarene, Madagascar and Mozambique Basins (LECLAIRE, 1974).

A drop of the CCD in the late Miocene is signaled by the presence of carbonates in the deeper parts of the basins. The event may have been caused by an accelerated seafloor spreading along the ridges and increased volcanic supply of calcium from the fresh basalt (LECLAIRE, op. cit.). Since late Miocene siliceous sediments have become increasing more common up to the Holocene time due to the development of modern high productivity belts and the subsequent rise of CCD (KIDD and DAVIES, 1978).

By the later part of middle Miocene and late Miocene, both the Red Sea and the Gulf of Aden had begun to open. Earliest synrift evaporite sequences in the Red Sea seem to have had a sabkha-type origin. The overlying evaporitic facies indicates a restricted environment, whereas younger sediments are of more normal marine origin, reflecting the opening of the Red Sea to Indian Ocean influence in late early Pliocene (STOFFERS and ROSS, 1974) (Fig. 12).

The Neogene oxygen-isotopic record of the Indian Ocean is consistent with the global patterns (Fig 9). A period of climatic amelioration in the early middle Miocene was followed by a rather dramatic fall in both the surface and bottom water temperatures around 15 m.y. ago. This was associated with the development of a major enlargement of the East Antarctic ice sheet in the middle Miocene and an intensification of vertical and horizontal thermal gradients (SHACKLETON and KENNETT, 1975; SAVIN, 1977). After this sharp drop, the temperatures stabilized for a while, but they entered a period of greater climatic fluctuations in the late Neogene and Quaternary.

The middle Miocene and younger circulation patterns look like their present day counterparts (Fig. 11). By middle Miocene the eastern limb of the Tethys was closed, but temporary connections may have been established during periods of high seastands, e.g. in the early middle Miocene. The Indo-Mediterranean link was finally severed in the late Miocene (around 12-11 Ma), following a worldwide sea-level drop (RÖGL and STEININGER, 1983). Early Pliocene patterns (Fig. 12) are essentially similar to the modern circulation patterns, with the exception of a open Central American passage, which finally closed around 3 m.y. ago, leading to the present hydrographic patterns of central Atlantic and Pacific regions.

CONCLUDING REMARKS

The preceding discussion is only a very brief overview of the dynamic history of Indian Ocean since its genesis after the fragmentation of Gondwanaland. Aligned along its continental margins are vast old and new sedimentary basins that potentially could contain extensive hydrocarbon and mineral resources. Any meaningful exploitation of these resources requires that the details of the geologic history of these margins must be deciphered. Unfortunately, marine geological/ geophysical research is expensive and requires specialized personnel. Such undertakings are often beyond the reach of individual coastal states, which underscores the need for regional cooperation and sharing of research resources, facilities and personnel. Both national and international scientific agencies need to foster such cooperation and promote research to help fullfil the national and regional aspirations for resource self-sufficiency.

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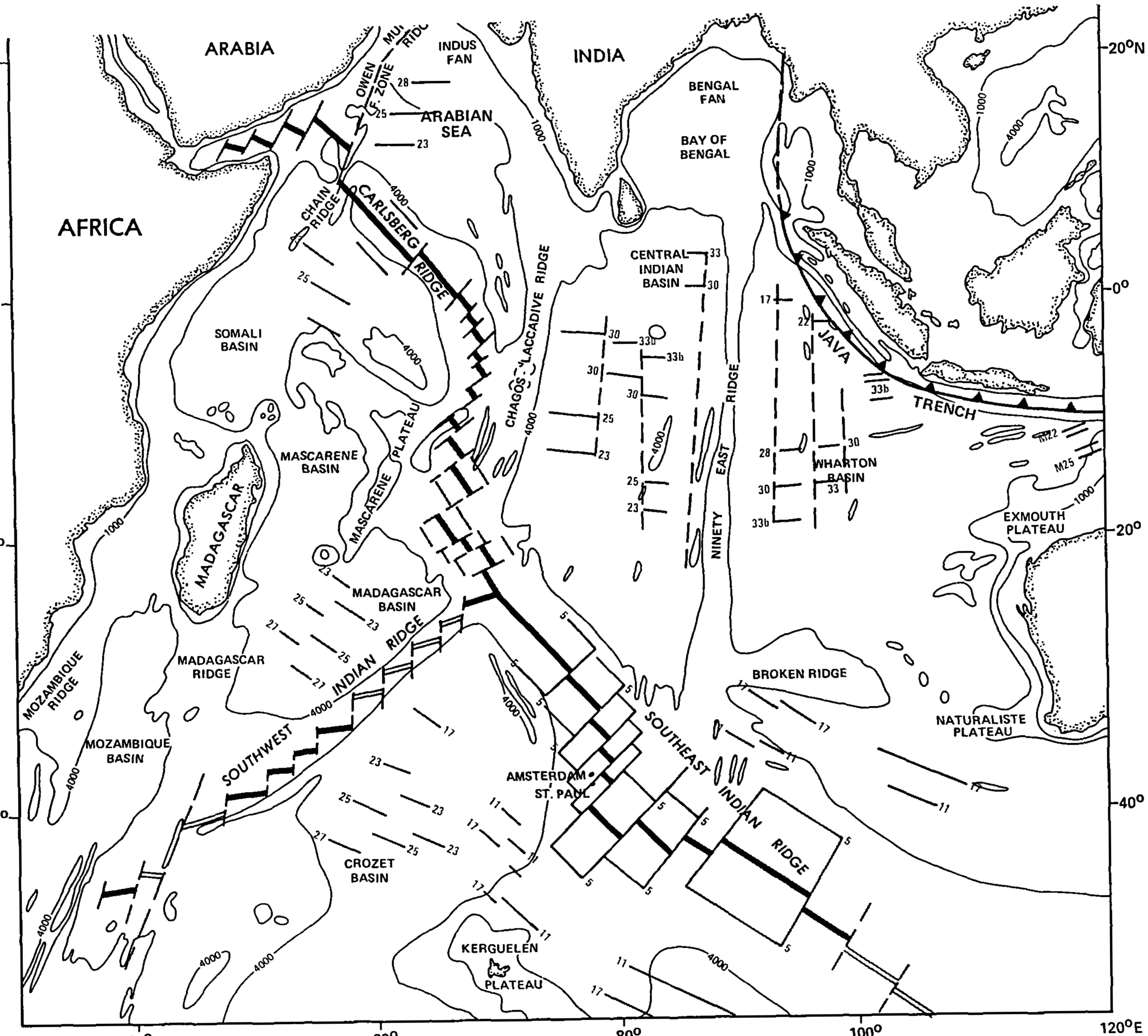

Figure 1. Bathymetric and morphologic map of the Indian Ocean.

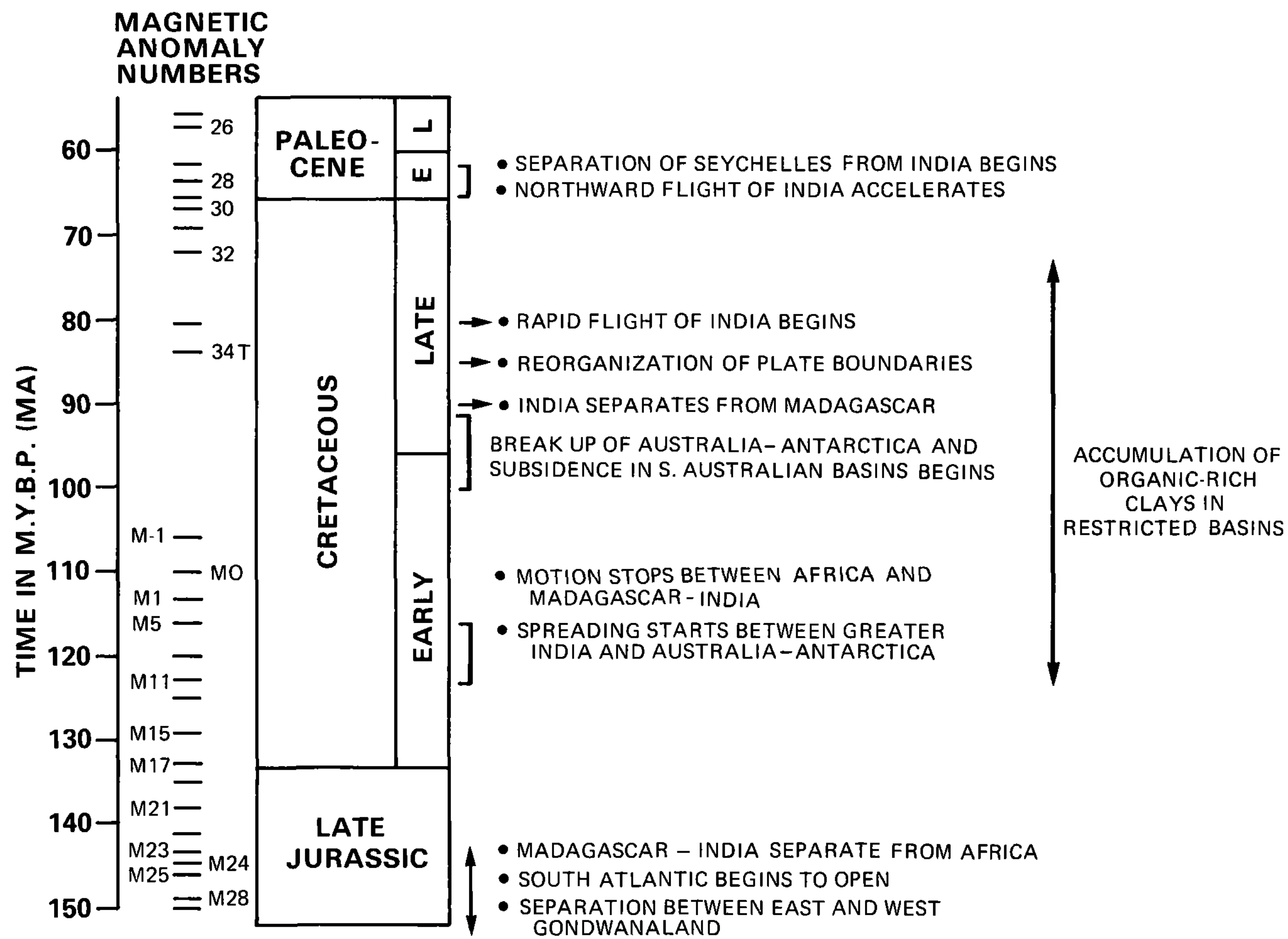

Figure 2. Chronology of major Mesozoic events in the Indian Ocean.

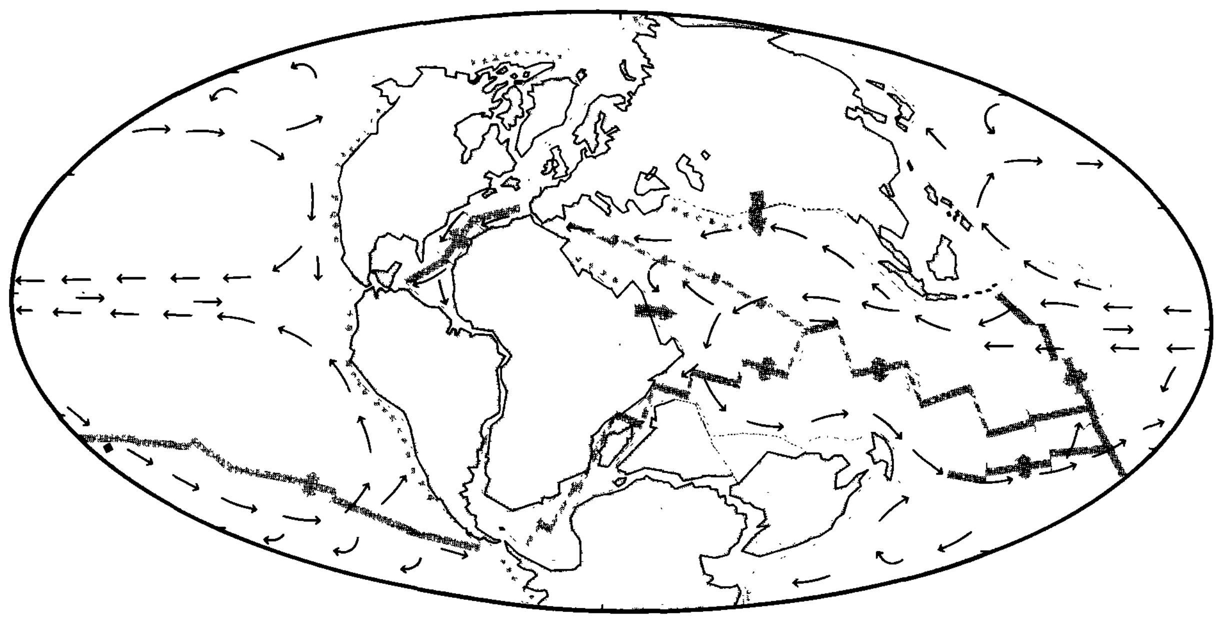

Figure 3. Inferred global paleocirculation scenario for the mid Jurassic time (ca 160 million years ago). Averaged land areas are shown shaded and epeiric sea unshaded (after BARRON et al., 1981). 1000 m depth contour is drawn around the continental margins. Inferred spreading ridges and major strike-slip faults are drawn in solid grey and dashed lines, respectively. Potential source areas of Warm Saline Bottom Water (WSBW) in the low and middle latitudes are indicated with larger arrows, and inferred surface circulation features are shown with smaller arrows. Potential upwelling sites (asterisks) are simulations after PARRISH and CURTIS (1982). The figure shows the circulation before the opening of the gap between North and South America and the beginning of the circum-global Tethys Current. (After HAQ, 1984).

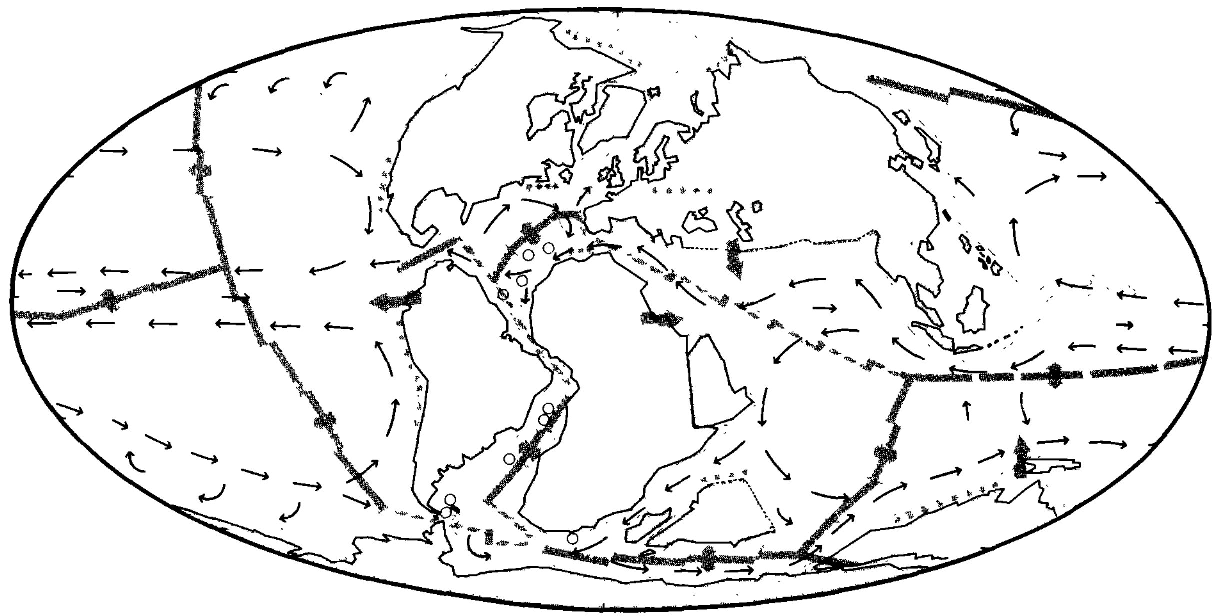

Figure 4. Inferred global circulation patterns for the mid Cretaceous time (ca 100 Ma). For explanation of symbols see caption of Figure 3. Open circles in the Atlantic basins represent deep sea sites where organic-rich black shales of early and middle Cretaceous have been recovered. This figure depicts the oceans after the establishment of the Tethys Current, following the opening of the Central American Isthmus. (After HAQ, 1984). Notice the new spreading ridges in the South Atlantic and change of the locus of ridges in the Pacific Ocean.

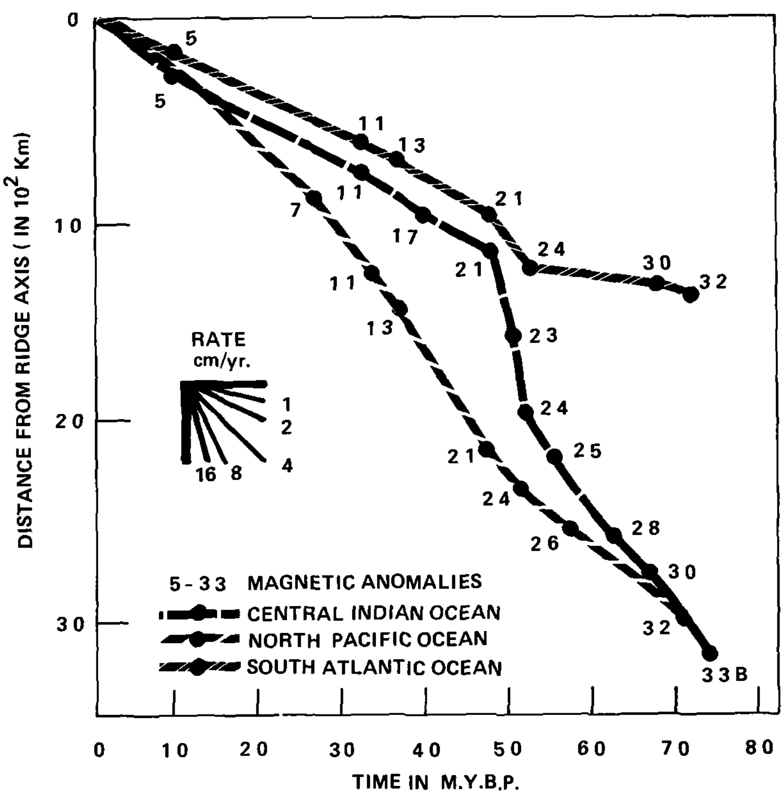

Figure 5. A comparison of half spreading rates for the latest Cretaceous and Cenozoic (magnetic anomalies 33 base to 5) in three major oceans. Notice the exceptionally fast rates in the central Indian Ocean up to anomaly 21. Pacific magnetic anomaly profile after PITTMAN et al. (1968), South Atlantic profile after DICKSON et al. (1968), and central Indian Ocean profile after MCKENZIE and SCLATER (1971).

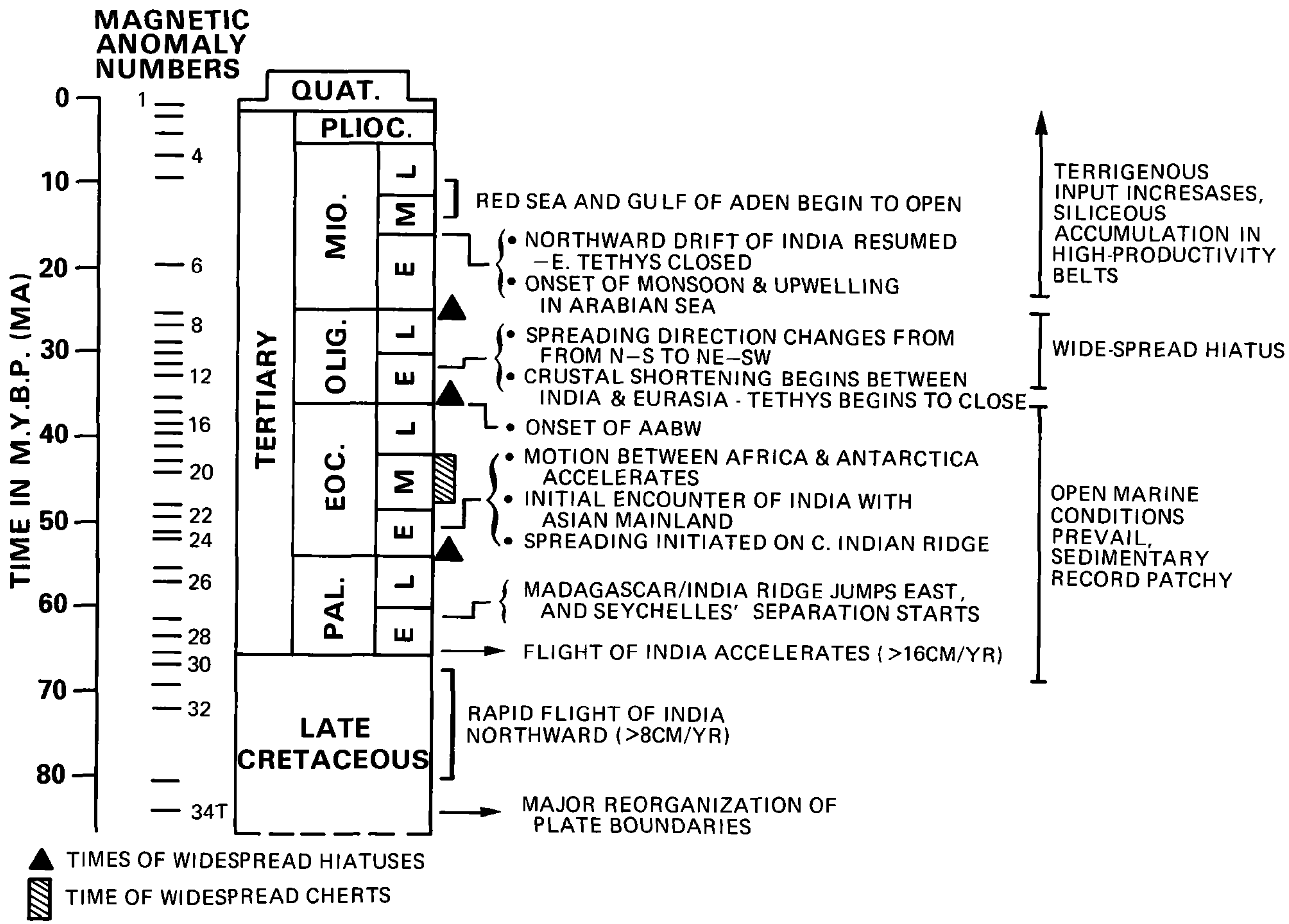

Figure 6. Chronology of major Cenozoic paleogeographic and climatic events and sedimentary patterns in the Indian Ocean.

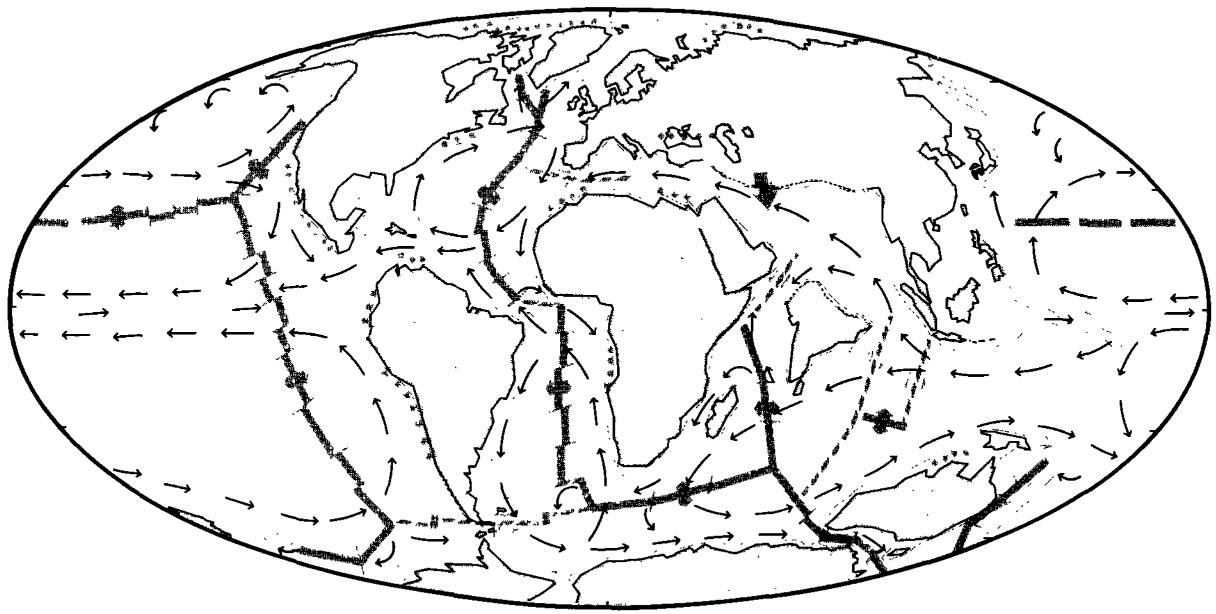

Figure 7. Inferred global circulation scenario for the mid Paleocene (ca 60 Ma). For explanation of symbols see the caption of Figure 3. (After HAQ, 1984).

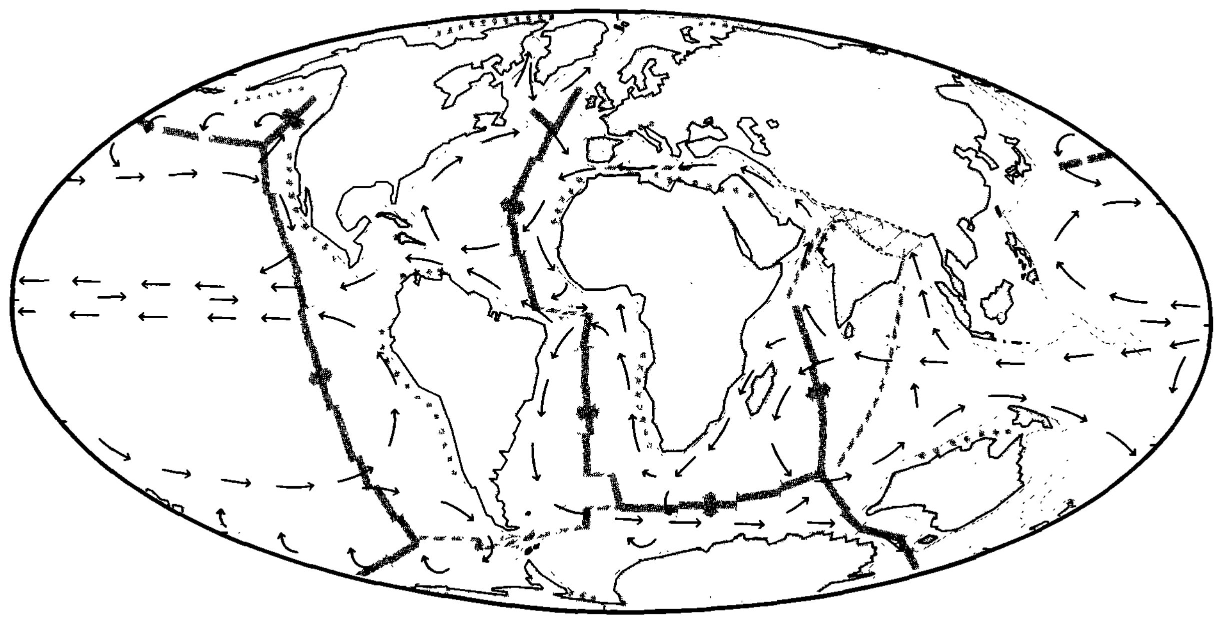

Figure 8. Inferred global circulation patterns in the mid Eocene (ca 45 Ma). For explanation of symbols see caption of Figure 3. India had made its first (soft) contact with the Eurasian Plate and the Tethyian flow to its north may have become somewhat restricted by this time. The main flow shifted towards the west of the Indian Plate during the remainder of the Eocene. (After HAQ, 1984).

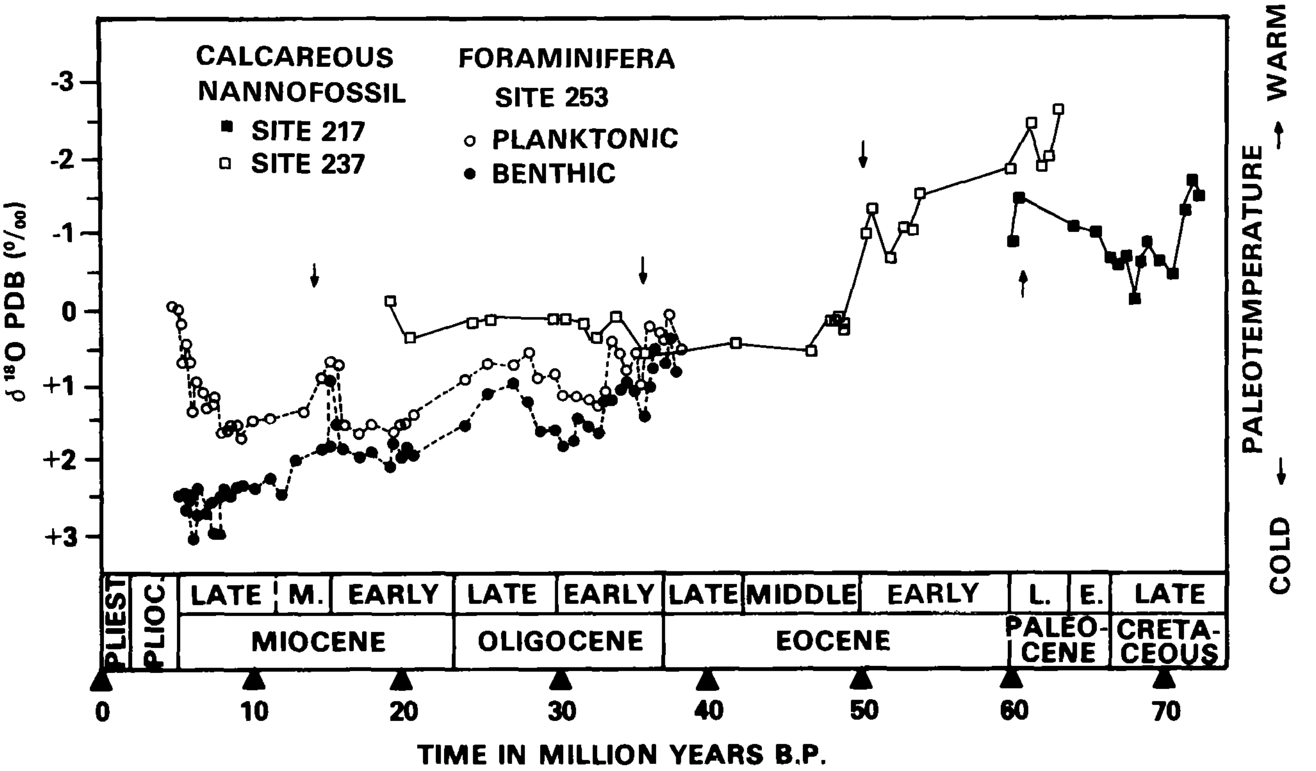

Figure 9. Cenozoic oxygen-isotopic record from DSDP sites in the Indian Ocean. The major climatic trends in the Indian Ocean follow global trends. (After OBERHÄNSLI, 1986).

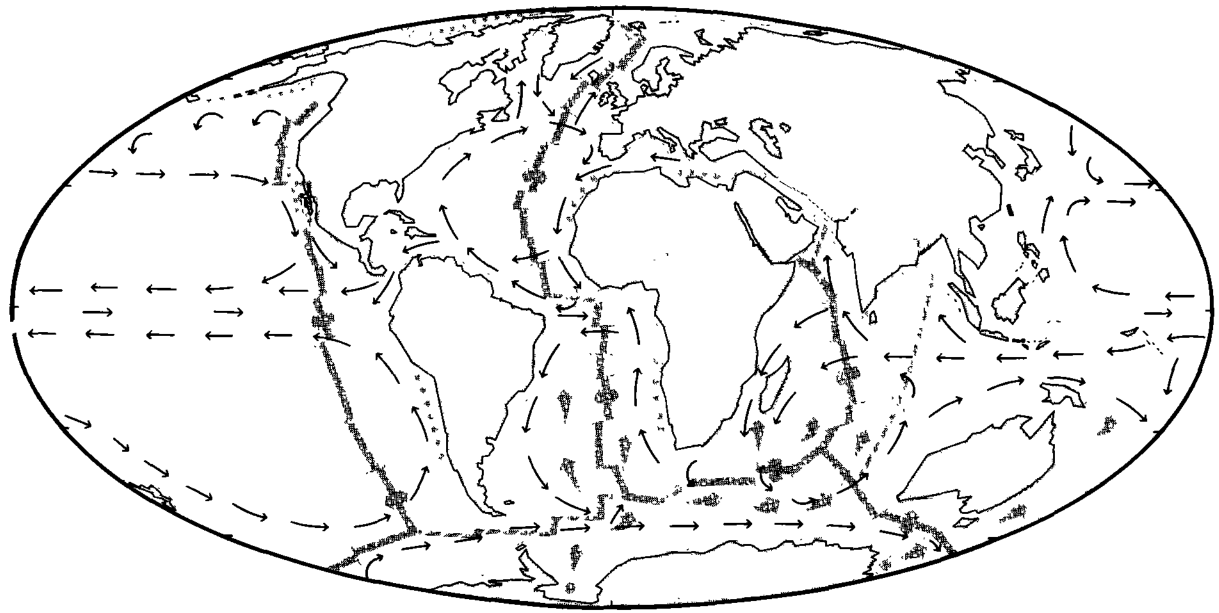

Figure 10. Inferred global circulation trends in the mid Oligocene (ca 30 Ma). For explanation of symbols see caption of Figure 3. By this time the locus of the origin of deep waters has shifted to southern high latitudes. These loci of Antarctic Bottom Water and its initial pathways are shown by grey dots and larger grey arrows. Eastern half of the Tethys Seaway is almost closed by this time and the Drake Passage is now open, establishing the circum-Antarctic Current and shifting the locus of circum-global circulation from low to high latitudes. (After HAQ, 1984).

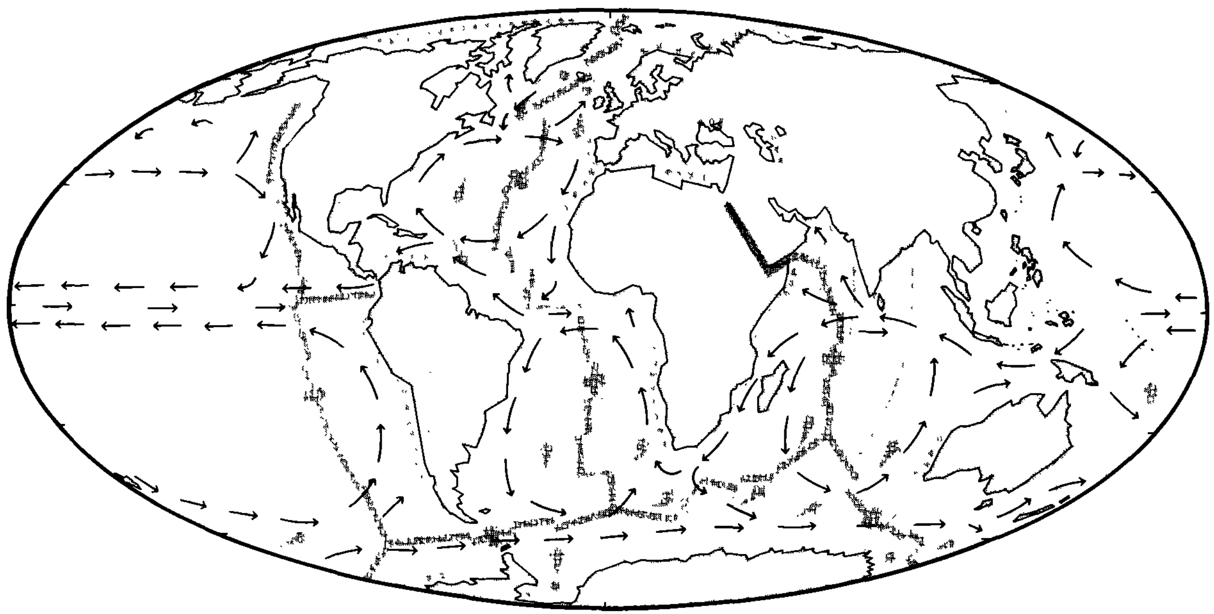

Figure 11. Inferred global circulation patterns for the middle Miocene time (ca 16 Ma). For explanation of symbols see caption of Figures 3 and 10. These patterns look much like their modern counterparts. The locus of the formation of North Atlantic Deep Water (NADW) in the North Sea and the flow of the Mediterranean Intermediate Water (large open arrow at the Gibraltar Straits) was established by this time. (After HAQ, 1984).

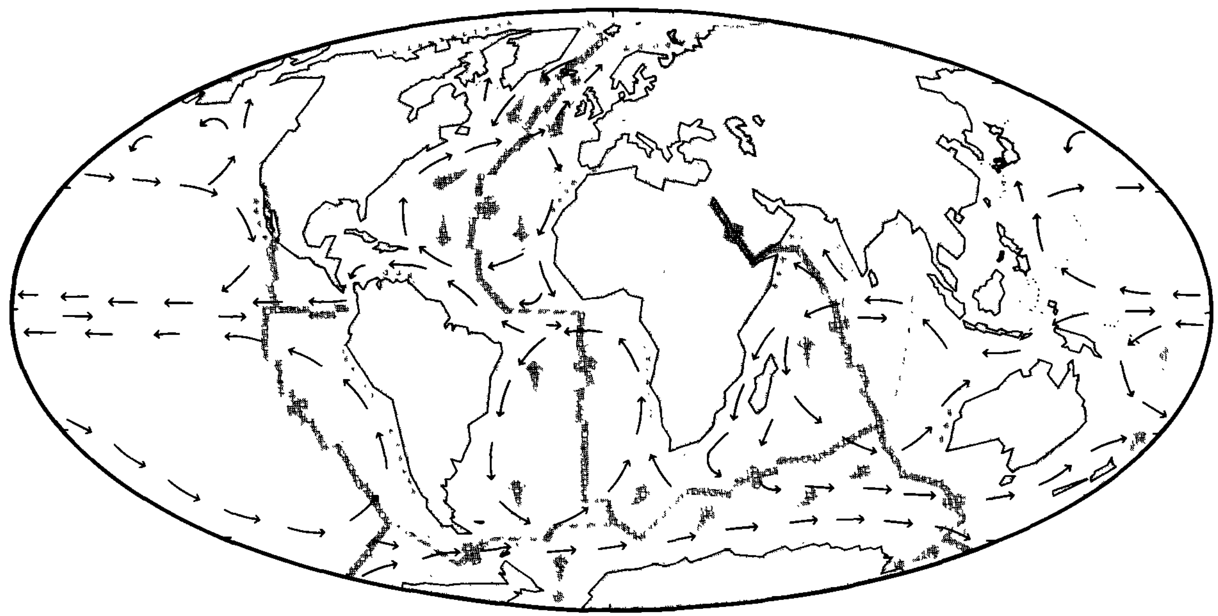

Figure 12. Inferred global circulation for the early Pliocene time (ca 5 Ma). For explanation of symbols see caption of Figures 3 and 11. With the exception of the open isthmus between North and South America (which closed around 3 m.y. ago), the patterns are essentially similar to those of today. (After HAQ, 1984).

GEOLOGICAL-GEOPHYSICAL MAPPING OF THE INDIAN OCEAN

N.N. TURKO, G.B. UDINTSEV AND D.I. ZHIV

Institute of Earth Physics of USSR
Academy of Sciences, P. Grusinakaya 10
113810 Moscow, USSR

HISTORICAL PERSPECTIVE

The history of the geological-geophysical mapping of the Indian Ocean bottom has encompassed several stages. The first stage started in the middle of the 19th century, when the laying of the first telegraph cables necessitated some knowledge of the sea floor. Early scientific expeditions, such as the "Challenger" "Askold", "Gazel", "Valdivia", "Enterprise", "Sealark", "Planet", "Snellius", "Investigator", also resulted in bathymetric soundings and the collecting of bottom sediments. The first bathymetric maps were compiled on the basis of these data (RYKACHEV, 1881; KRÜMMEL, 1881; SCHOTT, 1902; MURRAY and HJORT, 1912; GROLL, 1912). Three editions of the General Bathymetric Chart of the Oceans (GEBCO) also were published, the first in 1903 and the third in 1938-1942. The first gravity measurements were taken in the 1920's-1930's (VENING-MEINESZ, 1948).

The end of the World War II was the starting point of the second stage, which manifested itself by the use of new techniques, particularly the use of both high frequency and low frequency echo-sounding, seismic profiling, magnetics and gravity studies and increasingly accurate navigation. Large expeditions were undertaken by the research vessels "Albatros", "Galathea", "Challenger", "Ob" and "Lena". As in earlier cruises, however, the measurements were made along single profiles, and the resulting maps often merely supplemented the third edition of GEBCO (e.g. ATLAS OF THE WORLD, 1959; STOCKS, 1944). The maps were still very schematic.

INTERNATIONAL INDIAN OCEAN EXPEDITION

An important step in the study of the ocean was the organization of the International Indian Ocean Expedition (IIOE) in 1960-1965, with the encouragement of the Scientific Committee on Oceanic Research (SCOR) and UNESCO. The work of research vessels from more than 20 countries produced a great volume of new data on the topography, geophysics and geology of the ocean, and few major discoveries were made: the Ninetyeast Ridge, the Lanka Ridge, the Chagos Trench, and numerous seamounts. The results of IIOE research were published in the International Geological-Geophysical (IGG) Atlas of the Indian Ocean (1975). Because this enormous task took so much time, other maps based on IIOE results were published and later incorporated into the Atlas. Among them were the physiographical map of the ocean (HEEZEN and THARP, 1964), the Red Sea (LAUGHTON, 1970b) and the Gulf of Aden (LAUGHTON, 1966, 1970a), maps of the Somali and Mascarene basins and of the Karlsberg Ridge (FISHER, 1968; SCLATER and FISHER, 1974), maps of the Sunda Arc (MAROVA, 1966), the Andaman Sea (RODOLFO, 1969) and the Gulf of Oman (SEIBOLD and ULRICH, 1970), and many maps of individual canyons and seamounts. The IGG Atlas also contained the results of international projects which were conducted in the Indian Ocean after the end of IIOE: the Upper Mantle Project (UMP) and the Deep Sea Drilling Project (DSDP).

In discussing the maps in the IGG Atlas of the Indian Ocean, several points should be emphasized: 1) the topographic and of geophysical maps, both ocean-wide and local, are presented at a single scale, thus enabling easy comparison; 2) the tracks plotted on the maps make it possible to distinguish interpretation from real data. In this way the maps also serve to point out areas requiring further research. In the course of future studies it should be sufficient to enter new data on the maps and to reinterpret these parts without re-evaluating the previous data. Data interpretation, however, differs from author to author, as does the manner of cartographic representation. For example, in the bathymetric maps from the Indian Ocean in the IGG Atlas and the 5th edition of GEBCO chart, the boundaries between regions completed by different authors are distinctly visible. The tracks marked on the maps, therefore, make it possible to evaluate a single researcher's interpretation for the whole ocean or a part of it. The maps by KANAEV et al. (1977) and FISHER et al. (1971) are an example of such reinterpretations.

During the closing stages of IIOE and during the Upper Mantle Project, the study of key areas was conducted through a combination of geophysical and geological methods. For example, detailed research was conducted by the Soviet expeditions in the rift zones of the Red Sea (MONIN et al., 1980) and in the zones of faults of the Karlsberg Ridge (ZHIVAGO, 1982). A number of other expeditions also worked in the Red Sea and on the West-Indian Ridge (WHITMARSH et al., 1974; SCLATER et al., 1981).

OTHER MAPPING ACTIVITIES

The so-called aseismic rises are characteristic features of the Indian Ocean that have drawn much attention lately, for example the 54th and 58th cruises of the Research Vessel "Vityaz" on the Ninetyeast Ridge, and studies of other expeditions on the Mascarene, Chagos-Laccadive, Madagascar and Broken Ridges and on Kerguelen Plateau (HOUTZ et al., 1977; FEDOROV and DANILOV, 1978; GOSLIN et al., 1980; UDINTSEV and KORENEVA, 1982; AGAPOVA et al., 1983; BEZROUKOV and NEDPROCHNOV, 1983). These rises provide one of the keys to understanding the tectonics and history of formation of the ocean bottom and the "ocean-continent" transition zone. The study of transition zones also has economic interest, particularly with respect to petroleum. Extensive geophysical research has been carried out off the coasts of Australia (VEEVERS et al., 1974; VEEVERS and COTERILL, 1978) and India (NAINI and TALWANI, 1983), in the Bay of Bengal (CURRAY and MOORE, 1974) and on the Mascarene Ridge (KAMEN-KAYE and MEYERHOFF, 1980).

A certain amount of research has been done in the basins, where special interest has been focused on the fault zones (the so-called intraplate deformations). This kind of work was undertaken, for example, in the Central, Cocos and Oman basins (EITTREM and EWING, 1972; LARSON et al., 1978; BEZROUKOV and NEPROCHNOV, 1983).

Maps showing bathymetry, magnetics and gravity, and geology, as well as isopach maps often are prepared within specific areas and then compiled into larger-scale maps. The 5th edition of the GEBCO chart (scale 1:10 million) is such a generalization of bathymetric data, in which plotting sheets were at the scale 1:1 million. The sheets of the map were issued as soon as they were ready for publication, often resulting in considerable differences in dates of issue. Sheet 5.05, covering the ocean north of the equator, was published under the editorship of A.S. Laughton in April 1975, whereas sheet 5.09, covering the central part (from 0° to 47°S) and edited by R.L. Fisher, M. Yantsch and R. Komer, was published in April 1982. Actually the map of the northern part of the ocean is practically identical to the map in the International Geological-Geophysical Atlas. At present, the 6th edition of GEBCO is being prepared, its principal difference being the presentation of data in digital form and preparation of computer-processed plotting sheets in scale 1:1 million.

Regional magnetic field data have been represented mostly by plots of magnetic anomalies which have been used for different tectonic interpretations (LE PICHON and HEIRTZLER, 1968; McKENZIE and SCLATER, 1971). The gravity field data are published in a series of papers either covering the entire ocean (BOWIN et al., 1982) or smaller areas/basins (SAITO and TAKEI, 1982; WHITE and ROSS, 1979; WHITMARSH, 1979; LUI et al., 1982; KIECKHEFER et al., 1981).

Main features of the bottom topography, gravity and magnetic fields now can be mapped using satellite-observed data (HAXBY et al., 1983; LANGEL et al., 1982; ENGELIS and RAPP, 1984). These new possibilities are very attractive because such satellite missions have provided abundant information with uniform coverage, including remote areas poorly explored by surface ships.

Generalized sediment isopach maps for large areas have not been made, except in the review volume on the results of deep-sea drilling (HEIRTZLER, 1977) and by NEPROCHNOV (1979). Ocean bottom geology data were given in the Geological Atlas of the World (HEEZEN et al., 1978) and in SCHLICH's (1982) monograph.

The latest tectonic generalization is the International Tectonic Map of the World, scale 1:15 million, presented at the International Geological Congress in Moscow in 1984. The structure of the ocean bottom on most tectonic maps is now interpreted from the stand-point of plate tectonics.

FUTURE STUDIES

The detailed research in different regions has revealed the great complexity of bottom structure of the Indian Ocean: for example, great differences in morphology and geophysical character along the ocean ridges, the block structure of the so-called aseismic rises, different geological-geophysical characteristics of different blocks, the zones of intraplate deformations, etc. We believe that these data indicate the heterogenity of the ocean bottom and that they cannot be interpreted from the standpoint of plate tectonics alone. Further research using new techniques, such as multibeam echosounding, submersibles, etc., and accumulation of new data can bring us closer to the solution of many complex problems for which there are as yet no unambiguous answers.

The next stage in the geological study and mapping of the ocean bottom should be carried out on geotraverses, that is, coast-to-coast strips of detailed research crossing different structures. Trans-Atlantic Geotraverses (TAG) in the North Atlantic was the first of such studies (RONA, 1980). At present similar studies are being carried out between Angola and Brasil in the band about 2° wide (from 11° to 13°S). Similar traverses in the Indian Ocean will help to reveal the nature of the transition and conjugation of different types of structures to allow for construction of three- dimensional models of ocean bottom structure.

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AN EXAMINATION OF THE FACTORS THAT INFLUENCE THE MONTHLY-MEAN SEA LEVEL ALONG THE COAST OF INDIA

S.R. SHETYE AND A.M. ALMEIDA

Physical Oceanography Division
National Institute of Oceanography
Dona Paula, Goa 403 004, India

ABSTRACT

Available data on monthly-mean sea levels at eight locations off India — Veraval, Bombay, Marmagao and Cochin on the west coast, and Nagappattinam, Madras, Vishakhapatnam and Calcutta on the east coast — have been used to determine the annual cycle of the sea level. Three other sets of data — atmospheric pressure, ship-drift estimates and rainfall — have been utilized to examine the factors that control the observed variations in the sea level.

In general, the effect of the atmospheric pressure variation on the monthly-mean sea level along the coast is significant; the amplitude of the effect is dependent on location and varies between 4-20 cm. The sea level at Calcutta shows no influence of large scale coastal circulation. Off Bombay the monthly-mean coastal currents are weak and do not show significant correlation with the sea level. Sea levels at Veraval, Marmagao, Cochin, Nagappattinam, Madras and Vishakhapatnam (corrected for atmospheric pressure effects) show good correlation with the longshore component of the coastal current. The sea level records at these locations could be a good tool for the long-term monitoring of the surface geostrophic currents along the coast.

INTRODUCTION

Studies during the last few decades have shown that monthly-mean sea level can be used to monitor the large scale, low frequency circulation of the oceans (e.g., WYRTKI, 1979). There is now a network of tide gauges in the Pacific Ocean to monitor sea surface topography, and similar networks have been proposed for the other oceans (see WYRTKI and PUGH, 1983). The North Indian Ocean component of such a network will have to rely on the data collected from tidal stations located within the territory of India. Some of the existing stations may be used towards this purpose.

The existing Indian tidal stations, however, were not set up to monitor the general ocean circulation, but rather to measure the tides in areas of interest to maritime activities. Hence, not all of the existing stations may be suitable for sea level monitoring. It is therefore necessary to identify the factors that control the monthly-mean sea level at the existing stations. Such an exercise would help to determine how the data already collected might be used to understand ocean circulation in the North Indian Ocean. Furthermore, it might become possible to select tidal stations that would be suitable for inclusion in a sea-level network.

With the above considerations in mind, we have examined available data on monthly-mean sea levels, together with other relevant information, from eight locations along the coast of India (Fig. 1). Four of these, Veraval, Bombay, Marmagao and Cochin, lie on the west coast. The others, Nagappattinam, Madras, Vishakhapatnam and Calcutta, are on the east coast. The next section is devoted to a preliminary discussion on the likely causes behind the monthly-mean sea level variations in coastal areas.

PRELIMINARY DISCUSSION

PATULLO, MUNK, REVELLE and STRONG (1955; hereafter referred to as PMRS) related the annual cycle for monthly-mean sea levels to effects from: (1) astronomic tides; (2) atmospheric pressure; and (3) "steric" fluctuations. Here it suffices to point out that the upper limit of the tidal contribution is around a centimeter, which is much less than the amplitude of the annual cycles we shall be dealing with. The contribution to the sea level change due to atmospheric pressure changes depends on the difference between the local pressure and the mean global sea surface pressure. The annual cycle of this contribution has been computed in the next section.

The most important finding of PMRS was that variations in the recorded monthly-mean sea level at a location generally agree closely with the "steric" departures (Z_α) in the vicinity of the location.

Here, $$Z\alpha = \frac{1}{g}\int_{P_a}^{P_o} \Delta\alpha \, dP \qquad (1)$$

and, $$\Delta\alpha = \frac{1}{\rho(T,S,P)} - \frac{1}{\rho(\bar{T},\bar{S},P)}$$

ρ is the density, T and S are the temperature and the salinity, $\bar{T}$ and $\bar{S}$ being their annual mean, g is the acceleration due to gravity, P_a is the atmospheric pressure, and P_o is the pressure at a depth where all seasonal effects are assumed to vanish. PMRS did not examine the implications of the above results to circulation in the vicinity of the tide gauge, even though the mass field (which determines Z_α), the sea surface topography, and the geostrophic velocity field are interrelated.

Consider the situation depicted in Figure 2. A coastal current flows southward along a north-south coastline. We assume that the motion is restricted up to a distance R from the coast, R being of the order of a Rossby radius of deformation (approximately 100 km) from the coast. Assuming that the pressure gradient normal to the coast is in geostrophic equilibrium with the velocity field, we get

$$f v_1^s = g \frac{\partial Z_s}{\partial x} \qquad (2)$$

where f and g are the Coriolis parameter and the acceleration due to gravity, respectively. v_1^s is the longshore component of the surface geostrophic velocity, and $Z_s(x,y)$ is the topography of the ocean surface. Variations in Z_s near a coast can be determined from sea level data. We can write Z_s in terms of variations in dynamic height normal to the coast,

$$f v_1^s = \frac{\partial}{\partial x}\left[\int_{P_o}^{P_a} \delta dP\right], \qquad (3)$$

where

$$\delta = \frac{1}{\rho(T,S,P)} - \frac{1}{\rho(0,35,P)}$$

is the specific volume anomaly. P_o is the pressure at the depth of no motion. The term in the square brackets is proportional to Z_α. The observation made by PMRS, that variations in Z_α closely match sea level changes, when viewed in isolation bears no particular implication to the dynamics of the currents in the vicinity of the tide-gauge. A static ocean, heated or cooled uniformly at the surface, will show variations in Z_α which match with the sea level variations. But when we look at the result from the point of view of the dynamics behind Equations (2) and (3), and impose the restriction that the changes in the mass field are mainly due to advection, the PMRS result becomes a tool to monitor coastal geostrophic currents. To see this we approximate Equation (2),

$$f v_1^s \approx \frac{g}{R}\left[Z_c - Z_o\right] \quad , \qquad (4)$$

where Z_c is the sea level at the coast (point A in Figure 2), Z_o is the sea level at point B in Figure 2, at a distance R which is the offshore boundary of the coastal current. The variations in Z_α would be small in comparison to those in Z_c because point B is located in a regime which is quiescent in comparison to that A. Under these conditions variations in Z_α, Z_c, and v_1^s will match.

Our main concern in the present study is to see how well Equation (4) holds at the eight locations shown in Figure 1. One of the factors, other than tides and atmospheric pressure, which may introduce noise in the above relationship is the contribution to sea level from the accumulation of runoff due to rainfall in the vicinity of a tide gauge. This may be of special concern during the southwest monsoon season for gauges located at or near the mouth of a river. In the next section we assemble data to address these considerations.

DATA

The Permanent Service of Mean Sea Level (PSMSL) stores monthly and annual mean heights of sea level after collecting them from different national organizations (PSMSL, 1978). The monthly data used here have been supplied to the PSMSL by the Survey of India, Dehra Dun, India. These monthly mean values have been computed by averaging hourly values of a tidal record for a month (Table 1). We computed the normal monthly value for a month, say January, at a location by averaging the monthly values at that location for all Januaries of the data record. By averaging these normal monthly values we computed the annual mean at that location. The dashed curve in the top panel of Figures 3 to 10 gives the normal monthly values with the annual mean subtracted. The record of monthly mean sea level values shows considerable interannual variability, the extent of which can be seen from the standard deviations shown in Table 1. It should be noted that though the interannual variability may be a source of noise for the normal annual pattern considered here, this variability contains useful information on how the oceanic conditions vary from year to year.

To compute the contribution of the atmospheric pressure variations to the changes in sea level we need the global and local monthly-mean atmospheric surface pressure. The data on surface pressure for Bombay, Marmagao, Vishakhapatnam and Calcutta have been taken from RAO (1981). The data for Veraval and Cochin have been taken from WEST COAST OF INDIA PILOT (1975), and those for Nagappattinam are taken from BAY OF BENGAL PILOT (1978). The global monthly mean surface atmospheric pressure has been given by PMRS, who have also defined a procedure for correcting the sea level for atmospheric pressure changes. The same procedure has been followed here. The corrected sea level is given by the solid curve in the top panel of Figures 3 to 10.

KNMI Atlas (1952) gives the monthly-mean ship drift estimates in the Indian Ocean on a 2° x 2° grid. We have used these data to estimate the longshore component of the surface current (v_l^s). It is noted that though the use of ship-drift estimates as an indicator of surface currents has its own problems, no other source for estimating normal monthly surface currents along the coast of India is presently available. The longshore component has been determined by resolving the monthly-mean drift along a straight line tangential to the coast. The tangents and the 2° -squares for which the ship-drift data have been used are shown in Figure 1. The computed longshore components are shown in Figures 3 to 10. The following convention has been followed to determine the sign of these components. Along the west coast of India a component is taken as positive if the flow is northward. Along the east coast a component is taken as positive if the flow is southward. This choice was made to ensure that the sign of the sea level change and that of change in v_l^s would be the same if geostrophic balance as given in Equation (4) holds.

RESULTS AND DISCUSSION

The amplitude of the annual variation of the monthly-mean sea level at Veraval, Bombay, Marmagao and Cochin is around 20 cm. It is over 30 cm at Nagappattinam, Madras and Vishakhapatnam, and well over a meter at Calcutta. The magnitude of the correction due to atmospheric pressure is the highest, almost 10 cm, at Calcutta. It decreases towards the south. At Vishakhapatnam, Madras, Veraval and Bombay the maximum correction is around 7 cm. At Cochin it is less than 2 cm.

The circulation off the coast Nagappattinam, Madras and Vishakhapatnam is more energetic than that at the other locations. The longshore component off this coast is towards the north between February and August, the maximum magnitude of the component being around 40 cm s^{-1}. During September-January the longshore component is towards the south, its maximum magnitude being approximately 40 cm s^{-1} in November. At Cochin, Marmagao and Veraval the current is southward from March to October with a peak value of about 30 cm s^{-1} around July-August. During November-February the current is weak, less than 10 cm s^{-1}, and set towards the north. The monthly-mean circulation off Bombay remains weak with velocities less than 10 cm s^{-1} throughout the year, except during the month of July when a northward flow with magnitude of around 30 cm s^{-1} is noticed. The annual cycle of the longshore component off Calcutta, though similar to that at Vishakhapatnam, has an amplitude half of that at the latter.

Figure 3 to 10 reveal that at Veraval, Marmagao, Cochin, Nagappattinam, Madras and Vishakhapatnam the longshore current closely follows the pressure corrected sea level. Notice especially that the reversals in the currents are reflected in the sea level changes. The data that we have used here, particularly for currents, are not suitable for carrying out statistical significant tests. However, it is worthwhile to point out that the correlation coefficients between the atmospheric pressure-corrected sea level and the longshore current at Veraval, Marmagao, Cochin, Nagappattinam, Madras and Vishakhapatnam are 0.8, 0.7, 0.9, 0.9, 0.9 and 0.9 respectively. A scatter diagram of the pressure corrected sea level against the longshore current at the above six locations is given in Figure 11; the correlation coefficient for the 72 points plotted in the figure is 0.84. This value is high enough to conclude that the variation in the monthly-mean sea level at these locations is a good indicator of the longshore component of the coastal current. A least square linear fit through these points has a slope ($\Delta Z_c/\Delta v_l^s$) of 0.4 sec. We see that this figure is consistent with Equation (2) with g, f and R equal to 10^3 cm s^{-2}, 0.5 x $10^{-4}s^{-1}$, and 100 km respectively. This implies that a good physical basis exists to expect the high value of correlation seen above.

The sea level at Calcutta and Bombay does not fit in the above scheme. The Calcutta tide gauge is located on the River Hoogly about 60 km upstream from its mouth. It is likely that the dominant signal recorded by the gauge arises from the water level in the river. The coastal circulation off Bombay is weak. No clear signal arising from it is noticeable in the sea level data. Better quality data, particularly on the current distribution, are needed to examine the relationship between circulation and sea level.

The sea level at Veraval, Marmagao and Cochin decreases as the rainfall increases during the southwest monsoon (Figs. 3-6). This indicates that the contribution to the sea level change due to accumulation of rain runoff at these sites is small in comparison to that from large scale coastal circulation. At Nagappattinam, Madras and Vishakhapatnam the variation in sea level follows a pattern similar to that of the rainfall variation, suggesting that the observed monthly-mean sea level variation during these months is, at least partially, a consequence of rain runoff. Though such a possibility cannot be unequivocally ruled out with the present data, a more interesting scenario, consistent with the conclusion that the monthly-mean sea level is predominantly related to the longshore current, does exist. Two of the forcing functions for the current of the east coast are the local wind stress and the curl of the wind stress over the Bay of Bengal. The longshore component of the wind stress is poleward from March to October, with a peak value of around 0.5 dyne cm^{-2} in June-July; from November to February the component is equatorward with a peak of 0.2 dyne cm^{-2} (SHETYE et al., 1985). The Sverdrup transport computed by using the field of curl of wind stress over the Bay of Bengal suggests a poleward coastal flow in January, and an equatorward flow in July. In addition to the above two, another forcing function may be operative. GOPALA KRISHNA and SASTRY (1985) have shown that during the southwest monsoon season a longshore density gradient arises along the east coast of India. The main cause of this gradient is the dilution produced by the runoff from the rivers which are fed by the rains over the Indian subcontinent. Coastal currents driven by longshore density gradient are believed to exist along the west coast of Australia (McCREARY et al., 1986) and along the west coast of India (SHETYE, 1984). The observed close relationship between the sea level and the coastal flow at Nagappattinam, Madras and Vishakhapatnam encourages us to suggest that the similarities between the rise in sea level and the increase in runoff during June-November are not due to accumulation of runoff. If the rainfall along the east coast has any contribution, it is probably because the runoff helps to set up a longshore density gradient which then enhances a southward flow, which in turn affects the sea level.

In summary, with regard to the factors that control the monthly-mean sea level at the eight locations studied here, we note the following:

(1) In general, the effect of atmospheric pressure variations on the monthly-mean sea level along the coast is significant. The amplitude of the effect, which varies between 4-20 cm, is dependent on location and decrease towards the south.

(2) At Calcutta the gauge data show no influence of the large scale coastal circulation. It appears that the gauge here is predominantly under the influence of the level of the River Hoogly. The monthly-mean coastal circulation off Bombay is weak. No significant correlation between the longshore component of the current and the monthly-mean sea level is noticed.

(3) The sea level corrected for atmospheric pressure effects at Veraval, Marmagao, Cochin, Nagappattinam, Madras and Vishakhapatnam shows a good correlation with the longshore component of the coastal current. The sea level records at these stations would be a good tool for the long term monitoring of the surface geostrophic currents along the coast.

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Table 1.

(1)	(2)	(3)	(4)
Veraval	1959-64	17	25
Bombay	1957-63	3	23
Marmagao	1969-78	8	14
Cochin	1958-78	13	11
Nagappattinam	1971-77	15	21
Madras	1957-77	28	15
Vishakhapatnam	1957-78	28	17
Calcutta	1957-63	0	46

(1): Location

(2): Years covered in the data

(3): Number of months missing in the record

(4): Mean standard deviation (mm) of the normal monthly-mean value (standard deviation of the normal = $\sqrt{\frac{n \sum_i x_i^2 - (\sum_i x_i)^2}{n^2 (n-1)}}$ where x_i's are monthly-mean values for a given month, and **n** is the number of values used to compute the normal).

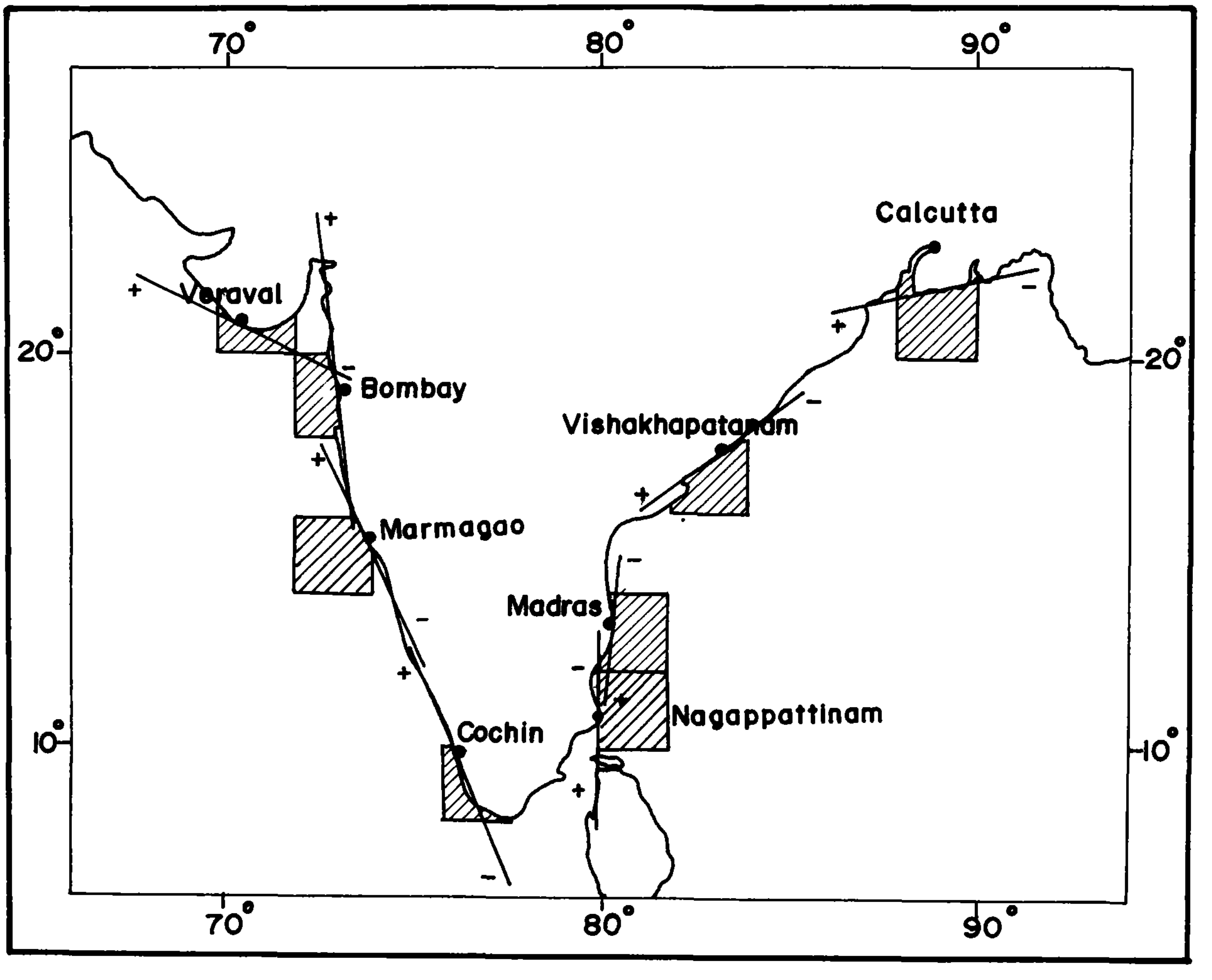

Figure 1. Area of interest, showing the locations of the eight stations where monthly mean sea levels have been documented. The hatched two-degree squares show the areas from where ship-drift values have been used to estimate coastal currents. The tangent to the coast used to compute the longshore component is shown at each location. The convention used to define the sign of the component is also shown.

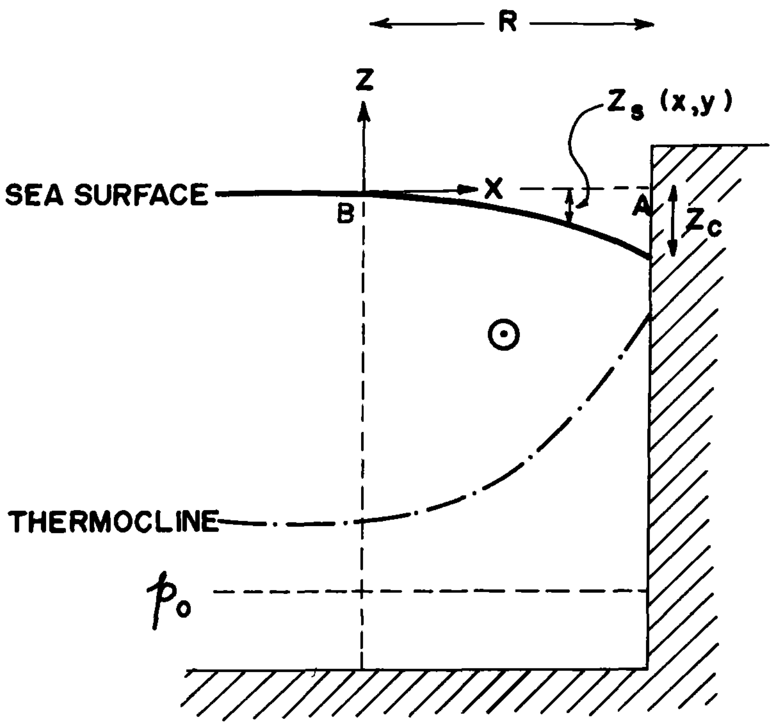

Figure 2. An idealized coastal current. The coastline stretches along the north-south direction. The current is southward, and the motion extends up to a distance R from the coast. $Z_s(x,y)$ is the sea surface. X, Y and Z are the eastward, northward and upward axes respectively. The sea surface tilts down by Z_c at the coast. The thermocline tilts upward towards the coast. P_o is the pressure at the level of no motion.

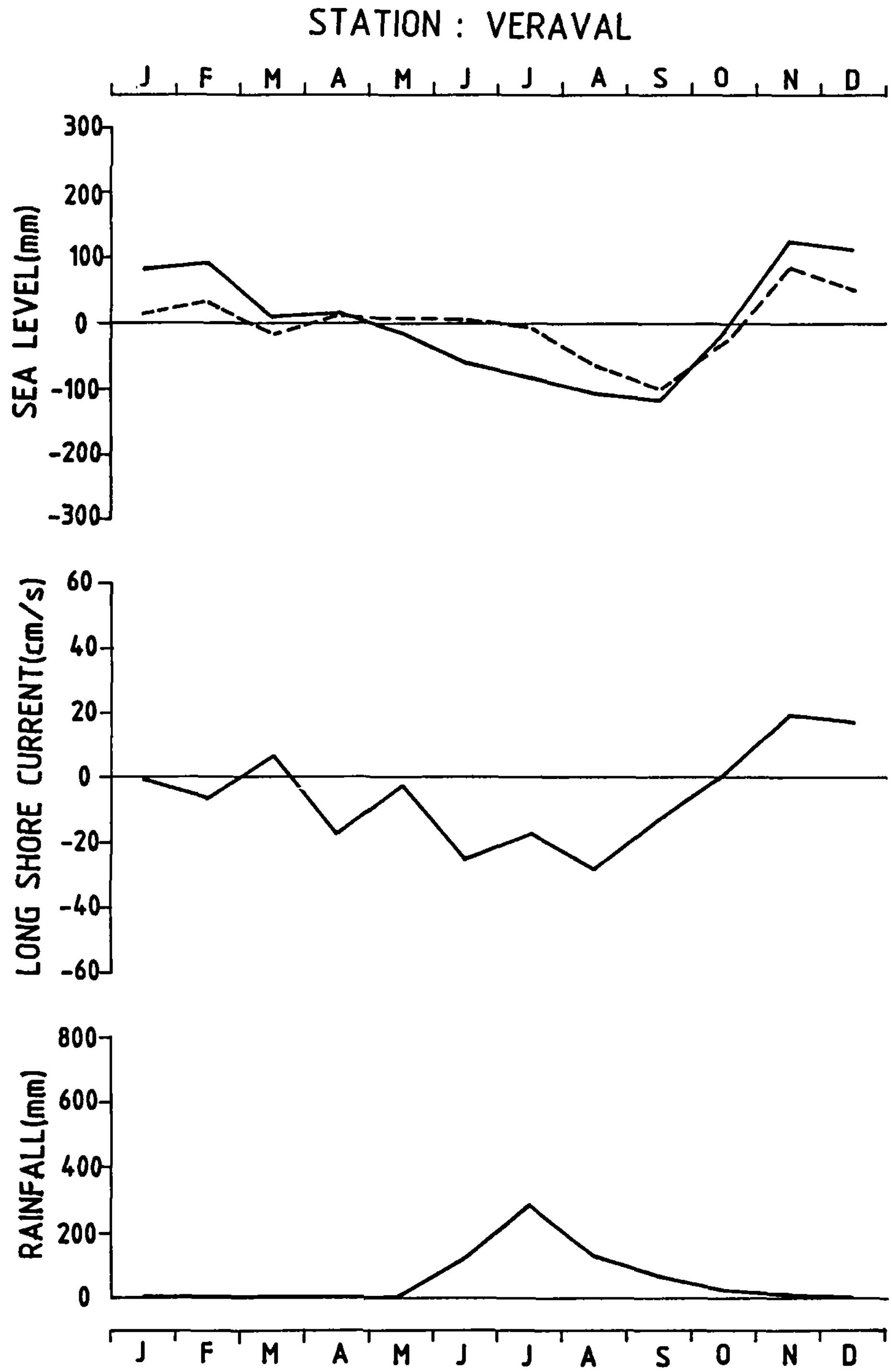

Figure 3. Normal monthly data for Veraval. The dashed line in the top panel gives the recorded monthly-mean sea level (mm). The sea level corrected for variations in atmospheric pressure is shown by the solid line in the same panel. The second panel shows the longshore component of the coastal current (cm sec^{-1}) based on the ship-drift values. The third panel gives the monthly mean rainfall in mm. The horizontal axis defines the month. Sources of data are given in the text.

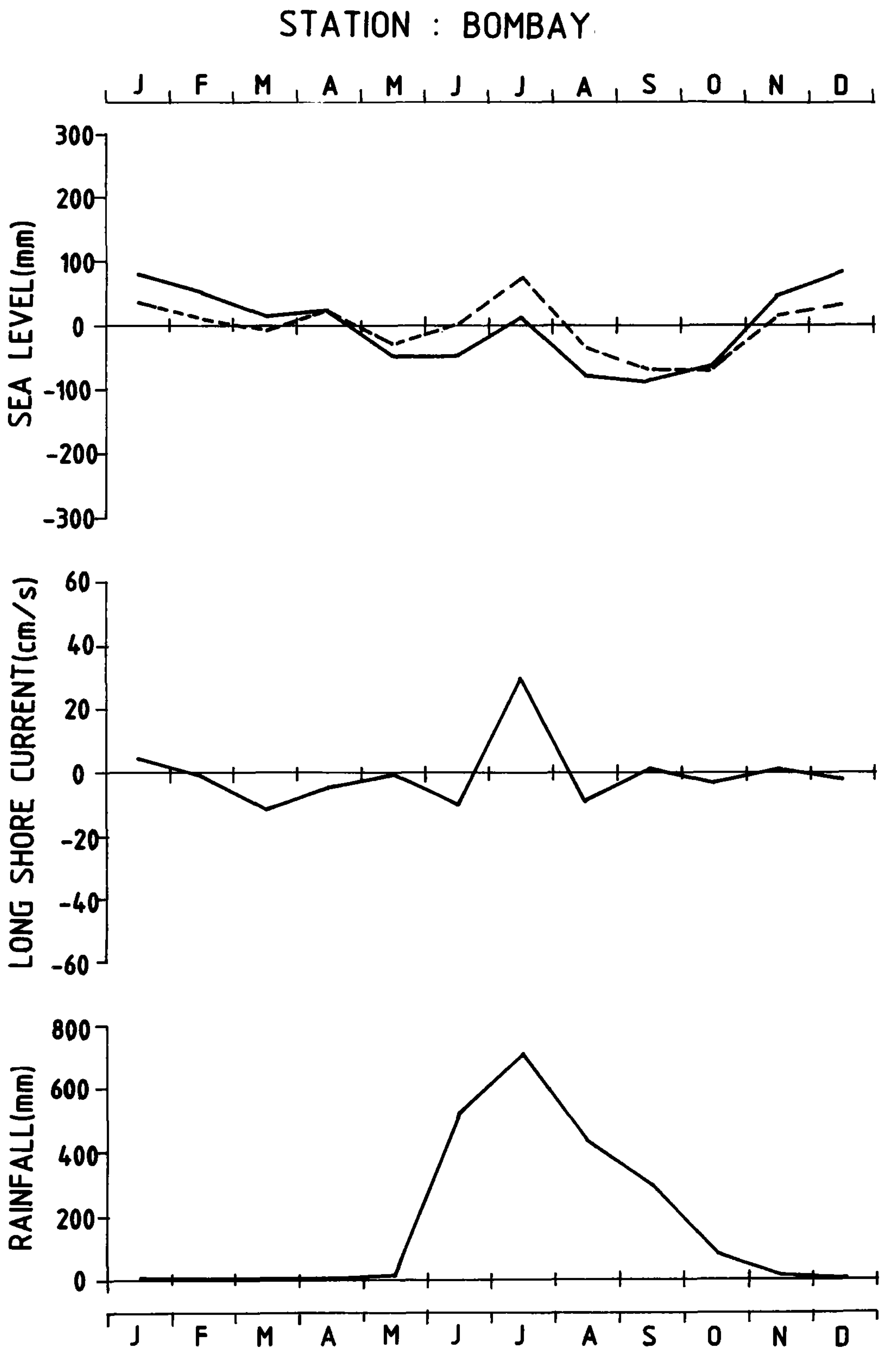

Figure 4. Same as in 3, but for Bombay (Apollo Bandar).

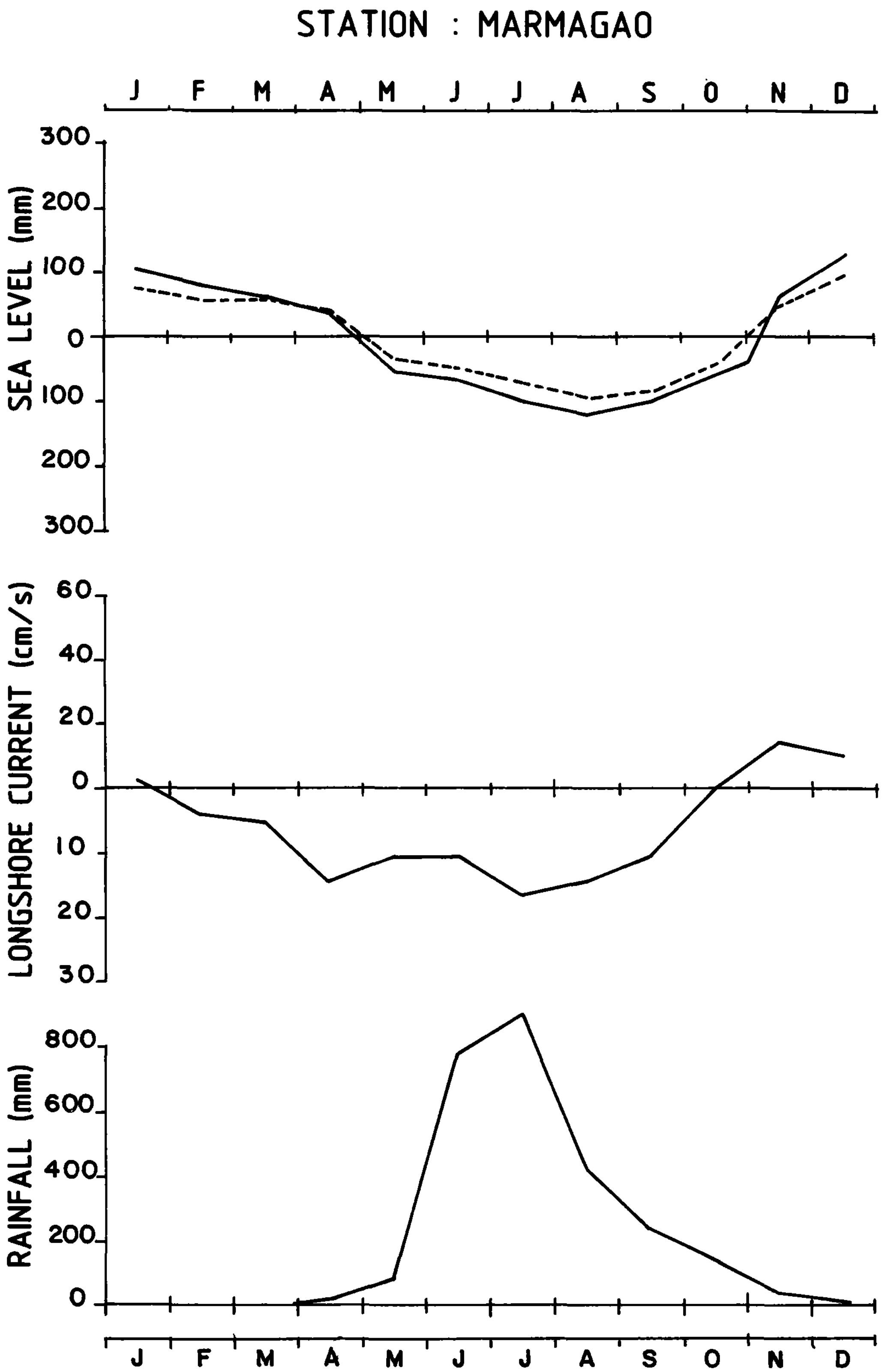

Figure 5. Same as in 3, but for Marmagao.

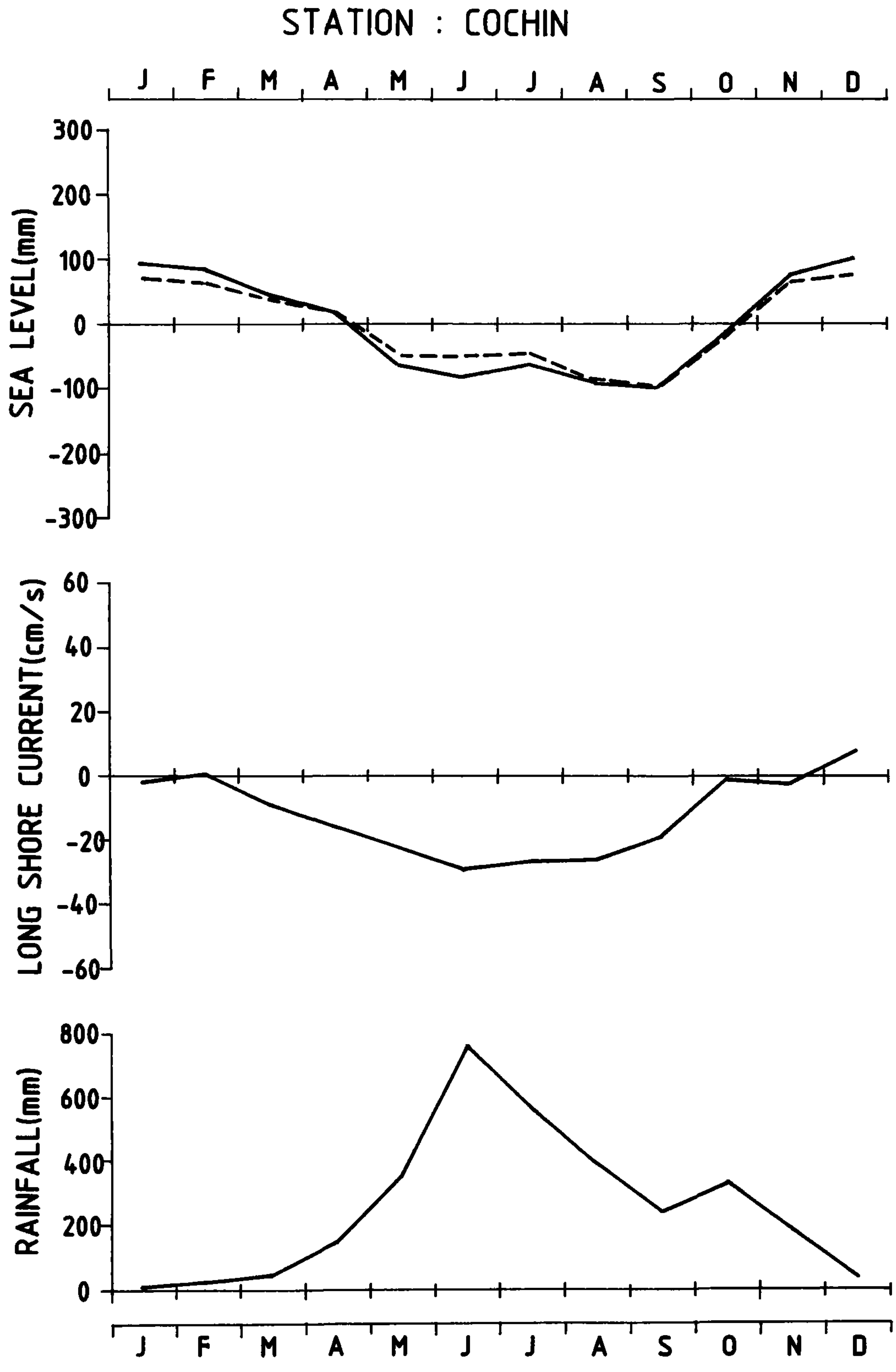

Figure 6. Same as in 3, but for Cochin.

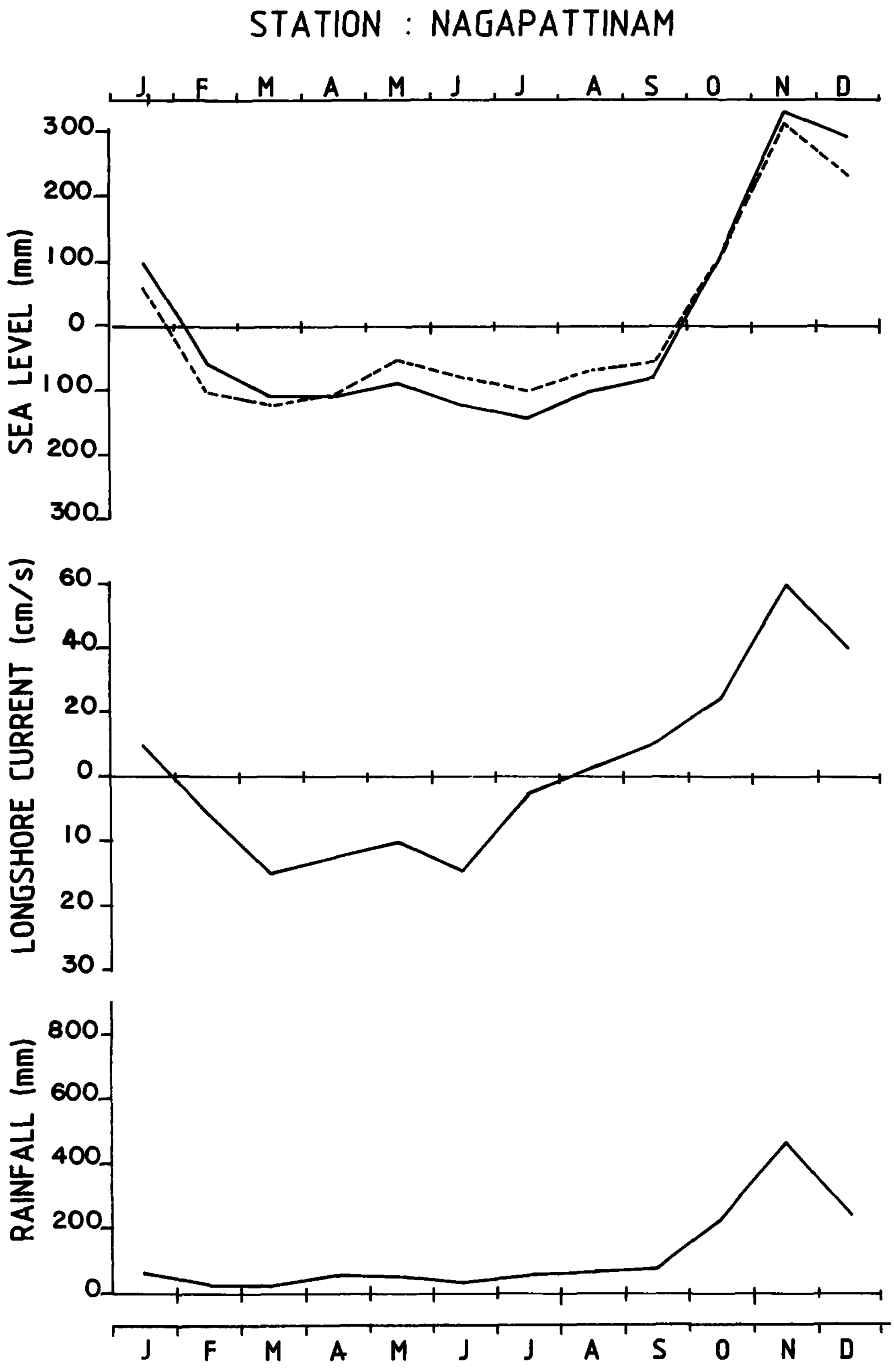

Figure 7. Same as in 3, but for Nagappattinam.

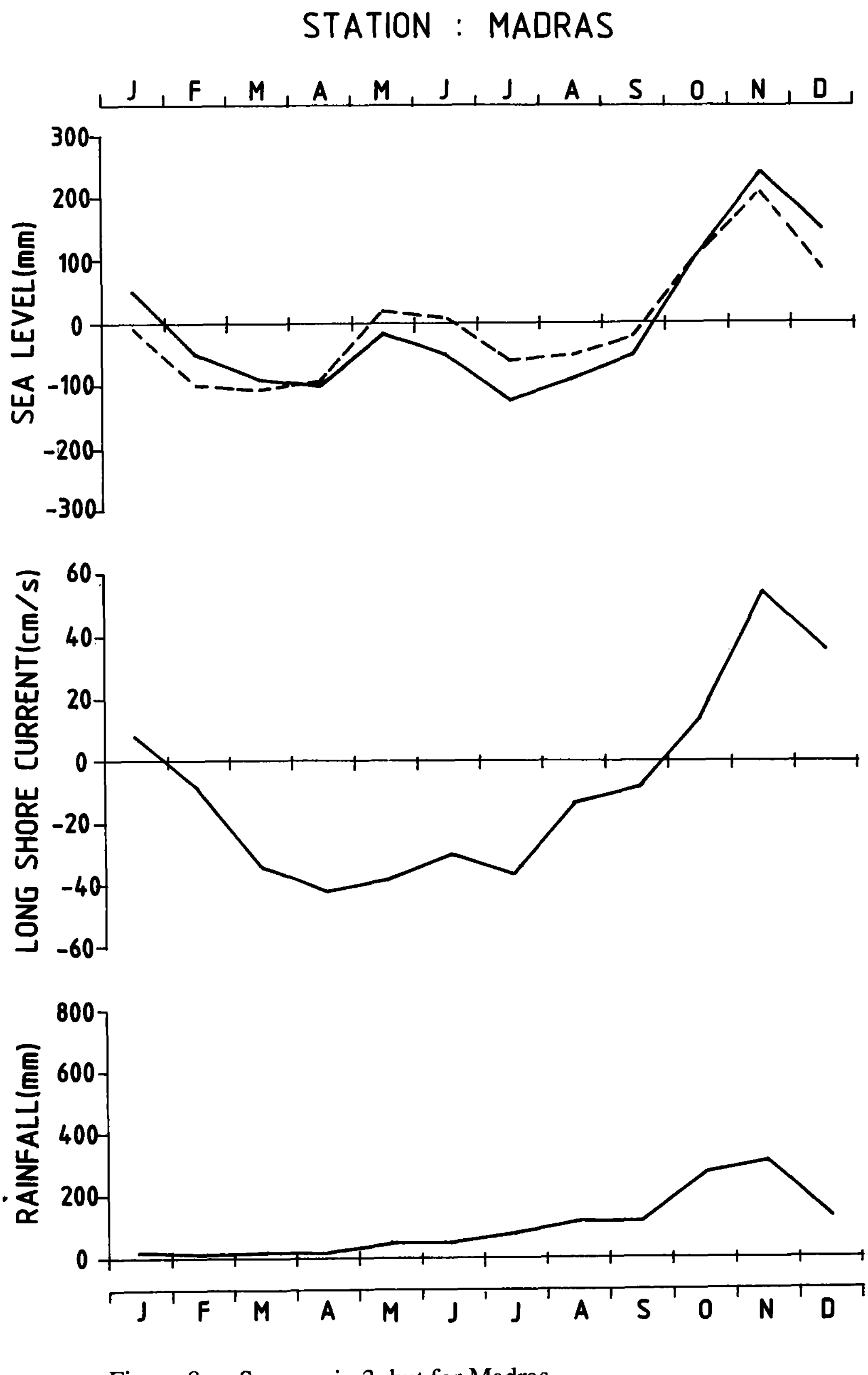

Figure 8. Same as in 3, but for Madras.

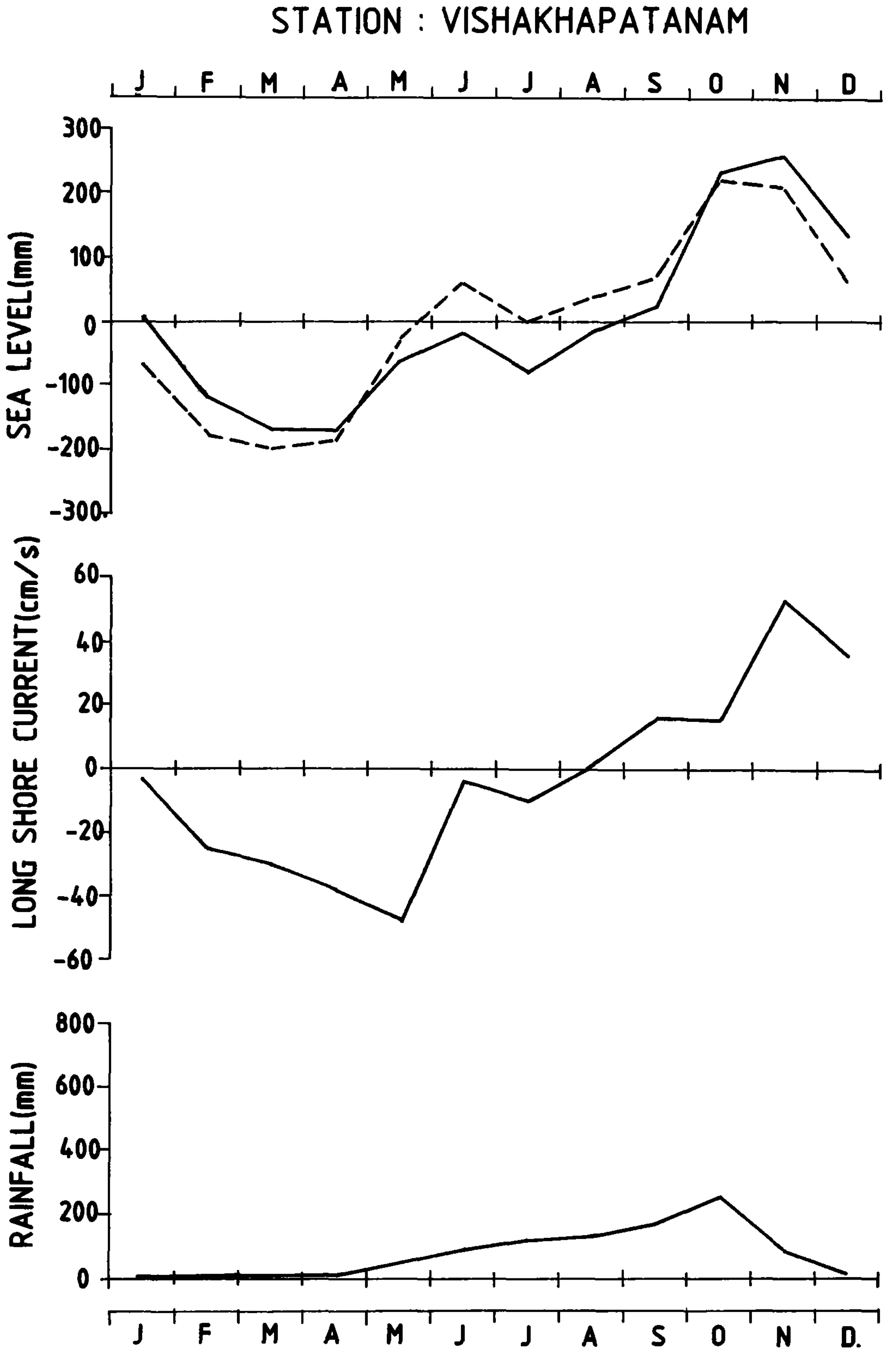

Figure 9. Same as in 3, but for Vishakhapatnam.

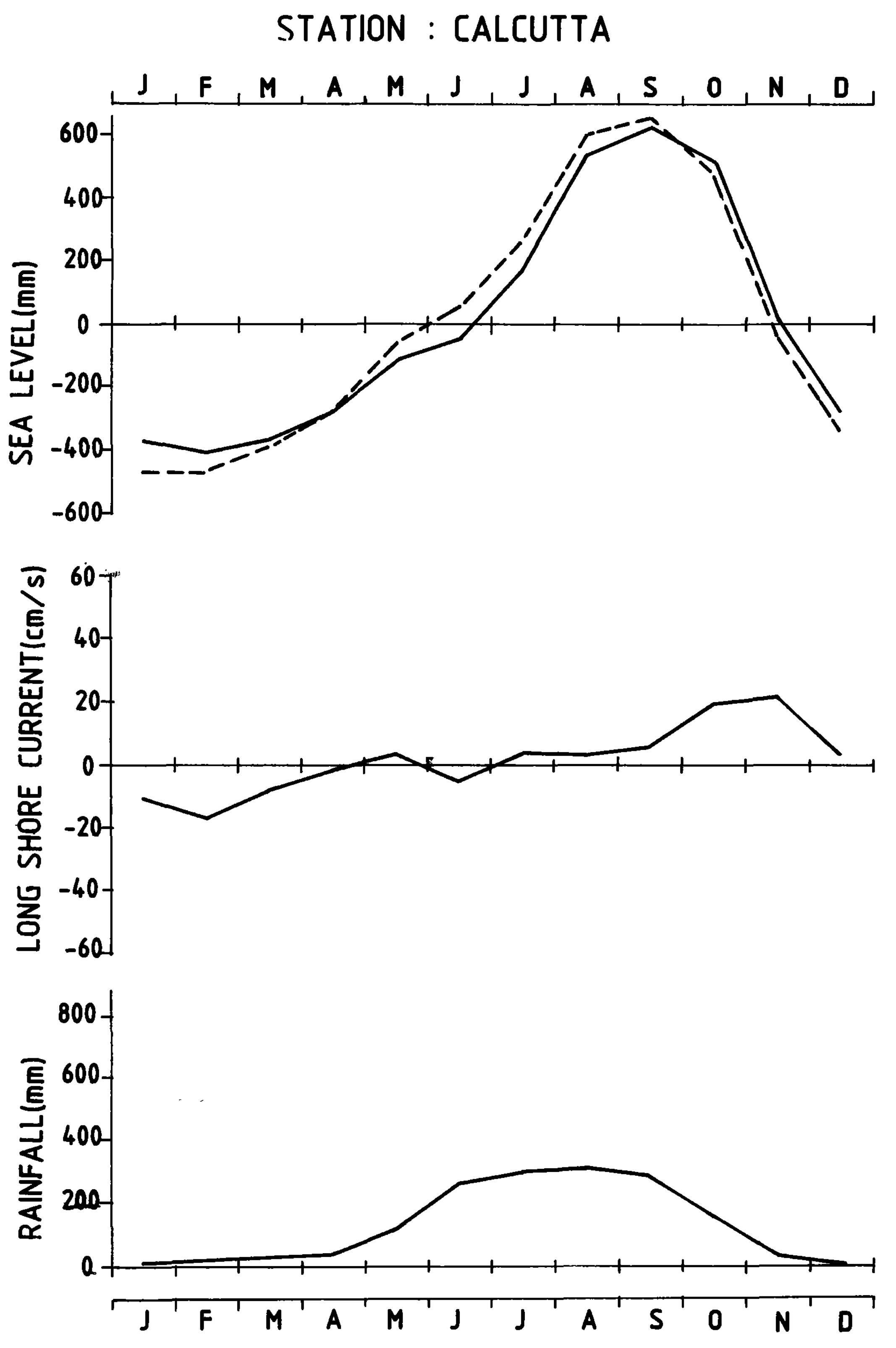

Figure 10. Same as in 3, but for Calcutta.

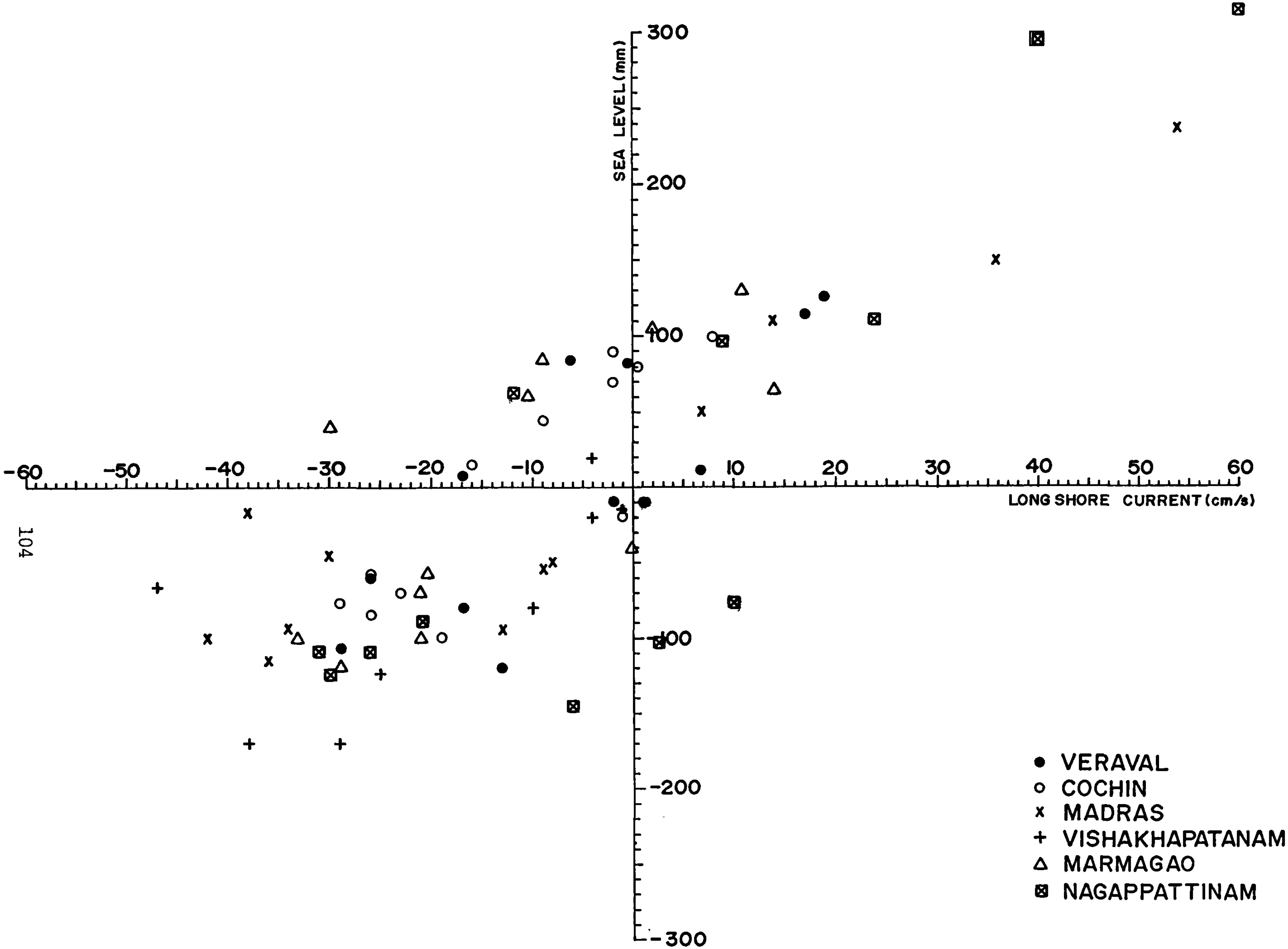

Figure 11. Scatter diagram of sea level against longshore current at Veraval, Marmagao, Cochin, Nagappattinam, Madras and Vishakhapatnam.

RED TIDES IN THE INDO-WEST PACIFIC REGION

GUSTAAF M. HALLEGRAEFF

CSIRO Marine Laboratories
GPO Box 1538
Hobart, Tasmania 7001
Australia

ABSTRACT

The most common red tide organisms in Indo-West Pacific waters, the blue-green alga *Trichodesmium erythraeum* and the dinoflagellate *Noctiluca scintillans*, produce mostly harmless water discolourations. Only in exceptional cases do such plankton blooms cause fish kills in sheltered bays due to the generation of anoxic conditions. The raphidophyte flagellate *Chattonella marina* has caused fish kills in India by releasing free fatty acids that destroy the gill tissues of fish. There is evidence that some red tide problems (e.g. *Chattonella, Noctiluca*) are aggravated by industrial pollution, but other plankton blooms (e.g. *Trichodesmium*) bear no relationship to eutrophication.

Red tides caused by toxic dinoflagellates, which produce potent neurotoxins that can find their way through fish and shellfish to man, are well documented from Europe, North America and Japan. At present, paralytic shellfish poisoning (PSP), diarrhetic shellfish poisoning (DSP) and ciguatera fish poisoning are virtually unknown from the Central Indian Ocean. Admittedly, surveys for the causative dinoflagellate species have been inadequate, and ethnic dietary habits (low shellfish consumption) may also play a role. Attention is drawn to the apparent spreading of *Pyrodinium bahamense* var. *compressa* (PSP) through the East Indies (Papua New Guinea, Philippines, Malaysia, Indonesia), and *Protogonyaulax tamarensis* has caused recent PSP fatalities in Thailand and possibly India.

It is recommended that a basic training workshop be conducted for local workers responsible for microscopic plankton taxonomy to enable them to recognize potentially toxic species and their benthic cysts. In addition, public health officials should be made aware of the clinical symptoms of various types of shellfish and fish poisoning to ensure that, if necessary, warnings can be issued promptly and effectively.

INTRODUCTION

Red tides are blooms of unicellular marine plankton algae which become so dense that they discolour the sea (e.g. Red Sea). Plankton blooms also may appear yellow, brown, green, blue or milky in colour dependent upon the organism involved. Red tide species in the Indo-West Pacific region include representatives of the blue-green algae, dinoflagellates and raphidophytes, with diatoms and prymnesiophytes only rarely involved. Most red tides are caused by motile or strongly buoyant species, and their high concentrations are achieved through a combination of high growth rates and vertical (behavioural) or horizontal (physical) aggregation. Dense plankton concentrations are therefore most strongly developed under stratified stable conditions, at high temperatures and following high organic input from land run-off after heavy rains.

The majority of plankton blooms appear to be completely harmless events, but under exceptional conditions non-toxic bloom-formers may become so densely concentrated that they generate anoxic conditions that cause indiscriminate kills of fish and invertebrates in sheltered bays (GRINDLEY and TAYLOR, 1962). Oxygen depletion can result from high respiration by the algae (at night or in dim light during the day), but more commonly it is caused by bacterial respiration during decay of the algal bloom. A basically different type of phenomenon involves the production by dinoflagellates of potent toxins that find their way through fish or shellfish to man. In this case low densities of the toxic algae in the water column may be sufficient to cause such illnesses in humans as paralytic shellfish poisoning (PSP), diarrhetic shellfish poisoning (DSP) and ciguatera fish poisoning. PSP can result both from eating bivalve shellfish and planktivorous fish (clupeotoxicity), while DSP is caused by eating shellfish, and ciguatera by eating tropical fish. The toxins involved evoke a variety of gastrointestinal and neurological symptoms in humans (Table 1), but rarely affect the nervous systems of fish or shellfish. Another group of toxins, called ichthyotoxins, selectively kill fish by inhibiting their respiration. On a global scale, close to 2000 cases of human poisoning through fish or shellfish consumption occur each year, and economic damage through reduced local consumption and reduced export of seafood products can be considerable.

Generally speaking, the numbers and intensity of algal blooms seem to be on the rise and their geographic extent seems to be spreading. While toxic plankton blooms appear regularly on a seasonal basis in temperate waters of Europe, North America and Japan (TAYLOR and SELIGER, 1979), they are still relatively rare in tropical regions. As a result, the Indo-West Pacific countries have little knowledge of how to deal with this environmental and economic threat. The present paper illustrates and discusses several potentially hazardous organisms that have been recognized in these tropical waters. Clinical symptoms of various types of fish and shellfish poisoning are summarized to assist their diagnosis by medical practitioners, and illustrations of the plankton organisms are provided to improve identification by local plankton workers. Blooms of the toxic species, in particular, need to be carefully monitored, and fish and shellfish products from affected areas should be tested for toxins by public health officials who should, if necessary, issue warnings promptly and effectively.

DESCRIPTION OF RED-TIDE ORGANISMS

Nowhere is there a more important need for correct taxonomic identification of plankton organisms than in the study of the toxic species. Red tides are often monospecific blooms, and understanding the autecology of the constituent species thus becomes crucial not only to understanding the bloom event but also when deciding on possible measures for its control. Some dinoflagellates (e.g. *Protogonyaulax*) produce benthic cysts that can seed further blooms, and the monitoring for these species must also take into account benthic cyst populations and sedimentary processes. Three categories of 'red tide organisms' are distinguished: species that produce mostly harmless water discolourations; species non-toxic to man but harmful to fish and invertebrates by damaging or clogging their gills; and species which produce potent toxins that can find their way through the food chain to man.

HARMLESS WATER DISCOLOURATIONS

The most common red tide organisms in the Indian Ocean, the blue-green alga *Trichodesmium* and the dinoflagellate *Noctiluca*, produce mostly harmless water discolourations. Only in exceptional cases do such plankton blooms cause fish kills in sheltered bays due to the generation of anoxic conditions.

Trichodesmium: This tropical blue-green alga produces seasonal (February — April) water blooms in the Andaman, Arabian, Java, Banda, Arafura and Coral Seas. These appear as yellow-grey (early bloom) or reddish-brown (late bloom) coloured windrows, occupying up to 40,000 square kilometers. The long filaments mass together to form raft-like (*T. erythraeum* EHR., Fig. 1a,b) or radiating or bundle-like aggregations (*T. thiebautii* GOMONT, Fig. 1c,d). At the start of the bloom the filaments usually appear throughout the water column, but during late bloom stages the strong gas vacuoles cause a massive rise of the algae to the surface layers. Differentiated cells within the centre of the colony are capable of fixing atmospheric nitrogen, which allows the algae to thrive under nutrient-impoverished oceanic conditions where they readily outcompete other phytoplankton. Wave action can break up the bundles and inactivate the central nitrogenase enzyme (CARPENTER and PRICE, 1976), which is why calm seas are a prerequisite for *Trichodesmium* blooms. The alga can be a nuisance to swimmers, but harmful effects on fish are seldom observed (DEVASSY et al., 1978) except in sheltered bays where the decaying bloom may generate anoxic conditions that can cause indiscriminate kills of fish and other marine fauna (e.g. CHACKO, 1942). An unusual mass death of corals caused by the decomposition of masses of *Trichodesmium* driven ashore by the wind has been recorded from New Caledonia (BAAS BECKING, 1951). *Trichodesmium* red tides ("sea sawdust")

were observed as early as 1770 during Captain COOK's voyage through the Coral Sea. There is no evidence of a relationship with industrial pollution.
Noctiluca scintillans (Macartney) Kofoid (= *Noctiluca miliaris* Suriray): This strongly buoyant, large, non-photosynthetic dinoflagellate (Fig. 2) causes spectacular water discolourations. Carotenoid globules within the cytoplasm (BALCH and HAXO, 1984) are responsible for "tomato soup" colored blooms in Japan, Hong Kong and Australia, whereas the presence of green prasinophyte endosymbionts (*Pedinomonas noctilucae*; SWEENEY, 1978) within *Noctiluca* may cause green water blooms in Indonesia, Malaysia, India and Thailand (SUBRAHMANYAN, 1954b, PRASENO and ADNAN, 1978). This dinoflagellate is mostly restricted to coastal waters and occurs especially in the vicinity of river mouths and following heavy rainfalls. *Noctiluca* has been known to bloom extensively off both the east (AIYAR, 1936) and west coasts of India, where it has been implicated in fishery decline (BHIMICHAR and GEORGE, 1950; PRASAD, 1953). No toxic effects are known, but it is possible that the high ammonia content of the vacuole irritates fish, which generally avoid the bloom areas.

Other Water Discolourations

A wide range of other non-toxic plankton blooms is known from the Indo-West Pacific region. The raphidophyte flagellate *Heterosigma akashiwo* (Hada) Hada (= *Olisthodiscus luteus* sensu Tomas, Leadbeater, Gibbs; Fig. 3) has caused red-brown water discolouration in Korea and Japan (YAMOCHI, 1984). The dinoflagellate *Gonyaulax polygramma* Stein (Fig. 4) has produced red water in the Arabian Sea (PRAKASH and SARMA, 1964; LEWIS, 1967), the dinoflagellate *Prorocentrum micans* Ehrenberg (Fig. 5) has caused brown water in New Zealand (CASSIE, 1981), and red tides by the dinoflagellate *Scrippsiella trochoidea* (Stein) Loeblich (Fig. 6) have resulted in anoxic conditions and fish kills in Australia (WHITELEGGE, 1891).

SPECIES DAMAGING OR CLOGGING THE GILLS OF FISH AND INVERTEBRATES

Selected species of diatoms, prymnesiophytes, raphidophytes and dinoflagellates can harm fish or invertebrates by damaging or clogging their gills.

Mucus-Producing Species

Diatoms are not usually included among the red tide organisms, but several species of the genus *Thalassiosira* (Fig. 7, e.g. T. *mala* Takano) can form gelatinous masses that may harm farmed oysters by clogging their gills (e.g. Japan; TAKANO, 1956). Similarly, colonies of the prymnesiophyte *Phaeocystis pouchetii* (Hariot) Lagerheim (Fig. 8) form irritant substances (acrylic acid) and mucilage. The latter can interfere with fishing by clogging gills of fish and bivalves and by fouling fishing nets (most recently in New Zealand, CHANG, 1983). A brown *Phaeocystis* bloom in the Arabian Gulf was originally mistaken for oil pollution. (M. BEHBEHANI, pers. comm.)

Species Damaging Fish Gills

Chattonella marina (Subrahmanyan) Hara et Chihara (= *Hornellia marina* Subr.) This raphidophyte flagellate (Fig. 9a) produces green-brown water discolourations which have been associated with fish kills in coastal waters of India (SUBRAHMANYAN, 1954a) and Japan. The alga releases into the seawater free fatty acids that destroy the gill tissues of fish (SHIMADA et al., 1983). Red tides by *Chattonella marina* and by the related species *Chattonella antiqua* (Hada) Ono (Fig. 9b) (not yet known from the Indo-Pacific) have caused multimillions of dollars' damage to the Japanese fishery of cultured yellowtail and seabream. In semi-enclosed basins, such as the Seto Inland Sea, the incidence of *Chattonella* red tides is steadily becoming more frequent due to increasing discharge of sewerage and industrial wastes.

Gyrodinium aureolum Hulburt; *Ptychodiscus* (*Gymnodinium*) *brevis* (Davis) Steidinger; *Gymnodinium nagasakiense* Takayama et Adachi The small, unarmoured dinoflagellate *Gyrodinium aureolum* (Fig. 10) is one of the most common red tide organisms in North European waters

(TANGEN, 1983), where its coffee-brown colored blooms have been associated with mortality of marine farmed fishes and benthic invertebrates. The alga has a destructive effect on the lamellar epithelium of fish gills (ROBERTS et al., 1983). This dinoflagellate has been positively identified from New Zealand and Tasmanian waters but has not caused fish kills in that region.

G. aureolum is morphologically similar to *Ptychodiscus* (*Gymnodinium*) *brevis* (Fig. 11a) and *Gymnodinium nagasakiense* (Fig. 11b). An unarmoured dinoflagellate resembling *Ptychodiscus* (*Gymnodinium*) *brevis* (Davis) Steidinger has been implicated in massive fish kills in Port Phillip Bay, Australia (WOOD, 1964), and *Gymnodinium nagasakiense* Takayama et Adachi (= *Gymnodinium* 'type 65') has caused fish kills in Korea and Japan (TAKAYAMA and ADACHI, 1984). *P. brevis* produces potent neurotoxic shellfish poisons (NSP) in the Gulf of Mexico (killing the fish by inhibiting respiration; STEIDINGER, 1983), but only weak toxicity thus far has been confirmed in *G. nagasakiense*. These species have not yet been recorded from the Indian Ocean region.

SPECIES TOXIC TO MAN

Some 20 out of the total number of 1500 extant species of dinoflagellates are known to produce potent toxins that can find their way through the food chain to man. Ten toxic dinoflagellate taxa occur in the Indo-West Pacific region.

Pyrodinium bahamense Plate var. *compressa* (Böhm) Steidinger, Tester et Taylor: This tropical dinoflagellate (Fig. 12) produces seasonal red tides (patches up to 300 km long) following enrichment by land runoff, especially near mangrove shores (MACLEAN, 1977). The alga can appear as single cells or in chains, and produces a spinose resting cyst that is involved in initiating the blooms. At present only the Indo-Pacific populations (var. *compressa*) are known to produce toxins, while non-toxic populations (var. *bahamense*) have been found in the tropical Atlantic (STEIDINGER et al., 1980). The toxins include a weak ichthyotoxin (which kills fish) and an array of PSP toxins. Because of its high toxic potential and its apparent spreading through the West Indies, this species is considered to be the "number one" red tide danger in the Indo-Pacific region with more than 700 human illnesses and 50 deaths attributed to it to date. The first fatal PSP cases were recorded in 1972 from Papua, New Guinea, in 1976 from Malaysia and in 1983 from the Philippines and Indonesia (MACLEAN, 1979; GACUTAN et al., 1985). This dinoflagellate is also known from the Red Sea and the Arabian Gulf but has not caused human illnesses there, possibly because of low shellfish consumption in these areas. Paralytic shellfish poisoning can result from eating bivalve shellfish (Table 2) and also planktivorous fish such as sardine and anchovy (clupeotoxicity). The toxicity of fish can be reduced by removing their gills and intestines before consumption, but shellfish can remain toxic for up to 5 months after *Pyrodinium* has disappeared from the water column.

Protogonyaulax tamarensis (Lebour) Taylor; *Protogonyaulax catenella* (Wedon et Kofoid) Taylor; *Gymnodinium catenatum* Graham: The cosmopolitan dinoflagellate *Protogonyaulax tamarensis* (Fig. 13a) can appear as single cells or in pairs. The alga causes widespread outbreaks of PSP in North America, Europe and Japan, but until recently was unknown from the Southern Hemisphere. *P. tamarensis* has now been implicated as the causative organism for a PSP case in 1983 in Thailand (SUDARA, 1984), and a morphologically similar species (to be described as a new taxon) was linked with fish and shellfish kills in 1983 in New Zealand (TAYLOR, 1984). Several cases of human poisonings after mussel consumption are also known from the East and West coasts of India (BHAT, 1981). The chain-forming *Gymnodinium catenatum* (observed in Tasmanian waters; HALLEGRAEFF and SUMNER, 1986) resembles *Protogonyaulax catenella* (Fig. 13b) in morphology and PSP toxins but does not have a cell wall divided into plates (MOREY-GAINES, 1982). All the above species produce smooth-walled agglutinous oval cysts (Fig. 13c) which may seed further blooms. The toxicity of these cysts may be ten times higher than that of the motile dinoflagellate cells. PSP occurred centuries before the modern industrial world developed and was known already to Indian tribes along the west coast of North America before Columbus' time.

Dinophysis fortii Pavillard; *Dinophysis acuminata* Claparede et Lachman; *Dinophysis acuta* Ehrenberg': These three dinophysoid dinoflagellates (Figs. 14a, b, c) are commonly encountered in temperate and tropical coastal waters, but they are seldom abundant (up to 10^3-10^4 cells l^{-1}). Seasonal population increases of *D. fortii* in Japan (YASUMOTO et al., 1978), of *D. acuta* in Norway and of *D. acuminata* in Holland, Spain, France and Ireland (KAT, 1983) have been implicated as the cause of diarrhetic shellfish poisoning (DSP). The Dinophysis-toxin involved has also been detected in New Zealand mussels. Humans can be poisoned through eating mussels and to lesser extent scallops and oysters. The clinical symptoms (Table 1) are easily confused with those of bacterial gastric infections, and DSP (first described in 1976 in Japan) may be much more common than realised at present. Control measures include seasonal closures of the shellfishery, purification of shellfish in laboratory tanks, and removal of the toxin-accumulating hepatopancreas from scallops before eating (not practicable with mussels and oysters).

Gambierdiscus toxicus Adachi et Fukuyo: This lens-like, benthic dinoflagellate (Fig. 15a) appears in epiphytic association with bushy red, brown and green seaweeds and can also be found free in sediments and coral rubble. The alga is probably circum-tropical in distribution, but thus far is only known from the Great Barrier Reef region (Australia), the Pacific Islands and possibly Mauritius and the Seychelles (D. ARDILL, unpublished data). *Gambierdiscus* has not yet been recorded from New Guinea, Indonesia, the Philippines, Thailand or India. Because of its dislike of land runoff, the species seldom occurs near large landmasses. Captain COOK suffered from the tropical fish poisoning "ciguatera" when visiting New Caledonia in 1774, but the causative dinoflagellate organism was only identified in 1978 (ADACHI and FUKUYO, 1979). The potent neurotoxins ciguatoxin and maitotoxin accumulate through the food chain, from small fish grazing on the coral reefs into the organs of bigger fish that feed on them. Control measures include eating only smaller fish and not eating species such as red bass, coral trout, chinaman fish, barracuda and moray eel (Table 3). Over 400 cases of ciguatera food poisoning are known from Australia (GILLESPIE, 1980) and over 3000 cases from French Polynesia. No adequate treatment is yet available. The symptoms persist for months and recur up to several years later (Table 1). The associated benthic dinoflagellates *Ostreopsis siamensis* Schmidt (Fig. 15b) and *Prorocentrum* lima (Ehr.) Dodge (Fig. 15c) also exhibit some toxicity (FUKUYO, 1981) similar to ciguatera.

CONTROL OF RED TIDES

Red tides can cause great economic damage, affecting both the seafood market and tourism. Destruction of fish and shellfish by anoxia and poisoning of fish and shellfish by toxin producing dinoflagellates are especially critical in countries that depend heavily on mariculture for protein. Cooking and other treatments of fish and shellfish do not destroy the toxins. The question arises as to whether these events can be controlled artificially. Various chemical and biological control methods that have been attempted have had mostly negative results. Copper sulphate was sprayed from planes to combat Florida red tides of *Ptychodiscus brevis* (STEIDINGER, 1983), but the killed algal cells released their endotoxins into the seawater and the decomposing algal mass generated anoxic conditions. Seed cultures of various predators, parasites or pathogens have been considered as candidates for biological control. However, the successful predator would accumulate the toxin and thus become a highly toxic vector to higher trophic levels in the food chain.

Moderate-scale coastal engineering involving deepening or filling affected areas or reshaping the coastline or diverting rivers may be a viable option to combat anoxic conditions in sheltered bays. Marine dredging operations should be alerted to the possible danger of seeding benthic dinoflagellate cysts into the water column, and caution should be used when transferring shellfish from one area to another because resistant microscopic cysts could easily be carried with them. Red tides that are related to increasing eutrophication (e.g. *Chattonella* in Japan) can be controlled by reducing discharge of sewerage and industrial wastes. On the other hand, phenomena such as PSP and ciguatera, which were known centuries before the modern industrial world developed, should be regarded as completely natural events. Careful monitoring of the causative dinoflagellates, of their cysts and of associated seafood products appears to be the only solution at present. Dependent upon the results, control measures may include elimination of the toxin-accumulating organs from fish or shellfish, avoidance of certain seafood species, depuration of shellfish in laboratory tanks and, in extreme cases, seasonal closures of particular fisheries.

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Table 1. Clinical symptoms of various types of fish and shellfish poisoning.

	Paralytic Shellfish Poisoning	Diarrhetic Shellfish Poisoning	Ciguatera
Causative organism	*Pyrodinium bahamense var. compressa, Protogonyaulaxxtamarensis; Protogonyaulax catenella; Gymnodinium catenatum*	*Dinophysis fortii; Dinophysis acuminata; Dinophysis acuta*	*Gambierdiscus toxicus, Ostreopsis siamensis; Prorocentrum lima*
Symptoms	mild case: tingling sensation or numbness around lips, gradually spreading to face and neck; prickly sensation in fingertips and toes; headache, dizziness, nausea, vomiting, diarrhoea severe case: incoherent speech; progression of stiffness and non-coordination of limbs; general weakness and feeling of lightness; slight respiratory difficulty; rapid pulse. Extreme case: Muscular paralysis; pronounced respiratory difficulty; choking sensation; death through respiratory paralysis may occur within 2 to 24 h after ingestion	after 30 min to a few hours: diarrhoea, nausea, vomiting, abdominal pain.	Symptoms develop within 12-24 hrs of eating fish. Gastro-intestional symptoms: diarrhoea, abdominal pain nausea, vomiting Neurological symptoms: Numbness and tingling of hands and feet; cold objects feel hot to touch; difficulty in balance; low heart rate and blood pressure; rashes In extreme cases, death through respiratory failure
Treatment	Patient has stomach pumped and is given artificial respiration. No lasting effects.	Recovery after 3 days, irrespective of medical treatment.	No antitoxin or specific treatment is available. Neurological symptoms may last for months and even years. Calcium may help relieve symptoms.

Table 2. Shellfish species implicated in diarrhetic (d) and/or paralytic shellfish poisoning (p) (Summarised from toxic incidents in Europe, Japan, North America, Philippines and South Africa).

Anadara antiquata	(ark shell)	p
Chlamys nipponensis	(scallop)	d,p
Chlamys nobilis	(noble scallop)	p
Choromytilus meridionalis	(black mussel)	p
Clinocardium nuttalli	(cockle)	p
Crassostrea gigas	(Pacific oyster)	p
Donax serra	(white mussel)	p
Fulvia mutica	(cockle)	d
Gomphina melanaegis	(venus clam)	d,p
Mactra chinensis	(surf clam)	d
Meretrix lamarckii	(Noto callista)	d
Mya arenaria	(softshell clam)	p
Mytilus coruscu	(mussel)	d,p
Mytilus edulis	(blue mussel)	d,p
Panope generosa	(geoduck)	p
Patinopecten yessoensis	(scallop)	d,p
Pecten albicans	(scallop)	d
Perna viridis	(green mussel)	p
Prototheca staminea	(little neck clam)	p
Ruditapes philippinarum	(venus clam)	d,p
Saxidomus giganteus	(butter clam)	p
Schizotherus nuttalli	(horse clam)	p
Spondylus butleri	(thorny oyster)	p
Venerupis japonica	(Japanese little neck clam)	p

Table 3. Fish species implicated in ciguatera fish poisoning.

Caranx melampygus	(spotted trevally)
Cheilinus undulatus	(green wrasse)
Epinephelus fuscoguttatus	(flowery cod)
Lethrinus miniatus	(long-nose schnapper)
Lutjanus bohar	(red bass)
Lutjanus gibbus	(paddle-tail)
Lutjanus monostigma	(one-spot sea perch)
Lutjanus rivulatus	(Maori chief)
Plectropoma leopardus	(coral trout)
Sphyraena barracuda	(barracuda)
Sphyraena picuda	(barracuda)
Symphorus nematophorus	(chinaman fish)
Variola louti	(coronation trout)

Table 4. **Summary of red tide orgamisms and their geographic distribution in the Indo-Pacific region.**

Harmless Blooms	**India**	**Thailand**	**Malaysia**	**Indonesia**	**Philippines**	**New Guinea**	**Korea**	**Hong Kong**	**Japan**	**Australia**	**New Zealand**
Trichodesmium erythraeum	0	+	+	+		+			+	+	
Noctiluca scintillans	+	+	+	+				+	+	+	
Heterosigma akashiwo							+		+	+	
Gonyaulax polygramma	0					+				+	
Prorocentrum micans											+
Scrippsiella trochoidea										0	
Noxious Species Damaging Fish Gills											
Thalassiosira mala									+	+	
Phaeocystis pouchetii										+	+
Chattonella marina	0								0		
Chattonella antiqua									0		
Gyrodinium aureolum										+	+
Gymnodinium nagasakiense							+		+		
Dinoflagellates Toxic to Man											
Pyrodinium bahamense v. compressa			•	•	•	•					
Protogonyaulax tamarensis	•?	•							•		0
Gymnodinium catenatum										•	
Dinophysis fortii/acuminata									•	+	+
Gambierdiscus toxicus										•	

\+ species present; 0 fish kills; • human illness

FIGURE CAPTIONS

Figure 1a. *Trichodesmium erythraeum;* filament 7-12 μm wide, 60-750 μm long.

Figure 1b. *Trichodesmium erythraeum*; raft-like aggregations, 1 cm size.

Figure 1c. *Trichodesmium thiebautii*; filament 3-16 μm wide, 1-2 mm long

Figure 1d. *Trichodesmium thiebautii*; radiating and bundle-like aggregation, 1 cm size.

Figure 2. *Noctiluca scintillans*; 260-720 μm diameter.

Figure 3. *Heterosigma akashiwo*; 11-24 μm long, 8-10 μm wide.

Figure 4. *Gonyaulax polygramma*; 29-66 μm long.

Figure 5. *Prorocentrum micans*; 35-70 μm long.

Figure 6. *Scrippsiella trochoidea*; 16-36 μm long, 20-23 μm broad.

Figure 7. *Thalassiosira mala*; cells 4-9 μm diameter, up to 2000 cells per colony.

Figure 8. *Phaeocystis pouchetii*; cells 4-8 μm diameter, colonies up to several cm.

Figure 9a. *Chattonella marina*; 20-30 μm wide, 30-55 μm long.

Figure 9b. *Chattonella antiqua*; 50-130 μm long.

Figure 10. *Gyrodinium aureolum*; 27-34 μm long, 17-32 μm wide.

Figure 11a. *Ptychodiscus brevis*; 15-35 μm long.

Figure 11b. *Gymnodinium nagasakiense*; 15-35 μm long.

Figure 12. *Pyrodinium bahamense* var. *compressa*; 35-68 μm diameter.

Figure 13a. *Protogonyaulax tamarensis*; 30-44 μm diameter, ventral and dorsal view.

Figure 13b. *Protogonyaulax catenella*; chain; cells 30-37 μm long.

Figure 13c. *Protogonyaulax tamarensis*; benthic cyst, 25 x 40 μm.

Figure 14a. *Dinophysis fortii*; 60-70 μm long.

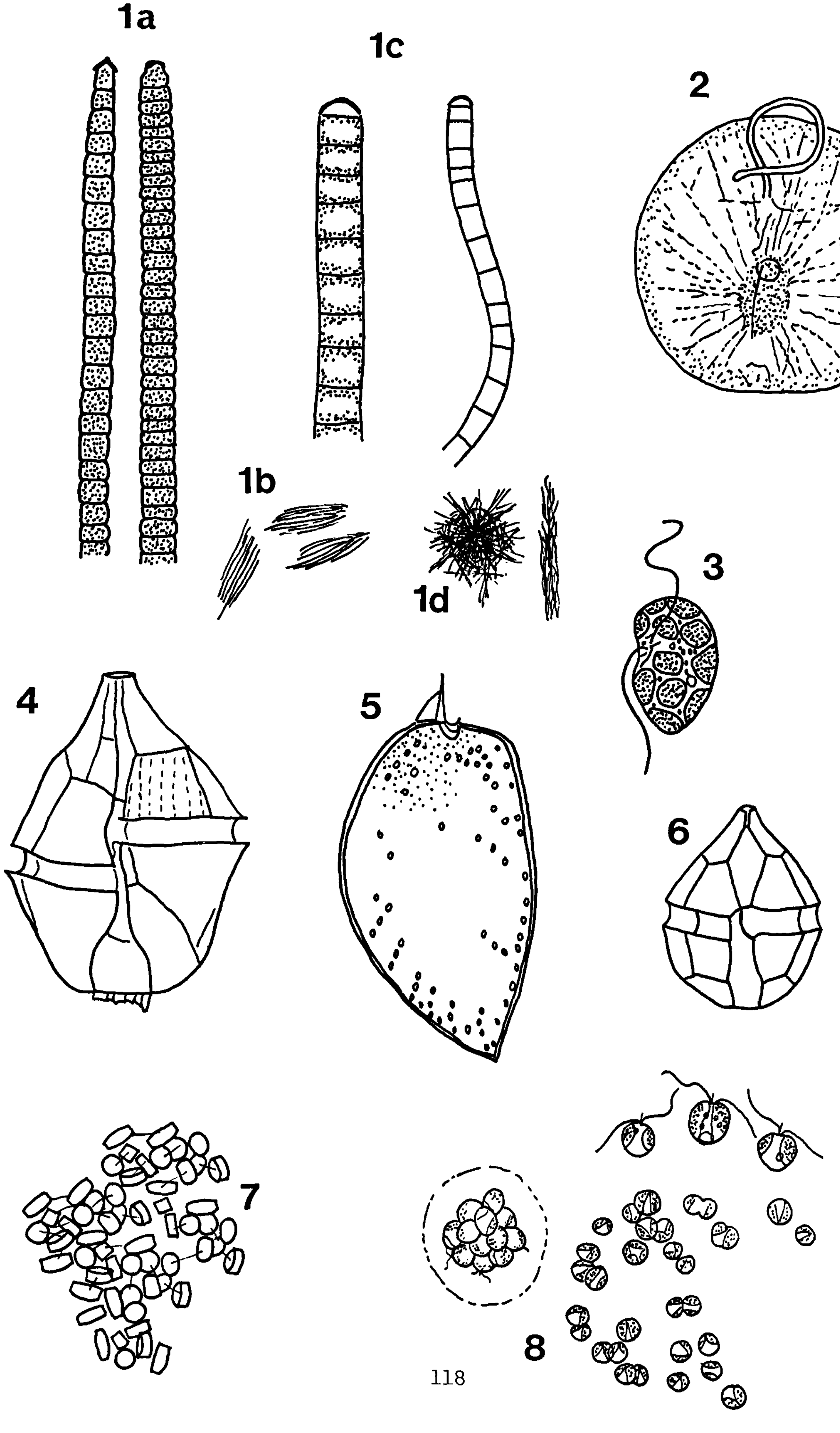
Figure 14b. *Dinophysis acuminata*; 25-47 μm long.

Figure 14c. *Dinophysis acuta*; 54-94 μm long.

Figure 15a. *Gambierdiscus toxicus*; 42-140 μm diameter; epitheca (e) and hypotheca (h).

Figure 15b. *Ostreopsis siamensis*; 45-90 μm diameter, epitheca (e) and hypotheca (h).

Figure 15c. *Prorocentrum lima*; 30-40 μm long.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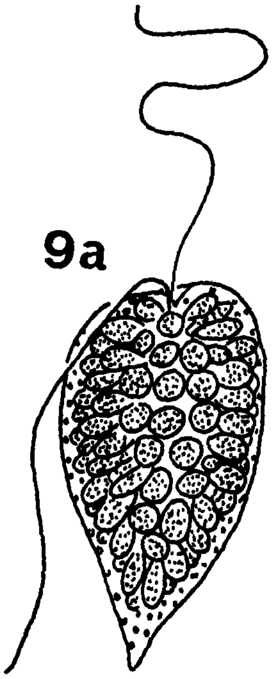

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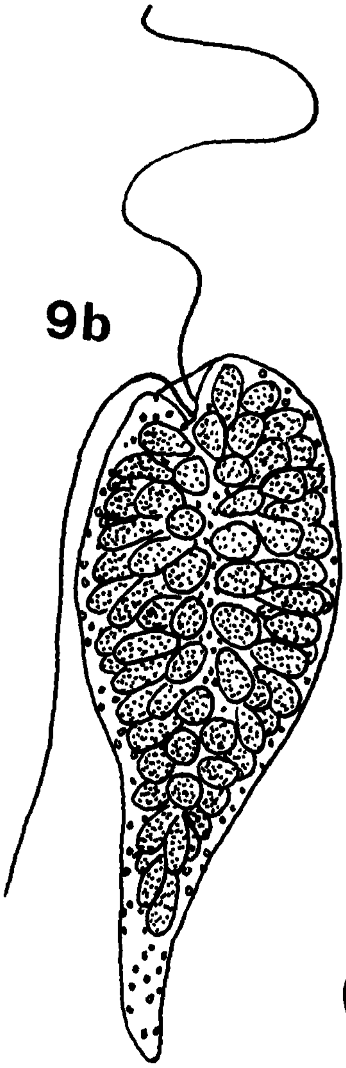

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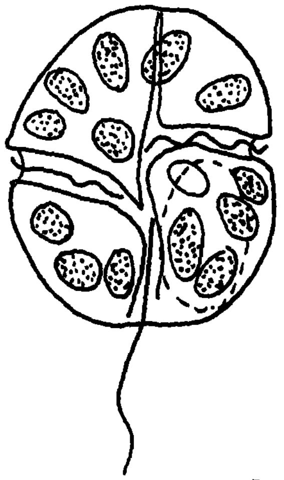

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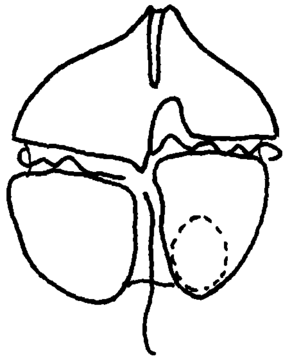

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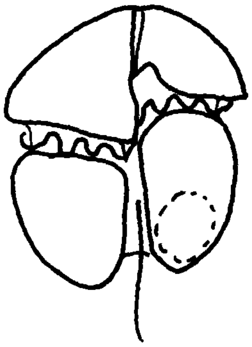

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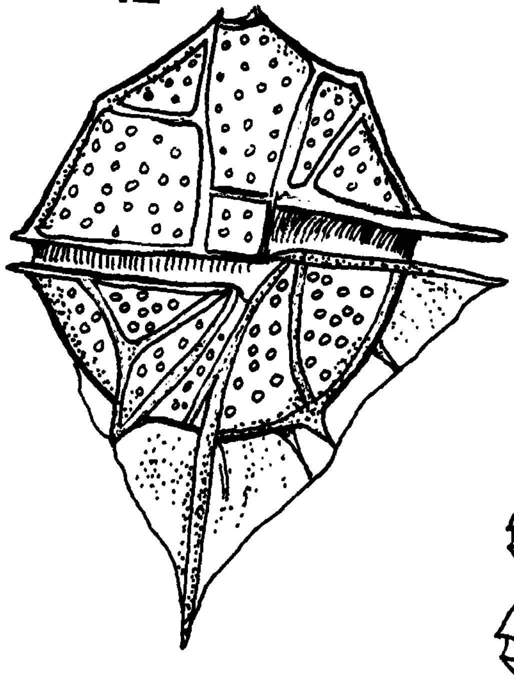

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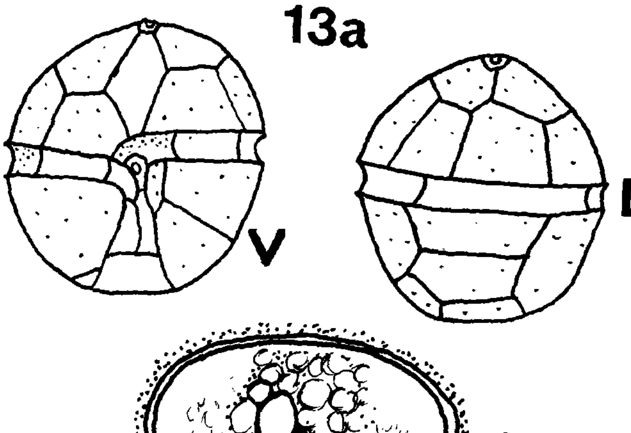

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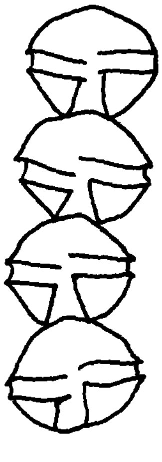

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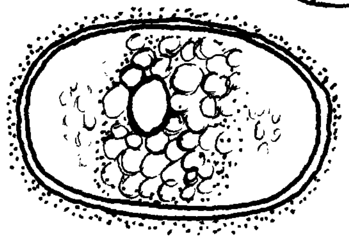

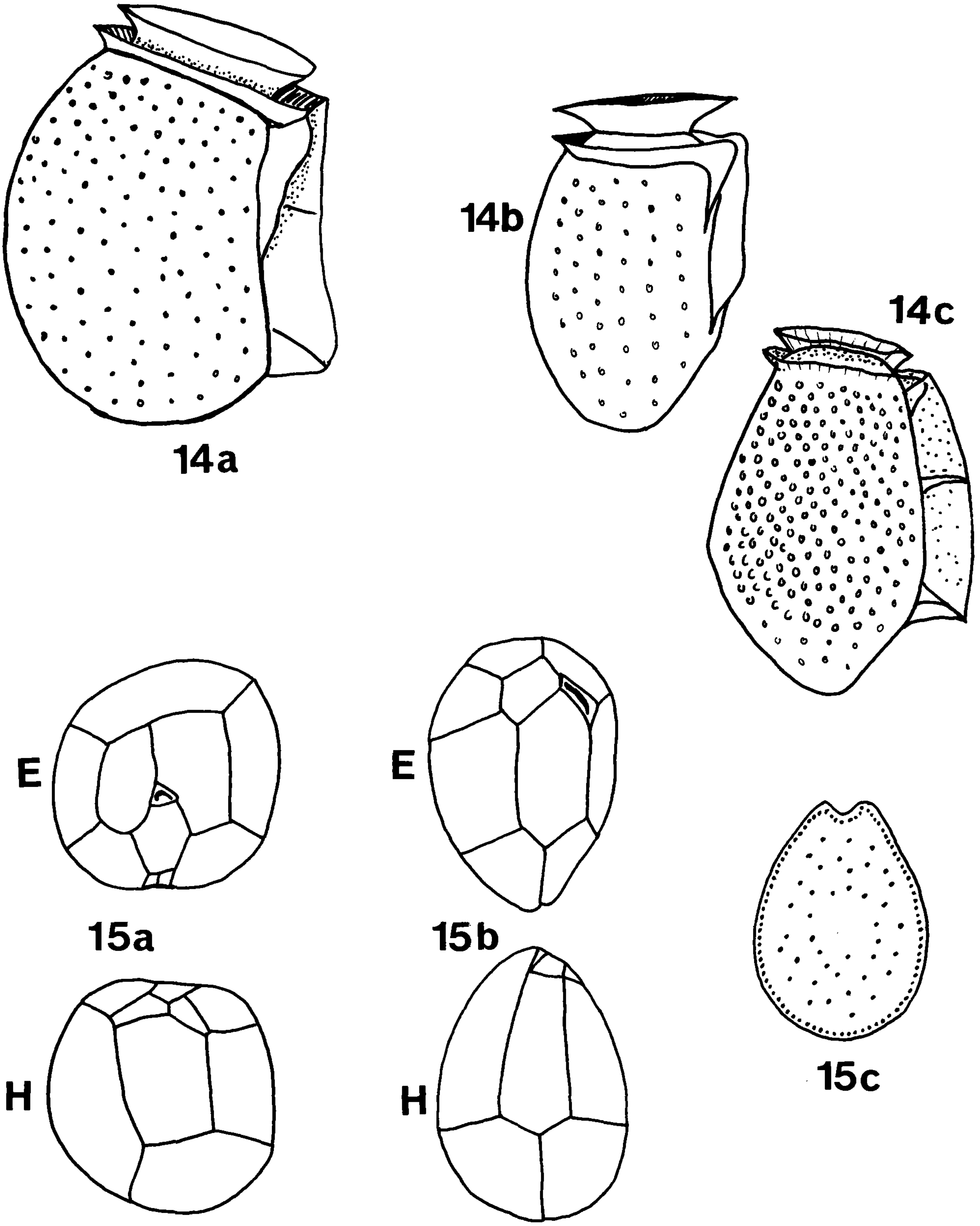
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ECONOMIC AND POLLUTION ASPECTS

MINERAL RESOURCES OF THE INDIAN OCEAN AND RELATED SCIENTIFIC RESEARCH

H.N. SIDDIQUIE*, A.R. GUJAR, N.H. HASHIMI
A.B. VALSANGKAR and B. NAGENDER NATH

National Institute of Oceanography
Dona Paula, Goa-403 004, India
*Deceased

ABSTRACT

The sea floor of the Indian Ocean and the continental margins bordering the ocean are covered by a wide variety of terrigenous, biogenous and chemogenous mineral deposits.

The humid tropical climate of some of the land areas bordering the Indian Ocean accelerates weathering of the source rocks. This, coupled with the large river runoff and wave and current conditions, favours the formation of a variety of terrigenous heavy mineral placer deposits. As such, the beach and offshore placer deposits of the Indian Ocean are particularly abundant. Heavy mineral placers are known from the beaches of South Africa, Mozambique, Tanzania, Kenya, western India, Sri Lanka, Burma, Malaysia, Indonesia and Western Australia. The exploration of these deposits in offshore areas has been particularly encouraging off Mozambique, western India, Sri Lanka, Malaysia and Western Australia.

Biogenous deposits in the Indian Ocean include shallow-water and shelf bank sediments and carbonate and siliceous oozes in the deep sea. Biogenous deposits are a low-priced commodity and are likely to be exploited only locally for construction, cement and chemicals where onshore limestone deposits are not available. Shells and corals have local values for semi-precious and ornamental purposes. Calcareous sands, shells and corals are being exploited on a large scale in India for the manufacture of cement, and exploration in some of the areas has indicated reserves of over a thousand million tonnes. In many countries, especially islands whose economy is largely dependent on tourism, restraint should be excercised on the exploitation of these deposits. A study of these deposits undoubtedly would lead to a better understanding of their formation, production and sustainable yields and, in turn, to better management.

Phosphorite deposits in the Indian Ocean are reported from continental margins (South Africa and western India) and seamounts (eastern and western Indian Ocean). Onshore deposits, however, are sufficient to meet requirements for the near future. Exploitation of the marine deposits would be feasible in those coastal areas where the cost of imported fertilisers is higher than the cost of marine mining. Future exploitation of phosphorites may provide a significant stimulus to the dominantly agricultural economies of many Indian Ocean countries.

The polymetallic nodules in the Indian Ocean cover an area of 10-15 million km^2, and the resources are estimated to be about 0.17 trillion tonnes. A study of over 900 chemical analyses from 350 stations shows that the deposits in most basins are submarginal. Deposits in the central Indian Ocean may be paramarginal (Ni + Cu + Co over 2.47 percent and abundance more than 5 kg/m^2).

Exploration of the mineral resources of the Indian Ocean not only would add to an inventory of the resources in the world oceans but also would lead to a greater understanding of the formation of marine mineral resources.

Present capabilities (trained manpower, equipment and vessels) for exploration for marine non-living resources in the region are confined to only a few countries. In this context, the programmes for exploration of non-living resources will require (i) compilation of the basic data for planning (ii) exploration of the inner shelf, (iii) exploration of the outer shelf, and (iv) exploration of the deep sea. In countries with limited manpower in marine geosciences, a beginning could be made by interactions between physical, chemical and biological oceanographers. This would be useful to identify the areas for the exploration for terrigenous deposits (heavy minerals placers), biogenous deposits (shells and corals) and chemogenous deposits. After initial planning, further work could be initiated on smaller vessels (even fishing trawlers) in the nearshore areas and later, with the acquisition

of equipment and larger vessels, on the continental shelf and in the deep sea. Indian scientists have developed collaboration programmes with Sri Lanka and Seychelles, and programmes are now being drawn up for collaboration with Mauritius.

INTRODUCTION

The Indian Ocean (Area: 74.917 x 10^6 km^2, Water Volume: 291.945 x 10^6 km^3, Average Depth: 3897 m) is the third largest of the world oceans. The lands bordering the ocean contain almost 40 percent of the world's population and contribute substantially to some of the essential mineral raw materials for the world economy; i.e., oil, tin, iron and manganese ores, mica and chromite.

The present paper reviews the surficial mineral resources of the Indian Ocean, excluding those in bedrock (oil, gas and sulphur etc.) and the metalliferous brine deposits of the Red Sea.

The minerals for the review have been grouped as (i) Terrigenous (ii) Biogenous and (iii) Chemogenous. The present review covers the Indian Ocean as defined by FAIRBRIDGE (1966), extending from the east of the African continent to the west of the Australian continent and from the Indian sub-continent to Antarctica.

SEDIMENT TYPES

TERRIGENOUS SEDIMENTS

Heavy mineral placers are known from the beaches of many countries bordering the Indian Ocean (Fig. 1). The placers contain ilmenite, rutile, zircon, magnetite, monazite, garnet, kyanite and tin in significant proportions. Many of these placers are economically exploited and contribute significantly to the world production (Table 1). The offshore extension of some of these placers has also been explored, but at present only offshore tin placers are being exploited.

The present onshore deposits of heavy mineral placers can provide a clue about the nature and distribution of offshore deposits in submerged valleys and beaches on the continental shelf. Though a large number of placers are known to occur on the beaches of the Indian Ocean, offshore extensions of only a few have been explored. These are reviewed below:

Mozambique: Known onshore placer deposits at Pebana, Vila Luiza and Micaune contain ilmenite, zircon, rutile and minor amounts of monazite. The R.V. VALDIVIA in 1971 and 1973 carried out detailed geological and geophysical surveys on the shelf off northern Mozambique (BEIERSDORF et al., 1980) to locate marine placer deposits. The heavy minerals included ilmenite (TiO_2 49.5; FeO 33.1 and Cr_2O_3 0.18 %), zircon (ZrO_2 63.05, TiO_2 0.71 %), rutile (TiO_2 87.7 %) magnetite, garnet, etc. Sediments on the Zambesi delta between water depth of 30 to 60 m are estimated to contain 50 million tons of ilmenite, 0.9 million tons of rutile and 4 million tons of zircon.

Tanzania: Placers containing garnet (4.3 million tonnes), ilmenite (3.6 million tonnes), rutile (0.4 million tonnes), kyanite, zircon, magnetite and monazite were reported (DUYVERMAN, 1981) along the coast of Tanzania. Heavy mineral concentrations, ranging up to 23 percent, decrease northwards except at Ndega beach. The data on either the size of resources or their offshore extension are not available.

India: Placers are known from a number of localities along the Indian coast (Fig. 2), of which the beach and dune placers along Kerala coast are best known. These deposits are reported to contain 17 million tonnes of ilmenite, 1 million tonnes of rutile, 1.2 million tonnes of zircon and 0.2 million tonnes of monazite. Similar deposits with varying proportions of ilmenite, zircon, magnetite, and garnet have been described from Maharashtra, Tamil Nadu, Andhra Pradesh and Orissa (SIDDIQUIE et al., 1984 and references therein). The total reserves of ilmenite are estimated to be 163 million tonnes, and production in 1981 was 0.189 million tonnes (INDIAN BUREAU OF MINES, 1981).

The extension of the heavy mineral placers from the beach to the offshore areas was a subject of speculation until recently. Some of the richest deposits in India occur along the 25 km-long stretch of the coast from Neendakara to Kavankulam in Kerala. PRABHAKARA RAO (1968) explored the shallow offshore areas of this stretch to a depth of about 10 to 12 m and suggested that these sediments have been derived from the area immediately to the east. The heavy minerals (4 to 56 percent) in these sediments consist of kyanite, sillimanite, zircon, garnet, ilmenite, leucoxene and rutile.

Ilmenite placers along the Konkan coast contain 17 to 74 percent ilmenite, and reserves have been estimated at about 4 million tonnes (MANE and GAWADE, 1974). The National Institute of Oceanography initiated the offshore surveys of Konkan in 1975 to (i) delineate the offshore extension of the promising onshore placers, (ii) understand the geology, structure, recent sedimentation and tectonic history of the area, (iii) define the environment of deposition of heavy minerals and formation of placers, (iv) evaluate the techniques for the exploration of the deposits, and (v) examine the feasibility of their economic exploitation. A detailed survey of the 130 km long strip along the Konkan coast (Fig. 3) was carried out during 1975-80. Heavy mineral concentrations range up to 90 percent (Fig. 4), and the assemblage is dominated by ilmenite (Fig. 5) and magnetite (Table 2) with minor quantities of augite, epidote, zoisite, apatite, tourmaline, topaz, hornblende, rutile, zircon, limonite and kyanite. The magnetites in the sands are titanomagnetite or finely intergrown with ilmenite. SIDDIQUIE et al. (1979) reported 42 to 57 percent TiO_2 in the ilmenite, and trace element analyses of these heavy mineral concentrates indicate appreciable concentrations of vanadium (up to 0.5 percent) and chromium (SAHOO, 1980).

A representative seismic profile off the Konkan coast (Fig. 6) shows four to five major reflectors consisting of sand, clay, sand with pebbles, weathered trap and trap. The thickness of heavy mineral-bearing sands varies from 2 to 10 m (Fig. 6), and thicker sands (more than 10 m) are associated with river mouths and buried ancient river channels and local sand lenses. This is also confirmed from shallow cores and boreholes drilled in Mirya Bay; the sands are underlain by weathered basalt and in turn Deccan basalts (Late Mesozoic to Early Tertiary). Magnetic surveys were carried out to correlate the magnetic anomalies with heavy mineral concentrations, basement configuration and structural trends.

The surveys carried out so far indicate that the ilmenite-bearing sands cover over 96 km^2 of the seabed. Assuming an average ilmenite concentration of 10 percent and a minimum thickness of 1 m for ilmenite-bearing sands, the reserves are inferred to be about 12.5 million tonnes. Seismic profiles indicate that the thickness of ilmenite-bearing sands ranges from 2 - 10 m and these extend to water depths of about 20 m over an area of 436 km^2. Thus the probable reserves in the area are many times more than the onshore reserves of 4 million tonnes. A comparison shows a higher concentration of ilmenite in the offshore than in the onshore sands. The extensive onshore and offshore deposits offer possibilities for an integrated development. Further exploration of offshore placers will be guided by the proximity to onshore placers and upon examination of the relict sands on the middle and outer shelves (Fig. 7).

Sri Lanka: The Sri Lanka beaches have been fairly well studied, and placers of ilmenite, rutile, monazite, zircon and garnet (KRISHNAN, 1968; HEARTH, 1983) are reported from many localities. However, little exploration for offshore heavy mineral placers has been carried out. Some of the localities are:

i) Beruwela in Induruwa (west coast) - extensive deposits of monazite with zircon and garnet.
ii) Beaches near Kokkilai lagoon (east coast) - 75% ilmenite forming an estimated reserve at about 4 million tonnes.
iii) Pulamoddai and Triukkovil beaches - 25% zircon.

Southeast Asia: Southeast Asia is the world's major producer of tin (WERNER, 1983); about 92 percent of the world's estimated resources are in this region. The tin province stretches for over 3600 km, from north Burma through peninsular Thailand and west Malaysia to the Tin Islands (Singkep, Bangka and Billiton) of Indonesia.

Offshore tin placers have been formed from the weathering and erosion of onshore deposits. These formations have been influenced by the interaction of the terrestrial and marine environments.

Although most of the production of cassiterite in southeast Asia has been from deposits on land, during the past ten years a considerable tonnage also has been obtained from submarine placers off the Indonesian Tin Islands and to the east of the Thai island of Phuket. A comparatively unimportant part of the belt is located in Laos, southeast Thailand and Anambas and Natuna Islands. Titaniferous magnetite is recovered from some of the beaches of Indonesia. Deposits of ilmenite, titaniferous magnetite, monazite and zircon also are reported in the offshore areas of western Indonesia.

The production of tin has declined since 1982 mainly because of depleted reserves, weak metal prices and cutbacks on exports.

Australia: The Australian beaches are well known for their heavy mineral placers. The total reserves proven and estimated for the east coast beaches of Queensland and New South Wales are 20 million tonnes and those of the west coast are 68 million tonnes. Between them these beaches produce over 90 percent of the world's rutile, 80 percent of zircon, 70 percent of monazite, >25 percent of ilmenite (CHARLIE, 1981).

BROWN (1971) reported encouraging results from offshore and estimated that a grade of 0.2 to 0.22 percent rutile and zircon could be mined. Reserves of over 375 million tonnes have been mapped, and indications are that another 500 million tonnes are present.

BIOGENOUS SEDIMENTS

Biogenous sediments cover large areas of the seabed (Fig. 1). They include pelagic oozes in the deep sea and coral and shell deposits in shallow areas.

Deep Sea

Deep-sea biogenic oozes accumulate in areas of high productivity (20°N - 40°S) and (for carbonate oozes) at depths shallower than the carbonate compensation depth (CCD). The approximate area covered by oozes in the Indian Ocean (FAIRBRIDGE, 1966) is given below:

Ooze	Area 10^6 km^2	Latitude
Globigerina	40.45	Between 20°N and 40°S
Pteropod	00.74	Small patches NW of Australia
Diatomaceous	14.98	South of 50°S
Radiolarian (with red clays)	18.72	Between 10°N and 40°S

The deposits match in physical properties and chemical composition a wide variety of rocks/ materials used on land, but they occur at too great depth and distance from land to be of any immediate economic interest.

Shallow Areas

Calcareous deposits dominate many shallow areas of the continental shelf, atolls and banks. Various maps by WELLS (1954) and ROSEN (1971) indicate that the highest generic diversity of reef corals occurs in a belt extending from 20°N to 20°S from Sumatra to Laccadives-Maldives, northern

Madagascar and the Red Sea. Within this area lie a number of atolls, coral reefs, submerged reefs and banks. Some of the banks are very large (i.e., Seychelles Bank covers 43,000 km^2, Saya de Malha, 40,000 km^2, Nazareth Bank 20,000 km^2 and the banks of the Chagos Archipelago 13,500 km^2). The banks have general maximum depths of 33-90 m and most have a shallow rim (reef) at 8-10 m depth around at least part of their periphery.

Deposits of calcareous shells also occur in the shallow areas of the continental shelf. The deposits usually occur in areas of high benthic productivity and where influx of terrigenous sediment is low (MILLIMAN, 1974). The deposits in some of the areas comprise sub-recent fossil shells which have been or are being concentrated by recent or sub-recent marine processes.

The best known deposits off India are in the Jamnagar (Gulf of Kutch) district of Gujarat, North Kanara district of Karnataka, Alleppey and Kottayam (backwaters of Vembanad Lake), Kozhikode, Mallapuram and Cannanore districts of Kerala, Chinglepet district of Tamil Nadu and Nellore district (backwaters of Pulicat Lake) of Andhra Pradesh. The present production from these deposits is over 130,000 tonnes (Fig. 8).

The other, though minor, use of corals and shells from shallow waters is as ornamental and semi-precious stones. Among the Indian Ocean countries, substantial amounts of unworked shell have been exported by Indonesia (2,395 tonnes - 1978), India (151 tonnes - 1978) and Maldives (5 tonnes - 1977). India is becoming increasingly important as a supplier of ornamental shells, and exports have risen from 20 tonnes in 1969 to 466 tonnes in 1979. Exports of shells from Kenya reached a peak of 107 tonnes in 1978. Ornamental corals are also exported by Indonesia (5 tonnes - 1976), Maldives (0.038 tonnes - 1977) and India (0.016 tonnes - 1978).

For many types of shells and corals a concern rightly has been expressed regarding the depletion of their stocks, and the need for conservation has been emphasized. Because of this concern some countries have imposed a ban on exports and others are regulating the exploitation on a sustainable yield basis.

CHEMOGENOUS SEDIMENTS

Phosphorite

Phosphorite deposits have been found along continental margins (South Africa and western India) and around seamounts (eastern and western Indian Ocean). The continental margins of South Africa, East Africa, southern Arabia, western India and the Andamans are marked by strong upwelling and provide non-depositional environments which are conducive to the formation of phosphorite (Fig. 9, Table 3).

Agulhas Bank: The Agulhas Bank phosphorites are the largest known seabed deposits (BATURIN, 1982, and references therein), occupying a belt about 600 km long and 100 km wide at depths of 100 to 500 m. They consist of phosphoritic nodules and phosphatized limestones. The total reserves of phosphorite and phosphatized rocks are estimated to be 140 million tonnes of P_2O_5, assuming a mean deposit thickness of 1 m.

Andaman: TIPPER (1911) reported phosphorite nodules off the northern Andaman Islands. In the late 1960's extensive sampling for phosphorite was carried out in the same vicinity, but the results were not encouraging.

Western Continental Margin of India: The detailed surveys of the shelf between Goa and Bombay have identified a series of well defined (25 m height and 150 km length) NNW-SSE trending ridges. These ridges are composed of algal, coral and shelly limestones. Analyses of rock samples from these ridges show a wide range of P_2O_5 concentrations, between 0.8 and 11 percent (NAIR, 1985).

Seamounts: BEZRUKOV (1973) noted the occurrence of phosphorite on Christmas Island. Subsequently phosphorite also was dredged from a seamount, 13°45'S and 99°56'E, depth 3689 m, during R.V. VITYAZ's 54th cruise on the Cocos Ridge. These rocks consisted of phosphorite (22.7

to 32.5 percent of P_2O_5) and limestone blocks with lenticular intercalations of dense white and unconsolidated brownish phosphorite (BEZRUKOV et al., 1979).

Recently ORV SAGAR KANYA dredged samples of phosphorites from the Error Seamount in the northwestern Indian Ocean (ANON, 1983), at a depth of 400-460 m. The haul contained corals, rocks and limestones with fragments of brownish to yellowish phosphorites.

Polymetallic Nodules

The first polymetallic nodules from the Indian Ocean was collected by HMS CHALLENGER (1872-1876). Subsequently, WISEMAN (1937), MERO (1965), CRONAN and TOOMS (1967, 1968, 1969), GLASBY (1970, 1972), CRONAN (1972, 1975a,b), BEZRUKOV and ANDRUSCHENKO (1974), SOREM and BANNING (1976), CRONAN and MOORBY (1976, 1981), SIDDIQUIE et al. (1978), COOLEY et al. (1979), and FRAZER and FISK (1980, 1981) carried out mineralogical and geochemical studies on nodules, encrustations and sediments from various basins in the Indian Ocean (Fig. 9, Table 4).

Polymetallic nodules from 7000 locations on the ridges, seamounts and deeper basins of the Indian Ocean have been collected. Based on these data, the area covered by the nodules is estimated to be 10-15 million km^2 and the resources about 0.15 trillion tonnes.

BEZRUKOV and ANDRUSCHENKO (1974), CRONAN and MOORBY (1976) and SIDDIQUIE et al. (1978) noted the lack of extensive nodule deposits in the Arabian Sea and the Bay of Bengal due to high sedimentation. A nodule-rich belt similar to that in the eastern Pacific is located in the Central Indian Basin, which perhaps extends further east to the Wharton Basin. BEZRUKOV (1962) estimated from photos that the nodules cover up to 83.2 percent of the sea floor, with abundances up to 43 kg/m^2. GLASBY (1973) estimated that the distribution of nodules ranges from 5 to 200 nodules/m^2 in the Indian Ocean.

Central Indian Basin: (Figs. 10-16): The average depth is about 5120 m. CRONAN and MOORBY (1981) reported that the nodules associated with the siliceous sediments are enriched in Mn, Cu and Ni. The Cu/Ni ratio (>1.0) is similar to the high-grade nodules from southwest of Hawaii in the Pacific Ocean (FRAZER and WILSON, 1980). The nodules associated with pelagic clays, especially in the northern area, are depleted in Mn (15.4%), Cu (0.34%), Ni (0.6%) and Zn (0.07%), and enriched in Fe (12%), Co (0.17%) and Pb (0.1%) compared to average contents of siliceous oozes. Nodules associated with terrigenous sediments are lower in average Mn (14.2%), Cu (0.22%) and Ni (0.34%). The nodules contain mainly todorokite. Many workers (BEZRUKOV and ANDRUSCHENKO, 1974; SIDDIQUIE et al., 1978 and CRONAN and MOORBY, 1981) have pointed out that the nodules at many locations have a higher average Cu (0.74%), Ni (0.86%), Co (0.14%), Mn (20%) and Zn (0.1%) than the other basins in the Indian Ocean. FRAZER and WILSON (1980) indicated that para- and sub-marginal grades of nodules cover an area of 0.7 x 10^6 km^2 between 10° to 16° S in the Central Indian Basin. Based on the grade and abundance of nodules, FRAZER and WILSON (1980) and CRONAN and MOORBY (1981) concluded that only the nodules from the Central Indian Basin meet the criteria for first generation mining. However, available data on the grade and abundance are still inadequate, and detailed exploration is necessary for the identification of a prime area and estimation of resources.

Wharton Basin: (Figs. 10-16): The average depth is 5260 m. CRONAN and MOORBY (1981) reported higher Mn, Cu and Ni and lower Fe, Co and Pb compared to the Indian Ocean average. Todorokite is common in the nodules. Nodules associated with siliceous oozes have significantly higher values of Mn (20.3%), Cu (0.56%), Ni (0.7%) and Zn (0.1%) and a higher average Mn/Fe ratio (1.78), and those associated with pelagic clays have lower Mn (16.1%), Cu (0.26%), Ni (0.48%), Pb (0.07%), Zn (0.06%) and Mn/Fe ratio (1.4). The Mn/Fe ratio (1.4) for the nodules with pelagic clays is higher than for similar nodules in the other basins in the Indian Ocean. The limited data on the abundance and grade (3 stations with a range of <5 to 15 kg/m^2) suggest that ore grade nodules in the Wharton Basin do not cover a large area.

South Australian Basin: (Figs. 10-16): Average depth is 4710 m; sediments in the basin range from pelagic clays to carbonate oozes. The nodules are marked by comparatively high averages of Mn (19%), Ni (0.68%), Cu (0.34%), and Zn (0.08%) and lower Co (0.17%) and Pb (0.08%). Nodules from the carbonate ooze region are higher in average Mn (19.7%), Cu (0.34%), Ni (0.71%) and Zn (0.84%) but lower in Fe (13.0%), Co (0.17%) and Pb (0.08%) than those from pelagic clays. The economic potential of possible resources in the basin is judged to be sub-marginal (FRAZER and WILSON, 1980).

Crozet Basin: (Figs. 10-16): The average depth is 4570 m, and the sediments include siliceous oozes, pelagic clays and carbonates. Nodules associated with each sediment type show marked differences in composition and mineralogy (LECLAIRE and PERSEIL, 1979). Average Fe, Cu, Pb and Zn contents are comparable to the Indian Ocean average, but the Mn/Fe ratio (0.79) is lower. Average values are near to those of the first two basins, whereas Mn (12.2%), and Cu (0.18%) are lower. Nodule concentrations vary from >5-10 kg/m^2.

Madagascar Basin: (Figs. 10-16): The average depth is 4620 m. The nodules in the basin are characterized by low average contents of Mn (12.5%), Ni (0.21%), Cu (0.12%) and Zn (0.05%) and high contents of Fe (16.4%) and Co (0.27%); the Mn/Fe ratio (0.75) is intermediate between the nodules on mid-ocean ridges and seamounts and aseismic ridges. The nodules associated with carbonate oozes and pelagic clays do not show significant compositional variations except for Co (0.31%) and Pb (0.12%). KOBLENTZ- MISHKE et al. (1970) reported high biological activity, which is perhaps reflected in the higher Mn/Fe ratio. Delta-MnO_2 is common in the nodules from seamounts in the basin.

Mozambique Basin: (Figs. 10-16): The average depth is 4880 m. The nodules, which are not widespread (VINCENT, 1972), are characterized by large variations in Mn and Fe and the Mn/Fe ratio (1.0 to 1.21). The average values for Cu (0.17%), Ni (0.27%), Co (0.12%) and Zn (0.05%) are much lower than the other basins in the Indian Ocean except perhaps the Madagascar Basin. Nodules associated with terrigenous sediment are higher in average Mn (11.9%), Fe (14.4%), Cu (0.10%), Ni (0.3%), Co (0.13%), Pb (0.12%) and Zn (0.06%) compared to the nodules from carbonate oozes.

Seychelles-Somali Basins: (Figs. 10-16): The average depth is 4500 m. The nodules average values for Mn (16%), Cu (0.22%) and Ni (0.41%) and Mn/Fe ratio (1.17) are higher than those reported for the Crozet, Madagascar and Mozambique Basins. Copper is higher (0.28%) in the nodules compared to the pelagic clays, whereas other elements are enriched in the nodules relative to the carbonate ooze.

PROSPECTS

TERRIGENOUS

The heavy mineral placers on the beaches and offshore areas of the Indian Ocean are perhaps more widespread and richer than those in other oceans. This is perhaps related to the source areas/rocks and favourable humid tropical climate in some of the land areas bordering the ocean which accelerates weathering of the rocks, erosion and transport of large volumes of sediments by large rivers. The heavy minerals are initially deposited as the rivers enter the sea and on the beaches where they are further sorted and concentrated by currents and waves.

The onshore placers provide the best clue for the exploration of the offshore deposits. While the offshore deposits in the vicinity of some onshore placers have been explored, most offshore areas still remain to be explored, a necessary step to assess the potential of these offshore deposits. A study of the Quaternary history of the continental shelves, the rise of the sea level, and the distribution of sediments (especially relict sediments) is also necessary in the areas where the placers occur. Such studies would not only provide data on the mineral potential of the shelf, but would also provide a conceptual framework for the exploration for offshore placers.

BIOGENOUS

Large areas of the deep and shallow Indian Ocean are covered by biogenous sediments. Calcareous and siliceous sediments, however, are low-priced commodities, and it is doubtful whether any deep-sea deposits could be of economic interest in the near future. The increase in demand for construction material, however, might lead to the exploitation on a comparatively large scale of calcareous deposits (i.e., corals and shells) from shallow areas.

CHEMOGENOUS

Systematic exploration of phosphorites has been carried out only on the continental margins of South Africa and India. The upwelling and non-depositional environments, particularly off the Somali and the Arabian coasts, need to be explored, as well as those areas where phosphorites already are indicated (the western shelf of India and the Andaman islands). Moreover, the Indian Ocean has a number of seamounts, and only a few (eastern Indian Ocean and Error Seamounts) have been explored. New data will add to an inventory of the resources and, under favourable situations, the deposits may even be exploited in the near future.

In terms of area covered and estimated resources, the polymetallic nodules in the Indian Ocean are second only to the Pacific Ocean. Available data show, however, that most Indian Ocean nodules do not meet the requirement of a first generation mine site, except perhaps those in the Central Indian Ocean. The exploration and exploitation of mineral deposits in the Indian Ocean (Fig. 17) will lead to a greater understanding of the formation and development of marine sediment deposits.

PLANNING

MARINE MINERAL EXPLORATION

In view of the limited expertise, equipment and research vessels, the exploration for marine mineral resources in the Indian Ocean region may have to take place in discrete and sequential phases: compilation of basic data for planning, exploration of the inner shelf, exploration of the outer shelf and the slope, and lastly the deep sea. In countries with limited expertise in marine geosciences, a limited beginning could be made by association of scientists from other disciplines (i.e., physical, chemical and biological oceanography) in the compilation of the basic data to identify target areas for exploration.

The planning for marine mineral exploration would require compilation of the available data from the literature (published and unpublished) and maps. A beginning could be made with the preparation of maps from notations on navigational charts (in addition to the seabed information or samples collected from research vessels) to prepare sediment distribution on the continental margins. These would provide some basic information on the distribution of sediments, shells, corals and rocks, and would lead to identification of the environments of sedimentation and the types of mineral deposits. It must be emphasized, however, that besides marine mineral exploration, it is necessary that basic research in marine geosciences also be encouraged, for the basis for further marine mineral exploration.

Sediment distribution maps combined with a study of the geology of the drainage basins would be useful for identifying potential placer areas and the expected minerals in these placers.

Offshore placers are likely to occur in areas covered by terrigenous sediments and near known onshore placers. Physical oceanographic studies also would be useful to indicate and to identify areas in which the energy regime is sufficient for such deposits to be formed. Collected data could be used for postulating models which would aid in the development of exploration strategies and techniques.

Biogenous deposits (i.e., corals and shells) are typical of tropical waters, and it would be appropriate if scientific studies result not only in economic potential and utilization, but also in achieving a basic understanding of the environmental parameters necessary for formation of these deposits. The basic studies on calcareous deposits also could provide insights into the formation of reservoir rocks for petroleum.

The countries in the Indian Ocean region with limited resources perhaps initially could study chemogenous deposits in shallower areas of the continental margins, likely areas for the formation of phosphatic deposits (i.e., upwelling and non-depositional environment), as these would provide valuable clues in understanding both recent and ancient deposits. This could be followed later by the exploration of the deep-sea deposits, i.e. polymetallic nodules.

PRESENT INDIAN COLLABORATION PROGRAMMES

Indian scientific institutions already have collaborative programmes with Sri Lanka and Seychelles, and a programme is being discussed with Mauritius. As a part of the programme with Sri Lanka, NIO has organized a cruise in Sri Lankan waters, and some Sri Lankan scientists have been trained in marine techniques at Goa. At the request of the Government of Seychelles, the R/V GAVESHANI surveyed the Exclusive Economic zone (EEZ) of Seychelles and collected interesting information on the topography, sediments and the distribution of polymetallic nodules of this island state. As a part of this programme, parallel work was carried out in the nearshore areas; the resources of seaweeds were assessed, and the morphology of the coast and rate of siltation in the harbour were studied.

At the request of Government of Mauritius, NIO is designing a programme to survey the EEZ of Mauritius. This survey will be begun after approval by the two Governments.

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Table 1. Heavy mineral placers, resources and production in the countries bordering the Indian Ocean (in thousand tonnes unless otherwise specified).

Country Offshore	Mineral	Resources		Production		Status of
		Onshore	Offshore	Onshore	Offshore	Exploration
1	2	3	4	5	6	7
South Africa	Titanium sand	NA	NA	41.74 (1979)	NA	Offshore occurrences reported
	Tin	NA	NA	2.7 (1983)	NA	
	Ilmenite	+99770 (1981)	NA	46.26 (1978)	NA	
	Rutile	+ 5442 (1981)	NA	45.25 (1981)	NA	
	Zircon	10884 (1981)	NA	P63.49 (1985)	NA	
	Monazite	+ 10.88 (1980)	NA	NA	NA	
	Garnet	4988 (1980)	NA	P11.79 (1985)	NA	
Mozambique	Garnet	NA	NA	11.2 (1979)	NA	
	Ilmenite	11791 (1980)	50,000 (1980)	NA	NA	
	Rutile	1088 (1980)	900 (1980)	NA	NA	
	Zircon	NA	4,000 (1980)	NA	NA	
Tanzania	Garnet	4280 (1980)	NA	30 kg (1979)	NA	Onshore placers reported
	Zircon & Kyanite	2329 (1980)	NA	8 kg (1979)	NA	
	Ilmenite	3628 (1980)	NA	NA	NA	
	Rutile	408 (1980)	NA	NA	NA	
Kenya	Garnet	NA	NA	274 kg (1979)	NA	
Egypt	Ilmenite	9070 (1980)	NA	NA	NA	
India	Ilmenite	163620 (1981)*	12500 (1982)	189 (1981)	NA	Offshore exploration partly carried out to the western shelf and planned for other areas on the western and eastern shelf. The offshore reserves are likely to be many times more than presently estimated.
	Rutile	8170 (1981)	NA	87 (1981)	NA	
	Zircon	20640 (1982)	NA	12 (1981)	NA	
	Monazite	1075 (1981)	NA	4.2 (1983)	NA	
	Garnet	52940 (1981)	NA	5.4 (1982)	NA	
	Kyanite	172200 (1981)	NA	34 (1982)	NA	
	Sillimanite	11990 (1981)	NA	12 (1982)	NA	
Sri Lanka	Ilmenite	4000 (1982)	NA	45.0 (1983)	NA	
	Zircon	1361 (1980)	NA	6.0 (1983)	NA	
	Rutile	272 (1980)	NA	13.0 (1983)	NA	
	Garnet	NA	NA	P90.7 (1985)	NA	
Bangladesh	--	--	--	--	--	Onshore placers of monazite and zircon are reported

Table 1. (continued)

Country Offshore	Mineral	Resources Onshore	Resources Offshore	Production Onshore	Production Offshore	Status of Exploration
1	2	3	4	5	6	7
Burma	Tin	907 (1980)	NA	4.2 (1982)	NA	Offshore dredging is being carried out from 1982.
Thailand	Tin (concentrated)	5560 (1980)	NA	4.2 (1982)	9.08(1982)	Offshore dredging for tin is being carried out but
			NA	0.78 (1979)	NA	production has now fallen
	Ilmenite	NA	NA	0.78 (1979)	NA	due to the depletion of the
	Monazite	NA	NA	0.72 (1979)	NA	reserves in the easily
	Zircon	NA	NA	2.0 (1983)	NA	exploited shallow areas.
Malaysia	Tin (concentrated)	4700 (1980)	NA	41.3 (1983)	NA	Production of tin has declined compared to 1982
	Ilmenite	7814 (1981)	NA	145.0 (1981)	NA	mainly because of the
	Monazite	NA	NA	2.0 (1979)	NA	combined effects of
deplet-						
	Zircon	181 (1980) NA		2.0 (1979)	NA	ing reserves weak metal prices and reserves cut back on export.
Described						as world's largest alluvial tin deposit in any one location, the Kuala Langate tin field will be jointly dredged by the Selonger State Government and Malaysia
Mining						Corp. If this project is successful it would help the country to maintain
its						position as a leading producer even with depleting reserves.
Indonesia	Tin	2630 (1980)	NA	26.5 (1983)	12.7 (1979)	Offshore dredging is being
	Ilmenite	907 (1980)	NA	NA	NA	carried out.

Table 1. (continued)

Country Offshore	Mineral	Resources Onshore	Resources Offshore	Production Onshore	Production Offshore	Status of Exploration
1	2	3	4	5	6	7
Australia	Tin	913 (1982-83)	NA	9.5 (1983)	NA	Offshore exploration has been completed in some areas.
	Ilmenite	42800 (1982)	NA	888 (1983)	NA	
	Rutile	9000 (1982)	NA	200 (1983)	NA	
	Zircon	13400 (1982)	NA	3832 (1983)	NA	
	Monazite	300 (1982)	NA	NA	NA	
	Garnet	5896 (1980)	NA	P18 (1985)	NA	
Madgascar	--	--	--	--	--	Onshore placers of ilmenite, monazite and zircon are reported.

P: Projected +: Reserve *: Inferred of Konkan, Maharashtra (SIDDIQUIE et al., 1982)

NA: Not available

From, INDIAN MINERALS YEAR BOOK (1977), (1979) and (1981)
U.S. BUREAU OF MINES (1979)
MINING JOURNAL, ANNUAL REVIEW (1982), (1983) and (1984)
MINERAL FACTS AND PROBLEMS, BUREAU OF MINES BULLETIN 671, US DEPARTMENT OF INTERIOR, 1980.

Table 2. The range (in percent) of sand, silt, heavy minerals, magnetic and non-magnetic fractions and ilmenite in the offshore placers of Konkan Coast, Maharashtra, India.

Bay	Sand	Silt	Heavy Minerals	Magnetite	Non-Magnetic	Ilmenite
Jaigad	20-98	1-24	20-70	10-40	1-10	1.30
Ambwah	20-98	1-20	5-40	1-71	3-15	3-30
Varvada	33-99	1-27	15-82	1.10	1-8	10-64
Kalbadevi	48-99	1-42	7-91	1-38	5-60	3-52
Mirya	12-99	1-37	5-72	3-48	1-51	1-35
Ratnagiri	8-99	2-69	1-78	1-59	1-50	1-22
Pawas	3-73	1-27	1-25	1-03	7-80	1-20
Purnagad	1-90	1-20	1-91	1-23	1-10	1-58
Vijaydurg	1-85	1-15	1-40	1-07	1-9	1-21

From SIDDIQUIE et al. (1982)

Table 3. Phosphate, resources and production (in million tonnes) in the countries bordering the Indian Ocean.

Country		Resources		Production	
South Africa		1000	(1982)	2.742	(1983)
Tanzania		NA		0.01	(1982)
Kenya (Guano)		NA		0.019	(1978)
Egypt		NA		0.65	(1982)
Iraq		NA		1.2	(1982)
Pakistan		NA		0.00065 per day plant capacity	
India	Phosphorite	129.8	(1981)	0.38	(1982)
	Apatite	8.8	(1981)	0.016	(1981)
	Guano	0.12	(1981)	NA	
	Phosphatic nodules	4.4	(1981)	NA	
Sri Lanka		* 40	(1984)	NA	
Thailand		NA		0.0045	(1979)
Indonesia		NA		0.001	(1979)
Australia		2000	(1980)	0.285	(1978)
Seychelles (Guano)		NA		0.005	(1979)
Christmas Islands		NA		1.095	(1983)

NA: Not available * Inferred
From: U.S. BUREAU OF MINES (1980)
MINING JOURNAL, ANNUAL REVIEW (1983), (1984)
INDIAN MINERAL YEAR BOOK (1981)

Table 4. Chemical composition of polymetallic nodules from the Indian Ocean

Basin/Ridge	No. of locations	No. of analyses	Maximum Mn	Fe	Ni	Cu	Co	Ni+Cu+Co
Central Indian Ocean	48	84	19.33	12.75	0.71	0.55	0.12	1.26
Wharton Basin	81	105	17.99	11.73	0.49	0.33	0.20	1.03
South Australian Basin	13	25	19.00	13.40	0.68	0.34	0.17	1.19
Crozet Basin	31	68	13.49	15.93	0.33	0.15	0.20	0.68
Madgascar Basin	28	64	14.55	19.60	0.28	0.14	0.34	0.68
Mozambique Basin	46	101	11.82	16.05	0.41	0.12	0.21	0.73
Seychelles/Somali Basin	37	323	18.12	16.61	0.36	0.12	0.40	0.85
Carlsberg Ridge	28	73	12.67	17.50	0.25	0.11	0.20	0.58
Average Indian Ocean	299	818	15.42	15.75	0.40	0.16	0.30	0.88

Basin/Ridge	No. of locations	No. of analyses	Minimum Mn	Fe	Ni	Cu	Co	Ni+Cu+Co
Central Indian Ocean	48	84	1.80	4.50	0.50	0.06	0.04	0.15
Wharton Basin	81	105	3.47	3.80	0.06	0.04	0.03	0.13
South Australian Basin	13	25	2.43	5.35	0.03	0.01	0.03	0.07
Crozet Basin	31	68	5.00	8.00	0.10	0.09	0.08	0.18
Madgascar Basin	28	64	8.50	9.70	0.06	0.06	0.08	0.20
Mozambique Basin	46	101	0.79	4.25	0.01	0.05	0.01	0.70
Seychelles/Somali Basin	37	323	3.72	1.33	0.09	0.02	0.03	0.14
Carlsberg Ridge	28	73	1.50	9.60	0.02	0.02	0.03	0.07
Average Indian Ocean	299	818	0.79	1.33	0.01	0.01	0.01	0.07

From: UNOET/SIO Data Bank

Table 4. Chemical composition of polymetallic nodules from the Indian Ocean (continued)

Basin/Ridge	No. of locations	No. of analyses	Mn	Fe	Ni	Average Cu	Co	Ni+Cu+Co
Central Indian Ocean	48	84	32.25	24.90	1.57	1.64	0.39	3.60
Wharton Basin	81	105	26.26	18.76	1.10	1.01	0.68	2.79
South Australian Basin	13	25	27.60	19.20	0.95	0.63	0.31	1.89
Crozet Basin	31	68	18.60	23.30	0.74	0.31	0.45	1.50
Madgascar Basin	28	64	19.50	26.45	0.75	0.29	0.69	1.63
Mozambique Basin	46	101	23.80	42.50	1.20	0.32	0.61	2.13
Seychelles/Somali Basin	37	323	32.30	24.13	0.79	0.48	1.04	2.31
Carlsberg Ridge	28	73	24.30	22.70	0.81	0.30	0.41	1.52
Average Indian Ocean	299	818	32.30	42.50	1.57	1.64	1.04	3.60

From: UNOET/SIO Data Bank

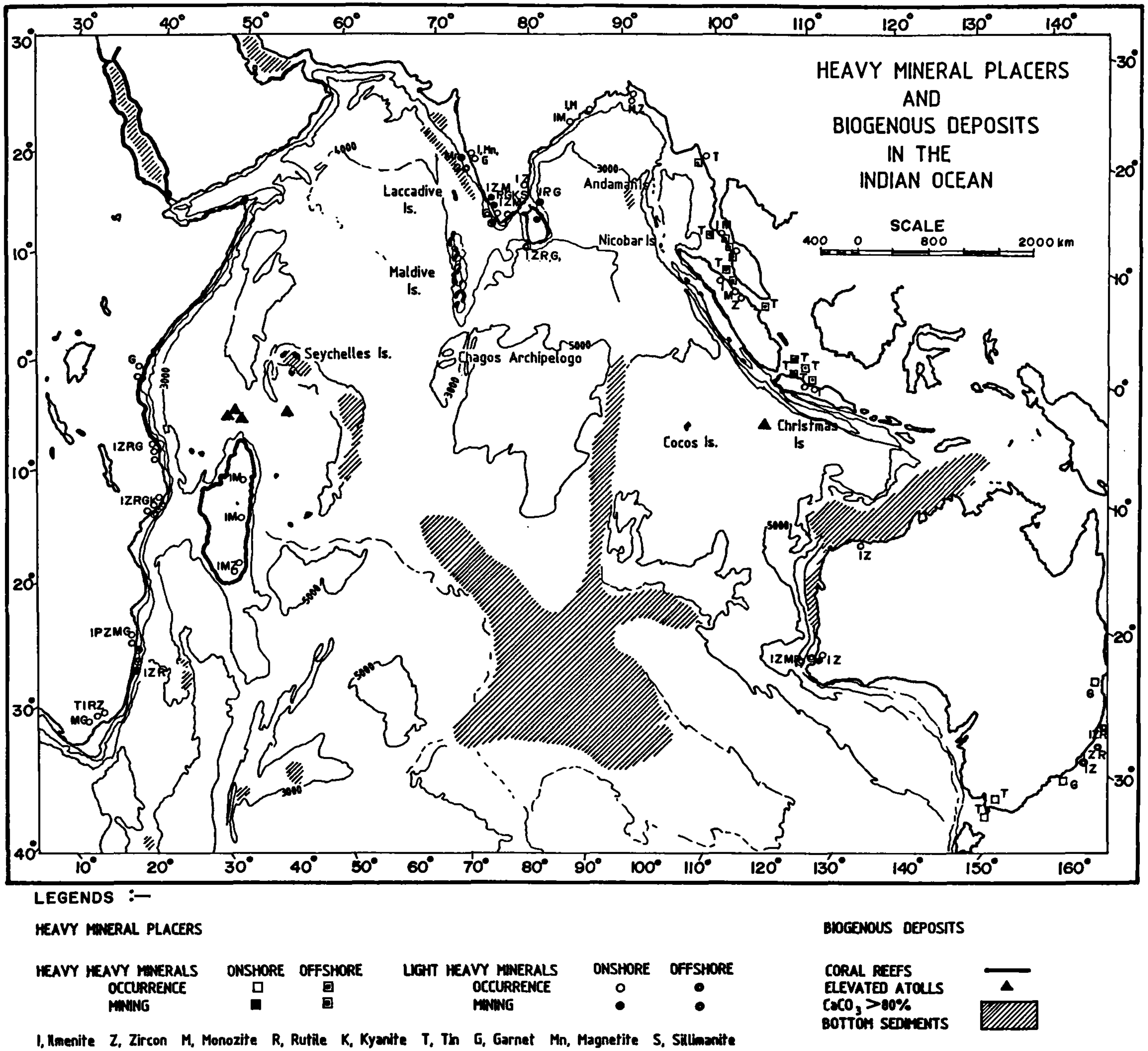

Figure 1. Heavy mineral placers and biogenous deposits in the Indian Ocean.

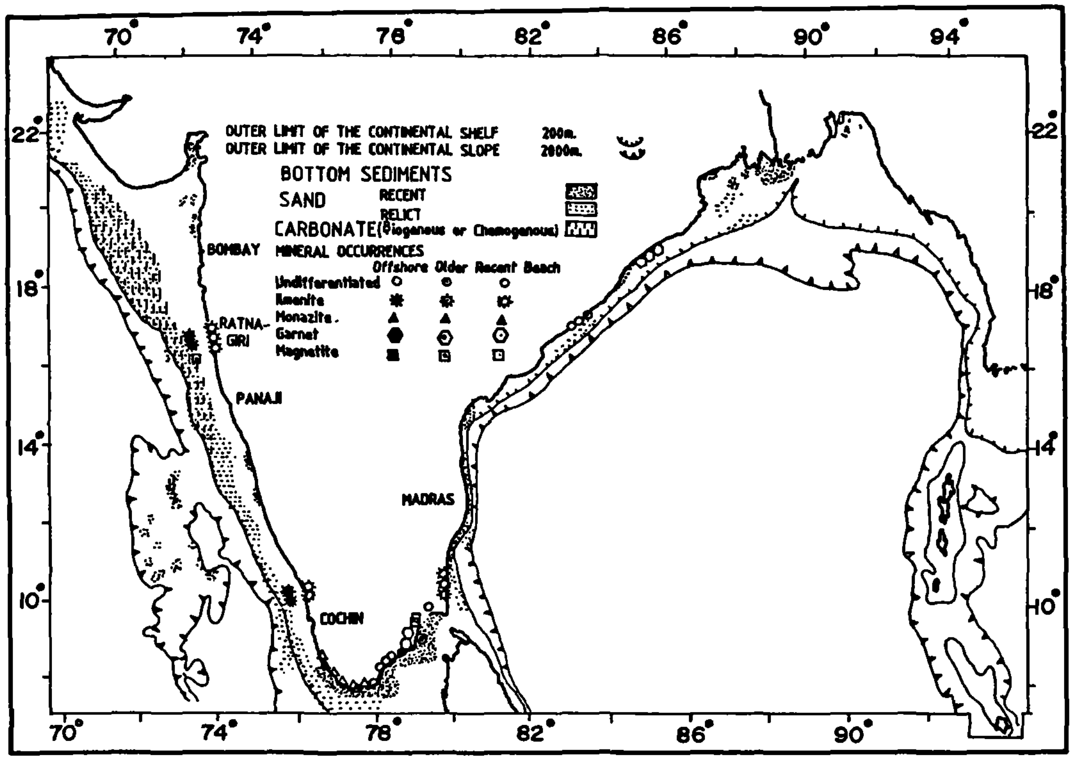

Figure 2. Map showing the terrigenous mineral occurrences on the continental margins of India.

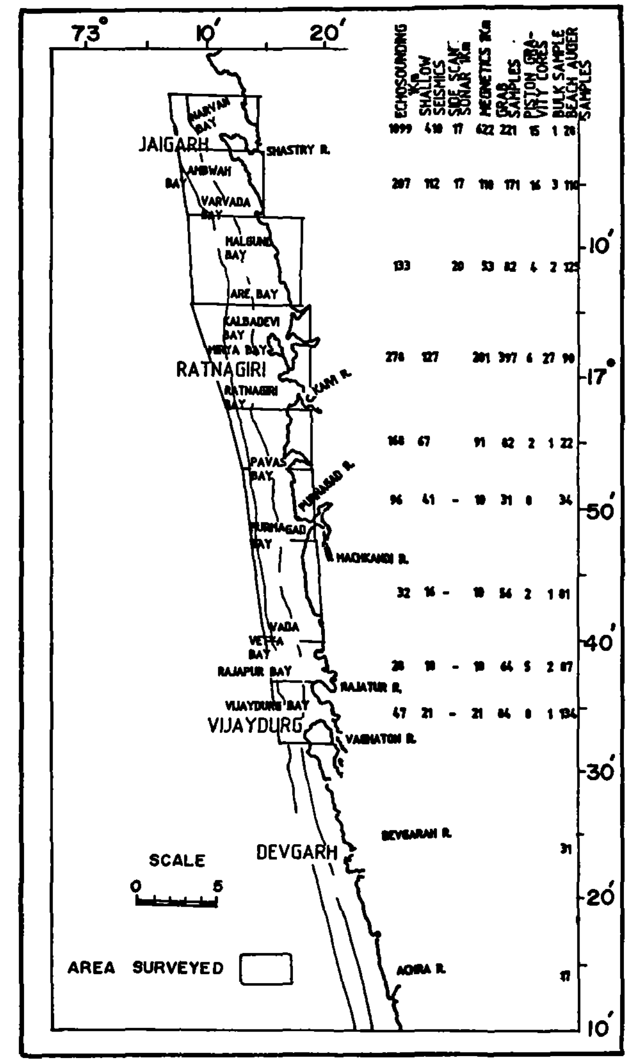

Figure 3. Map showing status of exploration for heavy mineral placers off the Konkan coast, Maharashtra, India (from Siddiquie et al., 1982).

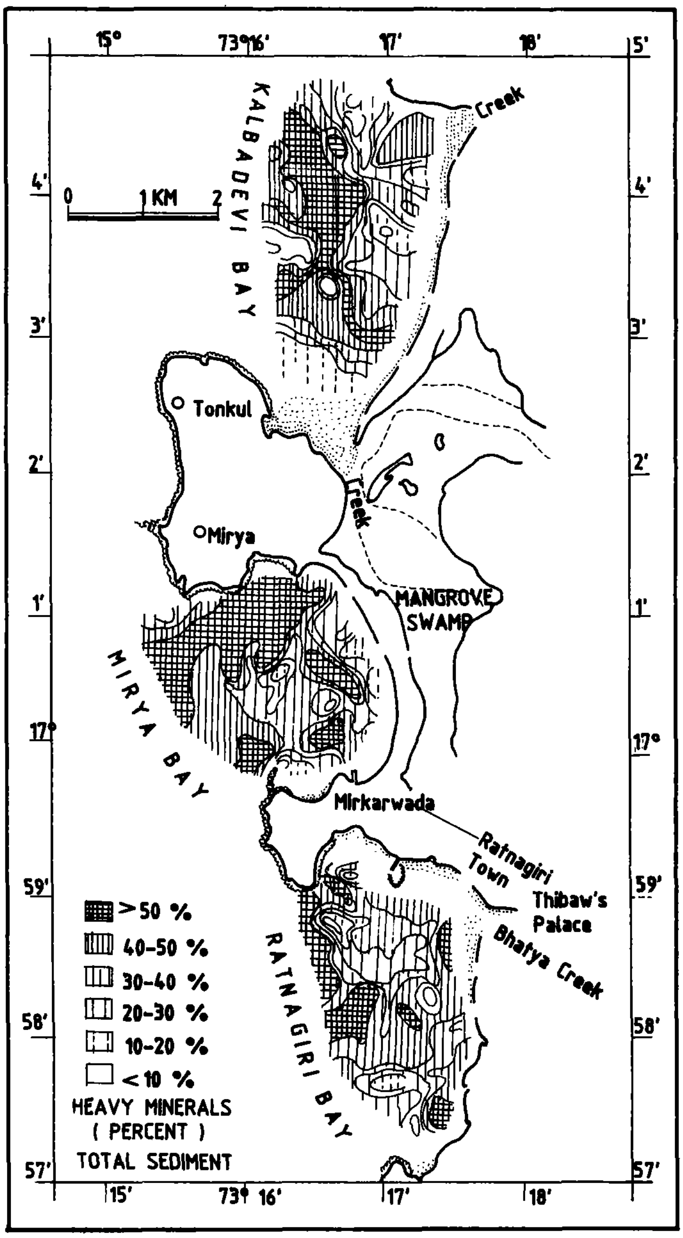

Figure 4. Map showing the distribution of heavy minerals in sediments of Kalbadevi, Mirya and Ratnagiri bays, Konkan coast, Maharashtra, India (from Siddiquie et al., 1979).

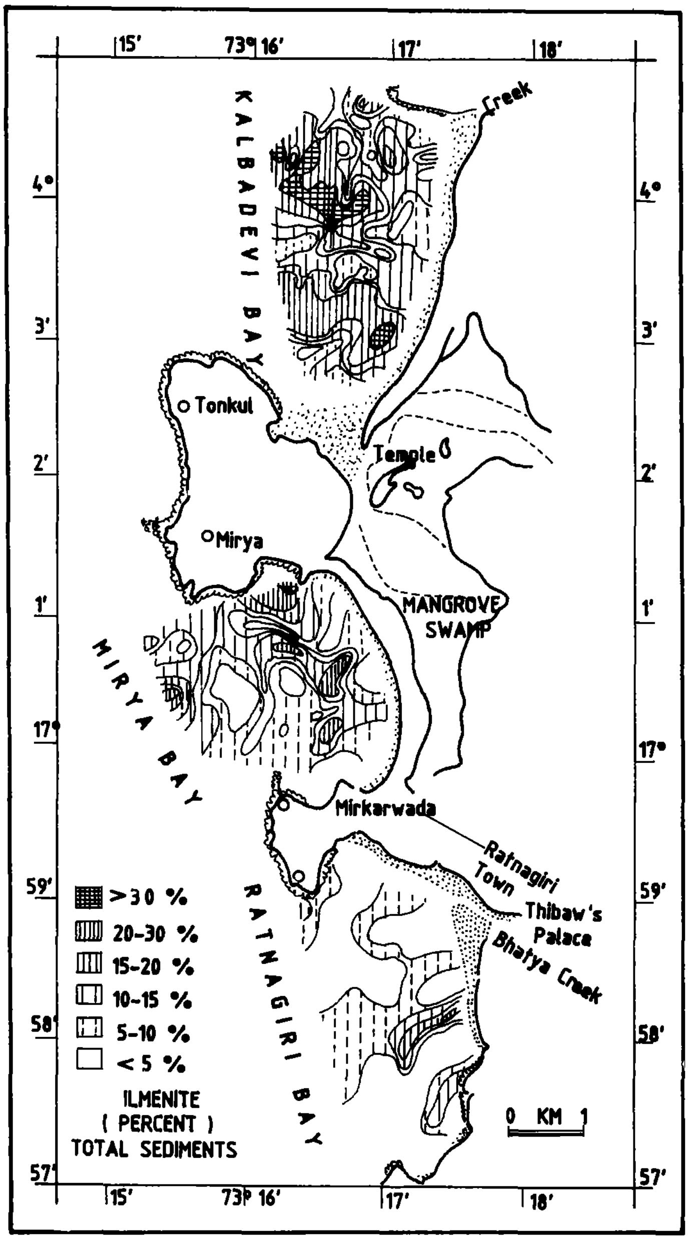

Figure 5. Map showing the distribution of ilmenite in sediments of Kalbadevi, Mirya and Ratnagiri bays, Konkan coast, Maharashtra, India (from Siddiquie et al., 1979).

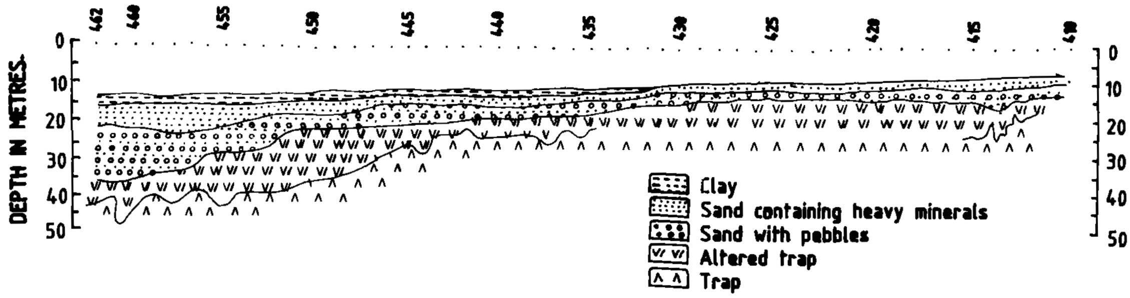

Figure 6. Interpreted seismic profile across the ilmenite placers off Jaigad, Konkan coast, Maharashtra, India.

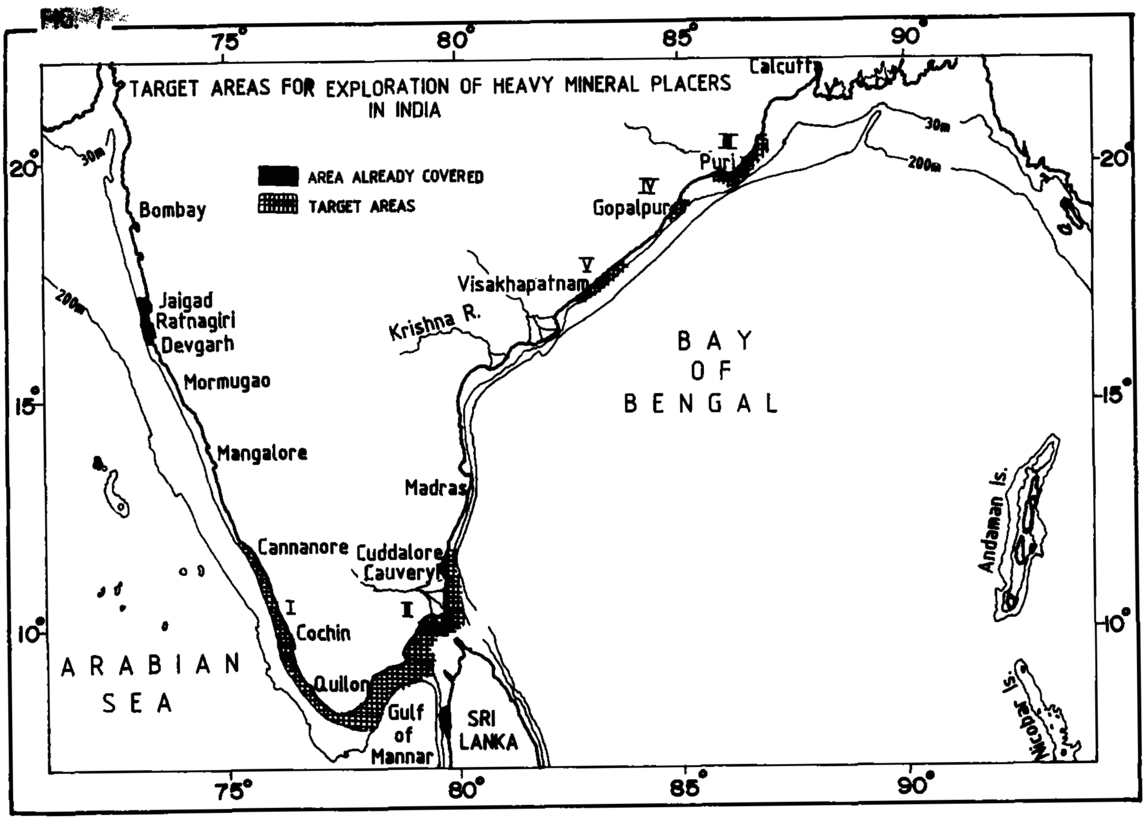

Figure 7. Target areas for exploration of heavy mineral placers in India.

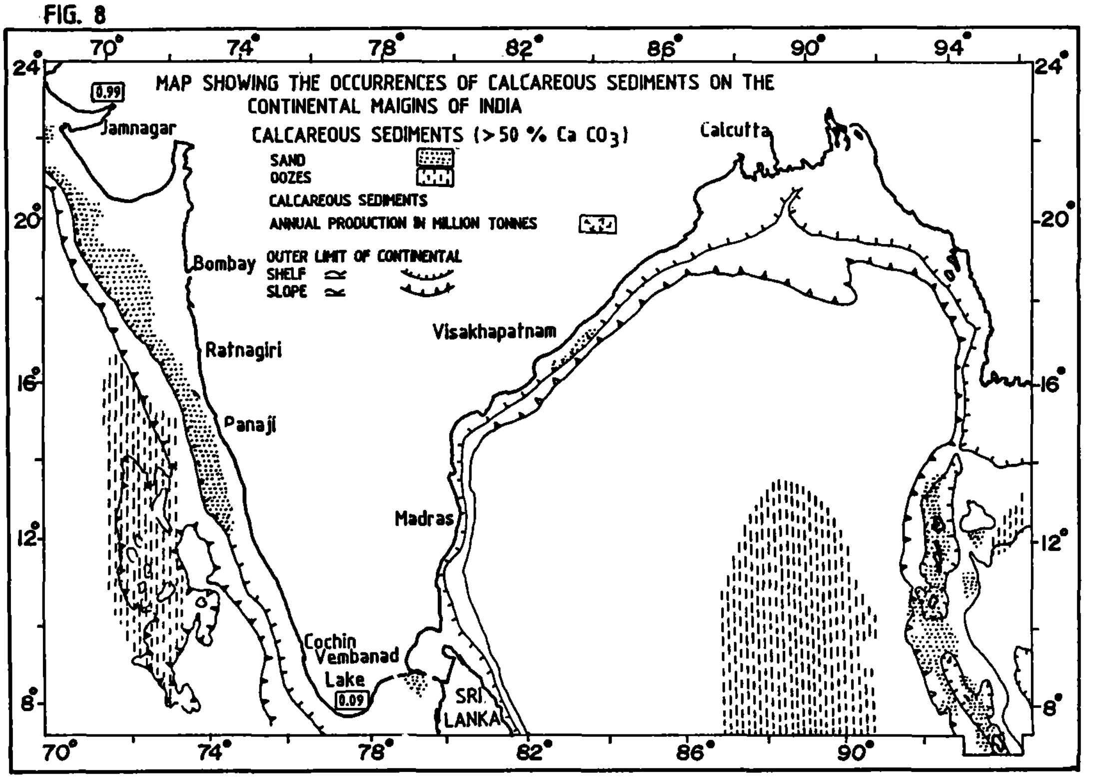

Figure 8. Map showing the occurrences of calcareous sediments on the continental margins of India.

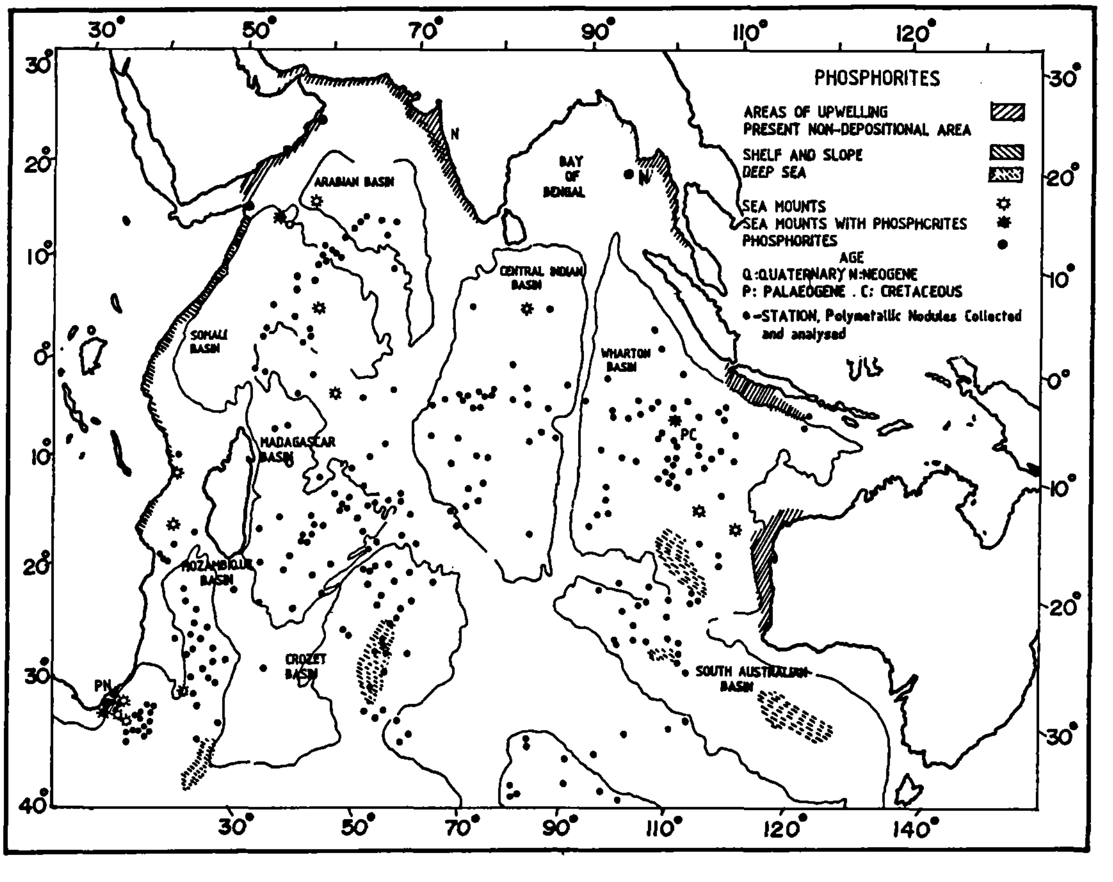

Figure 9. Map showing the phosphorites deposits and locations for polymetallic nodule samples in the Indian Ocean.

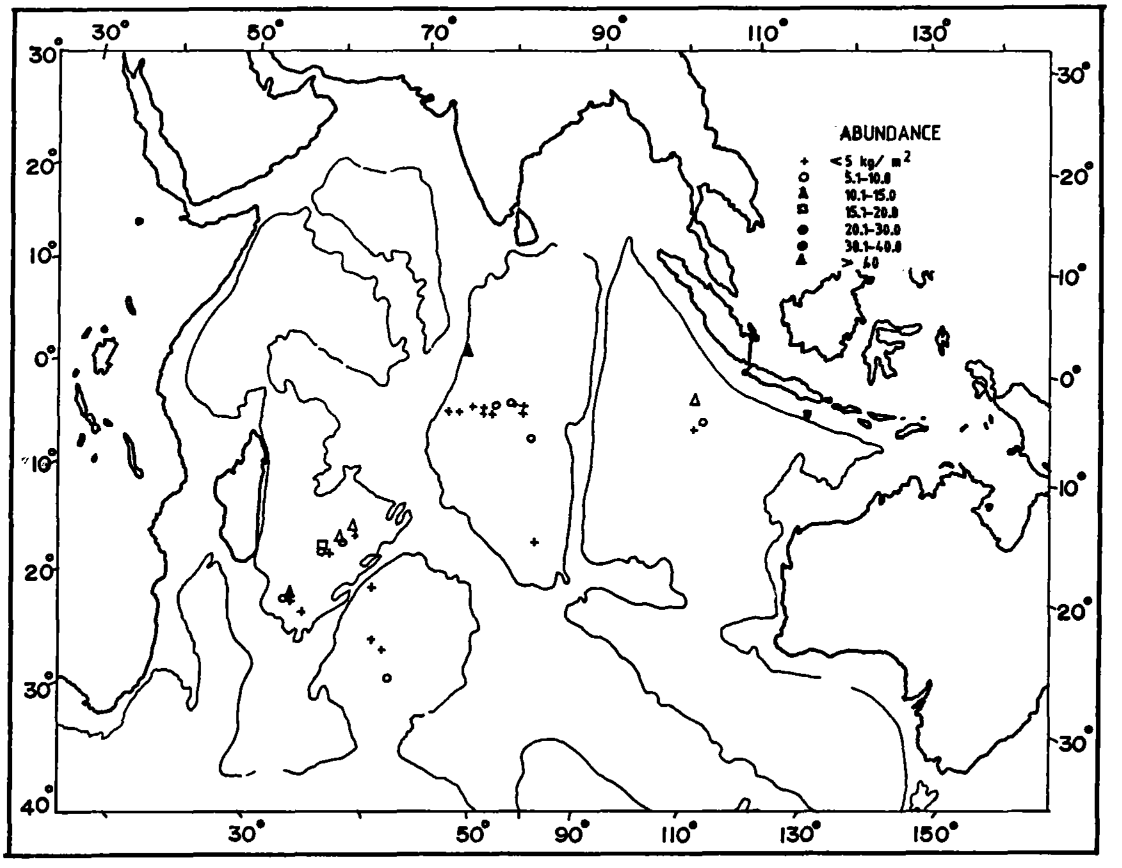

Figure 10. Map showing the abundance of polymetallic nodules.

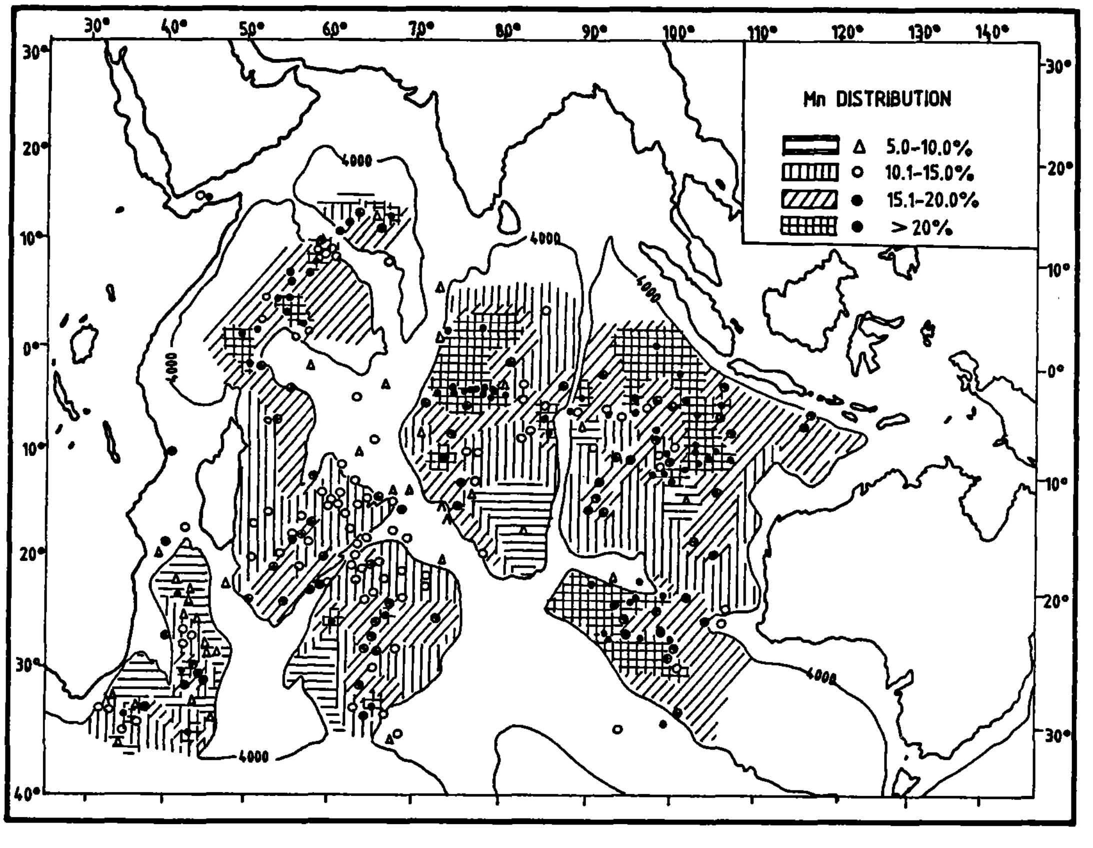

Figure 11. Map showing the distribution of Manganese.

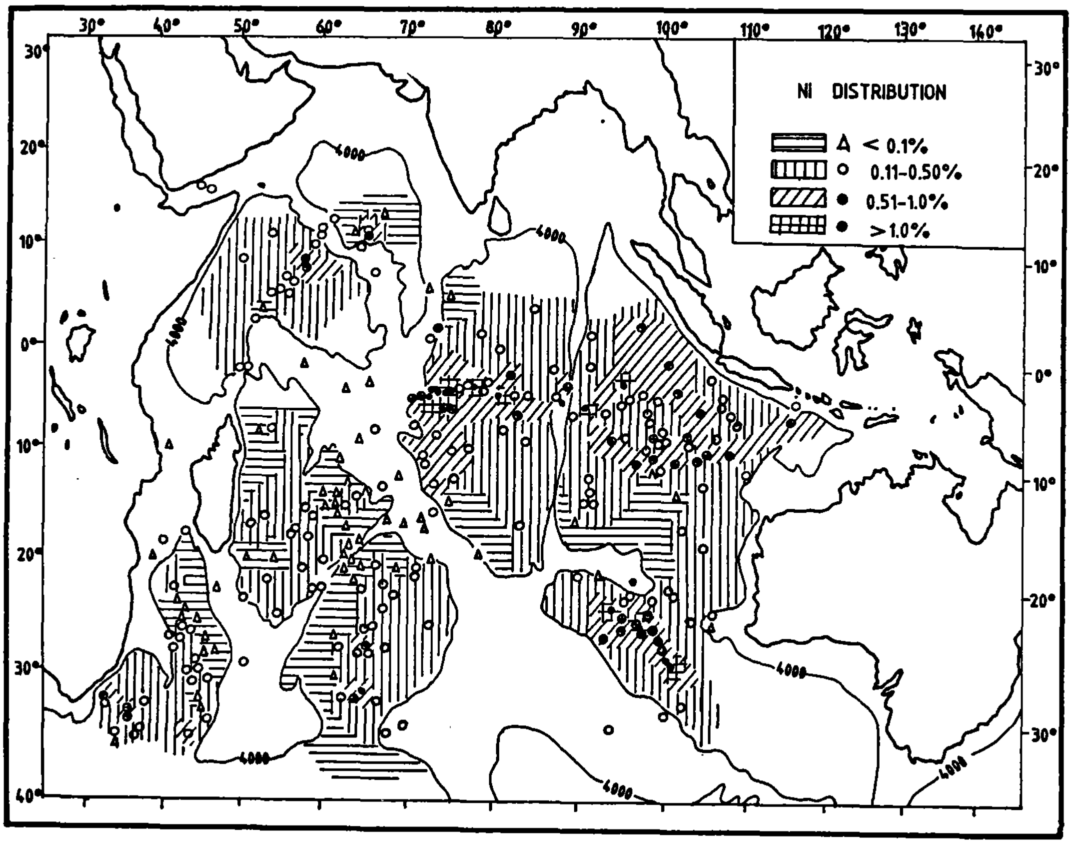

Figure 12. Map showing the distribution of nickel.

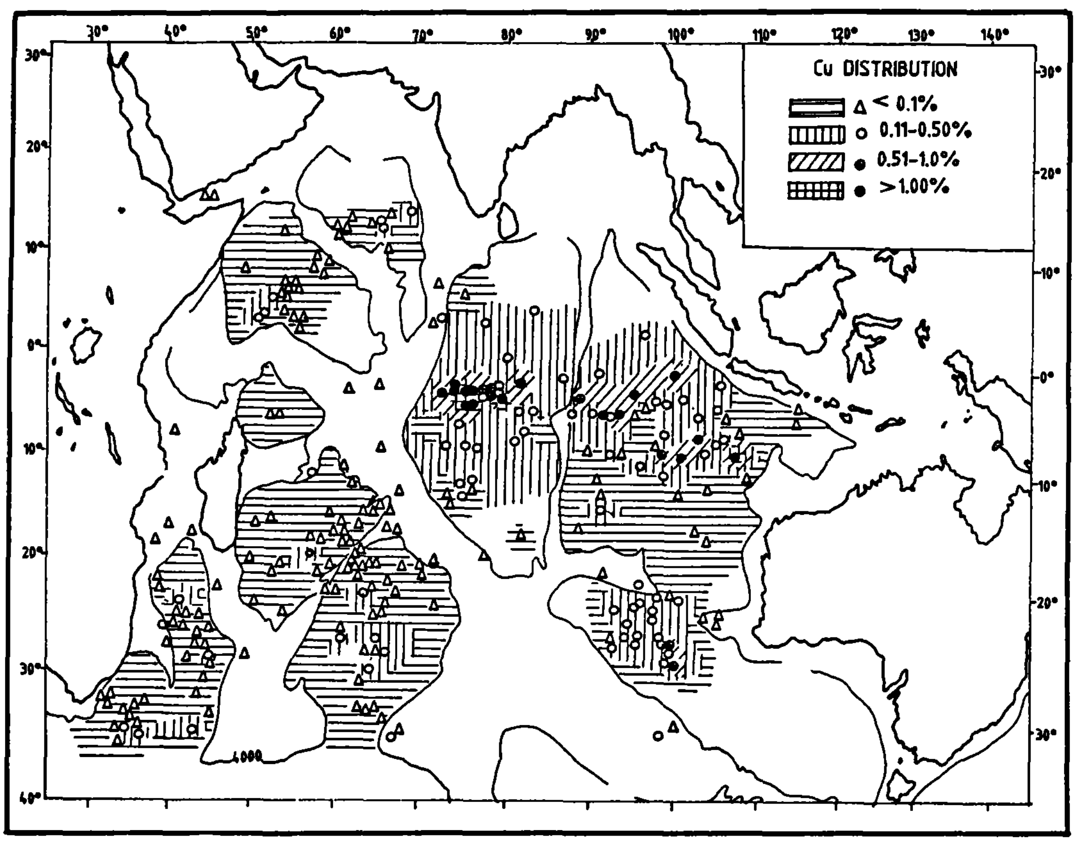

Figure 13. Map showing the distribution of copper.

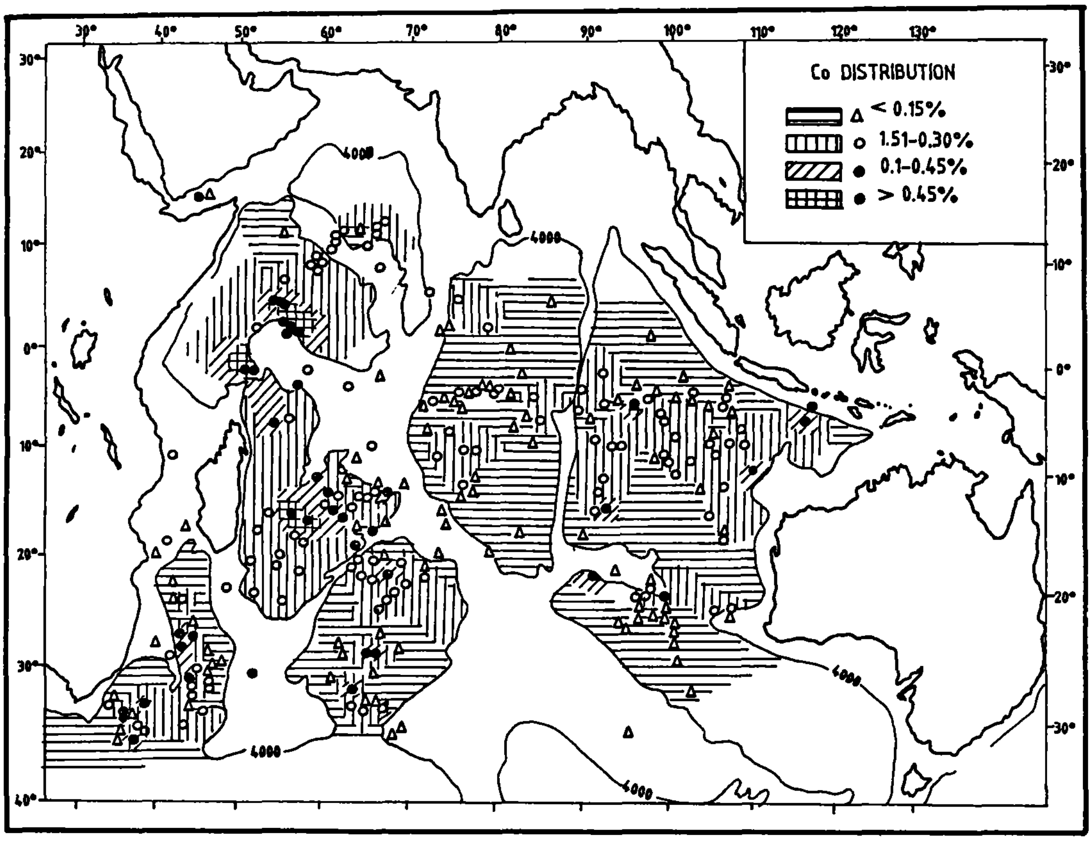

Figure 14. Map showing the distribution of cobalt.

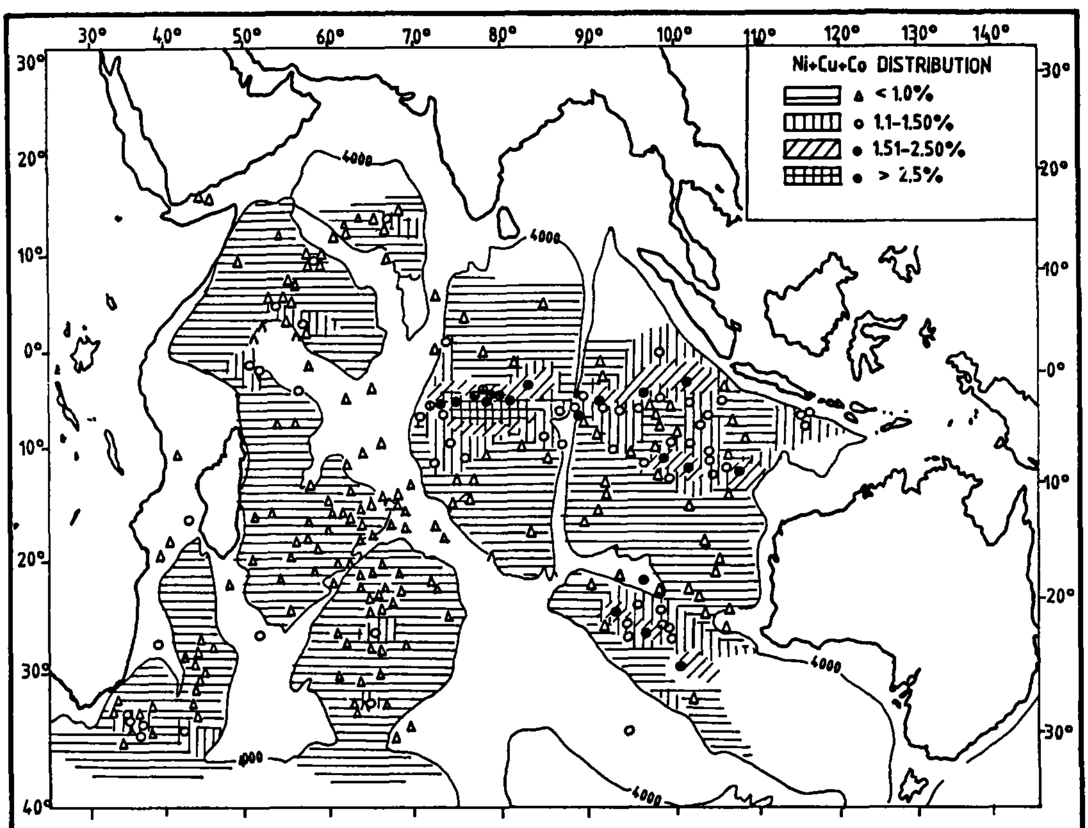

Figure 15. Map showing the distribution of nickel + copper + cobalt.

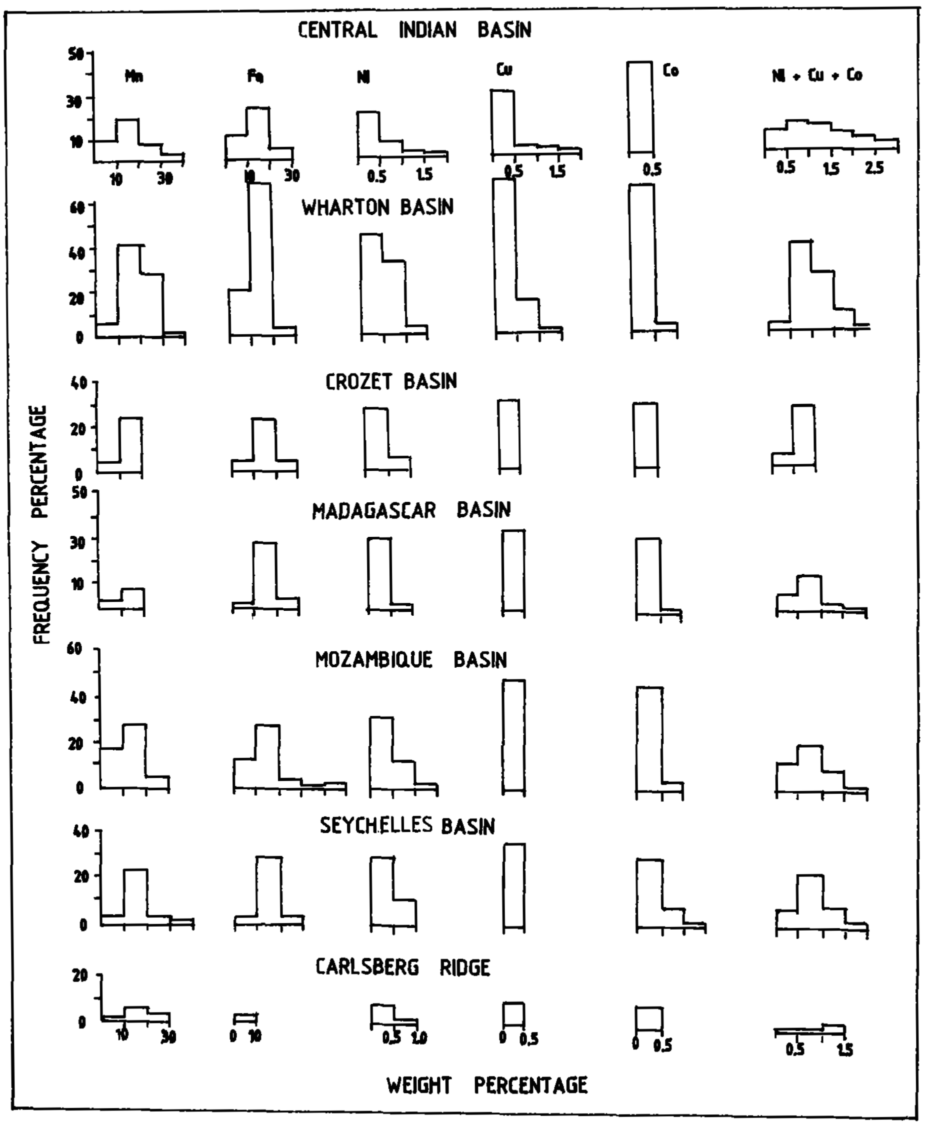

Figure 16. Frequency of manganese, nickel, copper, cobalt and grade (Ni+Cu+Co)

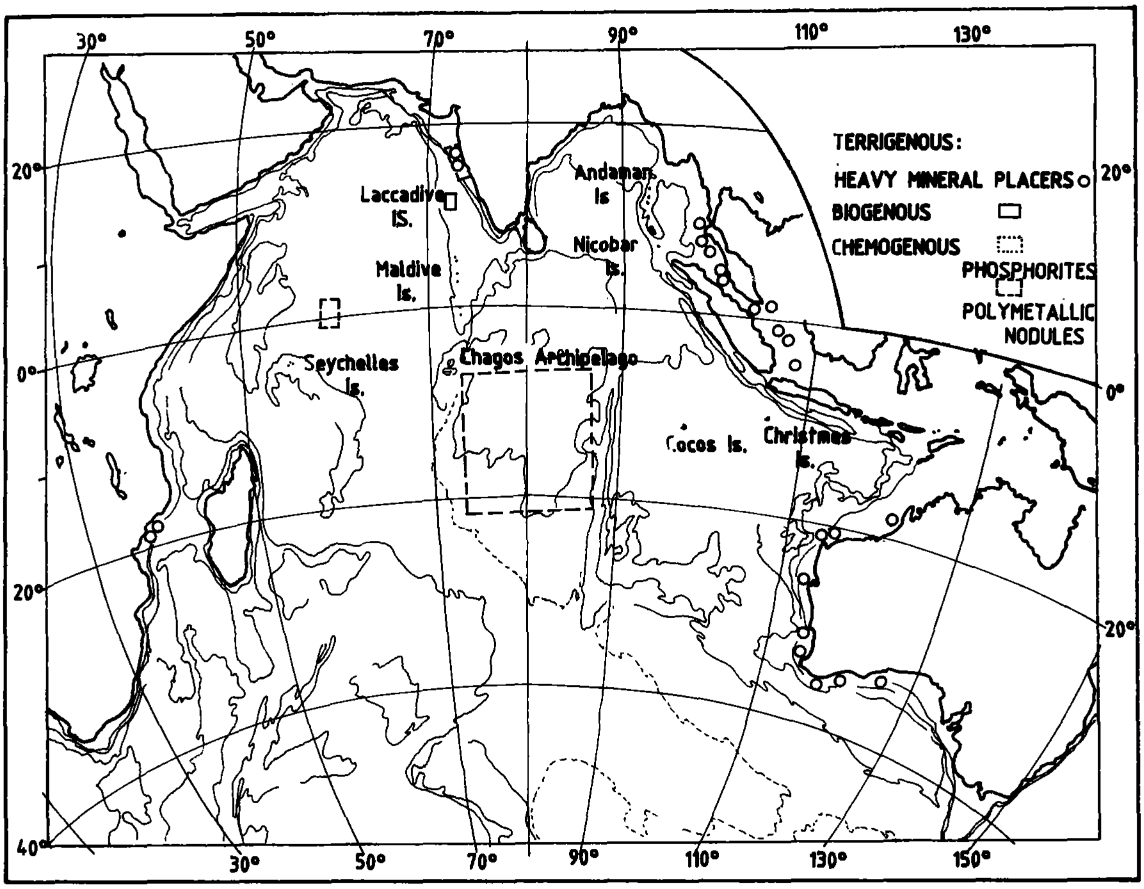

Figure 17. Marine mineral exploration/recent activities.

RESEARCH NEEDS FOR THE CORAL REEF ECOSYSTEMS OF THE CENTRAL INDIAN OCEAN

M.W.R.N. DE SILVA
National Aquatic Research Agency
Crow Island, Colombo 15
Sri Lanka

ABSTRACT

Coral reefs are unique tropical ecosystems that are important to the fisheries and tourist industries and are regarded as breeding and nursery grounds of many marine organisms. In addition, they act as barriers which prevent sea erosion. Unfortunately, in the Central Indian Ocean coral reefs have received neither the necessary attention nor respect. At present coral reefs in the Central Indian Ocean are over-exploited, damaged and subjected to destruction that raises grave doubts about their ability to survive in the future.

Most available information on the coral reefs of the Central Indian Ocean pertains to the taxonomy and biogeography of the reef biota. Information about the structure of reef communities and their physiographic zonations is limited to only a few reefs.

Major causes of coral reef degradation in the region are coral mining, use of explosives for fishing, natural sedimentation, tourist pressure and the removal of coral, reef fish and other reef organisms for export or sale as souvenirs. 'Crown of Thorns' starfish (*Acanthaster planci*) infestations have caused the destruction of some reef areas. Only a few countries in the Indian Ocean region have attempted to control the degradation of their coral reefs by the establishment of coral reef parks or reserves.

There is a need in the Central Indian Ocean and the adjacent regions for further surveying and mapping of coral reef areas as well as estimated economic productivity and potential, including the Maximum Substainable Yield (MSY) of exploitable coral reef resources. The causes and the rates of degradation of coral reefs must be determined and a concerted effort made to manage and conserve coral reefs in the area. Development of a central pool of Indian Ocean coral reef research and management expertise in collaboration with other relevant marine disciplines could provide the nucleus for much needed co-operative and comparative coral reef research in the region. Coral reef research data should be collected and disseminated; this can provide a regional basis for concerted efforts for managing and conserving coral reefs.

INTRODUCTION

For many South Asian coastal countries, coral reefs traditionally have provided food, building materials, pearls and shells for jewelery, as well as sheltered harbours for boats. In addition they act as buffers which prevent sea erosion. The potential of several coral reef organisms as producers of medicinally important compounds has received the close attention of scientists in recent years (e.g. Angeles, 1981).

For generations these reefs were able to provide for the needs of the local communities that were dependent on them. But today the coral reef resources are being exploited increasingly for export to cities and other countries. Because of these increased pressures, reefs have been destroyed or badly degraded in many areas. It is evident that immediate steps are required to stem the degradation of coral reefs for the benefit of present and future generations.

The state of equilibrium of a coral reef ecosystem hinges around the delicately balanced interactions within and between biotic and abiotic components. Critical disturbance, even to a single vital parameter, can cause an imbalance leading to the destruction of the entire community. JOHANNES (1975), commenting on the importance of corals to the reef ecosystem, stated that "So central are corals to the integrity of the reef community that, when they are selectively killed, migration or death of much of the other reef fauna ensues... therefore, the environmental tolerance of the reef community as a whole cannot exceed those of its corals".

UTILIZATION OF CORAL REEFS

EXTRACTIVE USES

Traditionally, many coastal communities dependant and/or associated with coral reefs have harvested the productivity of the reefs. Food in the form of fish, molluscs, crustaceans and seaweeds, ornaments and materials for hunting and fishing, building materials, fertilizers, and more recently aquarium fishes and invertebrates, as well as pharmaceutical chemicals, have been obtained from coral reefs. So long as Maximum Sustainable Yields are not exceeded, it is generally possible for a coral reef to cope with the demand for these products and recover in an ecological sense. However, the health of a reef is also crucial for the continued extraction of these resources.

NON-EXTRACTIVE USES

The most obvious asset of coral reefs is their fascinating beauty. The innumerable colourful fishes, shells, fan-worms and other organisms give tropical seas a romantic image which has been enhanced through the growing popularity of skin and scuba diving and exposure through the visual media such as movies and television (BETTERTON and DE SILVA, 1979). Maldive Islands and Sri Lanka have started to exploit the natural beauty of their coral reefs and associated beaches by establishing tourist resorts around such areas. Although local residents have been slower than tourists to use coral reefs for recreation, they have benefited by the opportunities for employment and other means of socio-economic improvement created by tourism.

Coral reefs represent a uniquely productive tropical ecosystem (MATHIAS and LANGHAM, 1975; DAHL, 1974; SOROKIN, 1973) that is important to fisheries (SMITH, 1978; STEVENSON and MARSHALL, 1974; ALACALA and LUCHAVEZ, 1981) and tourist industries, as well as providing breeding and nursery grounds of several marine organisms (summarised by DE SILVA, 1983). A coral reef also provide a natural storehouse of species for scientific research and investigation. In addition, they can be used as instructional tools in both formal and non-formal education. Coral reefs are natural laboratories which provide the opportunity for observing living examples of principles established in the classroom.

Establishment of marine parks can be a major non-extractive use of a coral reef. In addition to providing a means to monitor of a reef, a properly organized and managed marine park can generate an income and provide opportunities for socio-economic improvement by offering several supporting and recreational facilities such as scuba diving, snorkelling, tours in glass-bottom boats and similar water-based activities.

ENVIRONMENTAL PROBLEMS OF CORAL REEFS

Factors causing the degradation of coral reefs in South Asia can be considered under two separate categories: natural causes and man-made causes. As stated by GOMEZ (1981), "... in practical situations it is sometimes impossible to separate the effects of the two, because they are intertwined in a complicated and dynamic way with the multitude of factors and processes that characterize the exceeding complexity of coral reefs as ecosystems".

NATURAL CAUSES OF REEF DEGRADATION

A coral reef is subject to a variety of natural environmental stresses primarily associated with water movement, geological dynamics and biological interactions. Destructive wave action and strong currents experienced during monsoons, cyclones and freak storms can cause severe mechanical damage to coral reefs. PILLAI (1971a), reporting on the effect of a severe cyclone that swept through southeastern India and eastern Sri Lanka in December 1964, stated that "... the branching corals, especially *Acropora* spp. were the worst hit and very many large colonies of *A. corymbosa* and *A. surculosa* were found uprooted and washed ashore. Several corals, both ramose and massive, were later found dead on the reefs. The alcyonarians were reduced greatly in numbers at several places. On the whole there was a reduction of corals in most of the reefs". Currents also can bring about changes in water quality such as temperature, salinity, turbidity and nutrients which affect coral development. Exposure to anomalously low tides can expose reefs to excessive heat, freshwater inundation or destructive wave action.

Various geological phenomena also are known to cause reef damage. Sedimentation causes severe reef damage by inhibiting coral growth through reduced incident light available to the zooxanthellae within coral polyps, by interfering with the planktonic food supply, or by smothering the coral polyps. PILLAI (1971a) reported the mortality of a large number of coral colonies in the inshore waters of Palk Bay due to silt and suspended sand stirred up by monsoon winds. Large-scale destruction of corals due to silt stirred up during heavy winds also has been reported in Mandapam in the Gulf of Mannar. Other geological phenomena such as landslides, earthquakes, volcanic activity, and tectonic uplift and subsidence also can cause coral reef damage.

Among the biological interactions, competition constitutes a major limitation to the growth of reef-building organisms. Boring organisms, such as barnacles, molluscs, sipunculids, polychaetes, sponges and algae as well as grazing organisms such as parrot fishes (Scaridae), are also known to cause some degree of damage to coral reefs. Predation is another factor of great importance, the most well-known predator of coral being the 'Crown of Thorns' starfish (*Acanthaster planci*) which has been responsible for the complete destruction of several reef areas along the east coast of Sri Lanka (DE BRUIN, 1972).

HUMAN ACTIVITIES CAUSING REEF DEGRADATION

Human activities are of greater importance in several places than natural causes in causing coral reef degradation and have been responsible for extensive damage to many coral reef areas in South Asia. In 1980, the Coral Reef Working Group - 1 of the International Union for Conservation of Nature and Natural Resources (IUCN) recognized 24 human-related activities that could cause damage to coral reefs. Of these, the activities considered relevant to South Asia are listed in Table 1 on a three point scale to indicate the extent of the problem.

CORAL REEFS OF THE CENTRAL INDIAN OCEAN

SEWELL (1932, 1935), STODDART (1971, 1972), PILLAI (1972), DE SILVA (1983) SCHEER (1972, 1984), MAHADEVAN and NAYAR (1972), MERGNER and SCHEER (1974), and PILLAI and SCHEER (1976) have summarized the information available on corals and coral reefs of the Indian Ocean, with special emphasis on the Central Indian Ocean. Perhaps with the exception of SEWELL, most of the early workers "... were less concerned with the structure of reef communities and their physiographic expression than with the taxonomy and biogeography of the reef biota" (STODDART, 1971). However, many recent workers have made in-roads to bridge this gap in our knowledge of coral reefs (STODDART, 1971, 1972; MERGNER and SCHEER, 1974; DAVIES et al., 1971; and SCHEER, 1971).

Very little information is available on the utilization and cause of damage to the coral reefs in the Central Indian Ocean. Other than the Hikkaduwa Marine Sanctuary on the southwestern coast of Sri Lanka, which was declared in 1979, and the recently declared Pirotan Island Marine Park in the Gulf of Kutch in India, no other marine sanctuaries, reserves or parks exist in the Central Indian Ocean. And even the Sri Lanka sanctuary has experienced difficulty in implementation of regulations, mainly because of socio-economic considerations. Currently, moves are underway to rectify some of the problems.

The known distribution of the major coral reefs in the Central Indian Ocean are given in Figure 1, and various areas are discussed in the following paragraphs.

MALDIVE ISLANDS

The Maldive Archipelago, situated southwest of India, consists of 22 atolls (PILLAI and SCHEER, 1976) with about 1800 small islands, extending 764 km in a north-south direction. The northern limit is Ihavandiffulu Atoll (7°05'N, 72°55'E), while Addu Atoll — one of the most scientifically studied atolls — is located at the southern tip (0°40'S, 73°10'E). The atolls of Maldives are distinguished by "faros" or small atoll-shaped reefs round their rims. Microatolls and patch reefs/knolls are also well documented in the Maldives. Although most of the Maldivian coral reefs are yet to be investigated, they are still the best-known coral reefs in the Central Indian Ocean.

Investigations on the coral reefs of Maldives date back to Moresby's expedition to the atolls during 1834 to 1836. However, these reefs were first studied in detail during Stanley Gardiner's expedition to the islands during 1899-1900 (PILLAI and SCHEER, 1976). The results of his expedition were published in two volumes (GARDINER, 1903-1906) containing elaborate discussions on the reef formations of Maldives and Laccadives, and systematic analysis of the zoological and botanical specimens collected. Gardiner was followed by Alexander Agassiz, who spent two months cruising through the archipelago in 1901-1902 (AGASSIZ, 1903). However, his work is considered less scientifically valuable than that of Gardiner (STODDART, 1966).

DERANIYAGALA (1956) made a collection of specimens from the Maldivian reefs in 1932 for the Colombo Museum. This was followed by John Murray Expedition, led by R.B. Seymour Sewell (1933-1934), which conducted gravity surveys and deep soundings in the Maldives and published physiographic descriptions of Addu and Goifurfehendu Atolls (SEWELL, 1936).

Although several subsequent expeditions visited the Maldives, no major contributions were made to the knowledge of the corals and coral reefs of these islands until the 'Xarifa' Expedition led by Hans Hass in 1957. Based on the work done in Addu Atoll during this expedition, HASS (1962, 1965) formulated a new hypothesis on atoll formation. PILLAI and SCHEER (1976) described and summarized the geographical distribution of 143 species of hermatypic corals belonging to 49 genera and 4 ahermatypic corals belonging to 3 genera collected during this expedition. They also listed a total of 241 species of corals belonging to 75 genera from the Maldives together with their known occurrence at the various atolls.

A subsequent major study the reefs of the Maldives was made in 1964 by a team of scientists from the University of Cambridge led by D.R. Stoddart. STODDART (1966) studied the geomorphology of Addu Atoll and zonation of corals (STODDART et al., 1966; DAVIES et al., 1971; STODDART, 1972), coral fauna (WELLS and DAVIES, 1966; CLARK and DAVIES, 1966) and other aspects of the Maldives. PURDY (1981) discussed the possible evolution of the Maldive Atolls based on a combination of seismic stratigraphy and the results of a single well.

The Maldive reefs have a high diversity of coral species, with a total of 241 species of hard corals belonging to 75 genera (PILLAI and SCHEER, 1976). Available information indicates that many of the reefs studied in the Maldives are in a fairly good condition, with a live coral cover of between 70-80 percent for Addu Atoll, 40-70 percent for Rasdu Atoll, 60-90 percent for Fadiffolu Atoll and 20-50 percent for Ari Atoll (SCHEER, 1974). There is no mention of any human or other factors contributing to the deterioration of the Maldivian reefs. Although coral rock is used in some islands as building material and some reefs have traditionally been exploited for fish and shells, no impact of these activities on the coral reefs has been recorded at these islands. Sporadic use of explosives to catch fish is apparently carried out in some islands.

The coral reef formations in India are restricted to the atolls of the Laccadive Archipelago, and to the fringing and patchy reefs located in the Gulf of Kutch along the northwestern coast, the Palk Bay and Gulf of Mannar on the southeastern coast and the Andaman and Nicobar islands in the Bay of Bengal. SEWELL (1932) related the absence of coral reefs from a major part of the Bay of Bengal and the subcontinental coasts of India to the large influx of freshwater and silt from the major rivers of the area. Further, SEWELL (1935) suggested that the presence of high levels of nitric acid in the Ganges River, especially during the monsoon season, could discourage the growth of corals in the Bay of Bengal. More recently, PILLAI (1967, 1969a, 1969b, 1971a, 1971b, 1971c, 1972) made a valuable contribution to the knowledge of corals around India; there appears to be no work on the Scleractinia from the Gulf of Kutch" (PILLAI, 1971a).

MAHADEVAN and NAYAR (1972), as well as VENKATARAMANUJAM et al. (1981), discussed the exploitation and utilization of corals in the Gulf of Mannar and Palk Bay for making lime as building material and for road construction. RAJENDRAN and DAVID (1972) estimated the extent of coral reefs in and around some of the islands in the Gulf of Mannar, while MERGNER and SCHEER (1974) described the physiographic zonation and ecological conditions of some coral reefs in southern India.

Gulf of Mannar and Palk Bay

The coral reefs of this area are of the fringing or patchy type, thriving in shallow waters either near the shores of the mainland or encircling a few islands. PILLAI (1971a), referring to the reefs of the area, stated that "... shallow waters with muddy or sandy surroundings considerably restrict the growth of corals and at present the reefs seem to be in a state of deterioration".

Running almost parallel to the coast of South India in the Gulf Mannar are 21 islands of limited extent lying between 8°47'N — 9°15'N latitude and 78°12'E — 79°14'E longitude. Many of these islands have well-developed fringing reefs of varying sizes located 100 to 150 m from the shoreline. In addition to these islands, coral reefs occur around Rameswaram Island, the largest island in this group. One reef starts from north-northeast shore of Rameswaram, runs around Devil's Point parallel to the shore and extends up to Mandapam in the Palk Bay (MAHADEVAN and NAYAR, 1972). This reef, which is located 300-600 m offshore, has a 1-2 m deep (at low tide) sandy lagoon which is generally devoid of any coral (PILLAI, 1971a). Most of the recent investigations on the coral reefs of Palk Bay and Gulf of Mannar have been carried out on these reefs.

PILLAI (1972) recorded 110 hermatypic coral species belonging to 25 genera and 7 ahermatypic corals belonging to 7 genera from Mandapam area. From Tuticorin in the Gulf of Mannar, he recorded 16 hermatypic coral species belonging to 14 genera. All these species have been collected from shallow waters not deeper than 2 m at low tide. The deeper waters are yet to be investigated.

The Gulf of Mannar appears to be a unique zone with regard to the variety of fishing activities. MAHADEVAN and NAYAR (1972) stated that "... the nature of the sea-bottom in the inshore area of this zone supports certain fishing occupations which are not met within most other zones of the Indian coast. These are the chank fishing, pearl oyster fishing and coral quarrying. The fishermen find a steady income in bringing coral stones or coral blocks broken by them from the shallow areas adjoining many small islands between Tuticorin in the south and Pamban at the head of the Gulf". The Gulf fishermen also exploit the coral reefs from Palk Bay. Of the many coral species that occur in the area, only a few species such as *Acropora formosa, Porites compressa, P. solida, P. somaliensis, Favia valenciennesii* and *Tubipora* spp. are exploited for commercial purposes (PILLAI, 1967). *A. formosa* fragments, popularly called 'challi', are collected extensively for manufacturing lime, while other corals such as the massive *Porites* sp. and *Favia valenciennesii* are collected for other purposes. VENKATARAMANUJAM et al. (1981) estimated that at Tuticorin alone some 30,000 cubic meters (approximately 15,000 tons) of coral stones are landed every year. PILLAI (1971a) indicated that coral quarrying has resulted in large scale destruction of these reefs and their associated fauna. He stated further, that "... some of the islands in the Gulf of Mannar are almost completely exploited, leaving little trace of their existence". Explosives are also used for catching fish in some areas. PILLAI (*op cit*) also reported large scale destruction of corals at Mandapam in the Gulf of Mannar due to siltation.

Laccadive Archipelago

This archipelago is composed of 14 atolls with two sand cays, Cherbaniani Reef and Beleapani Reef (12°18'N, 71°54'E) to the north and Minicoy Atoll (8°30'N, 73°11'E) to the south. Our knowledge of the corals and coral reefs of Laccadives is extremely limited. The reefs were first studied by OLDHAM (1895) and GARDINER (1903, 1904, 1905). More recently, PILLAI (1971a) worked on a collection of corals from Minicoy Atoll and Chetlat Island (72°40'E, 11°41'N), and the productivity of the reef in Minicoy Atoll was measured by NAIR and PILLAI (1972).

A total of 69 species of hermatypic corals belonging to 26 genera have been recorded from the Laccadives, all of which are known to occur in Minicoy Atoll. Twelve hermatypic corals belonging to 10 genera were reported from Chetlat Island; this list, however, is incomplete since no intensive collections have been made. PILLAI (1971a) suggested that the huge colonies of *Diploastrea*, which are fairly common at present but was considered to be rare by GARDINER (1904), have grown in the last 70 years.

Andaman and Nicobar Islands

Little is known about the corals and distribution of coral communities of the Andaman and Nicobar Islands. Only fringing and patch reefs are known to occur in these islands, but very little information is available on the community structure and quality of the reefs. Whatever knowledge we have comes from the early work of ALCOCK (1893, 1898) on the deep sea corals around these islands, SEWELL's (1935) work during the John Murray Expedition (1933-1934), SCHEER's (1971) work on the Nicobars during the second 'Xarifa' Expedition (1957-1958), and MATTAI's (1924) and PILLAI's (1972) studies of corals from the Andaman Island.

Fifty-seven species of hermatypic corals from 23 genera and 11 species of ahermatypic corals belonging to 8 genera have been reported from the Andaman Island. SCHEER (1971) listed a total of 40 genera of hard corals for the Nicobar Islands. ROSEN (1971) reported that 41 genera of hermatypic and one genus of ahermatypic corals probably occur in the Nicobar Islands.

SRI LANKA

Although fringing coral reefs dominate much of the coastline of Sri Lanka, the better known ones occur along the northeastern coast near Trincomalee, and the southwestern coast near Dondra Head and Hikkaduwa. Information on the Sri Lankan corals and coral communities are almost non-existant except for the early work by RIDLEY (1883), ORTMANN (1889) and BOURNE (1905) who made extensive collections of corals from Sri Lanka (then known as Ceylon). Recently MERGNER and SCHEER (1974) studied the zonations and ecological conditions of some fringing reefs at Hikkaduwa, on the southwest coast of Sri Lanka. Seventy species of hermatypic corals belonging to 27 genera and 20 ahermatypic corals belonging to 12 genera have been recorded for Sri Lanka (PILLAI, 1972).

Many workers (DE BRUIN, 1972; DE SILVA, 1981) have reported on the destruction caused to several coral reef areas in Sri Lanka due to the use of explosives to catch fish, mining of coral for making lime, catching colourful reef fishes for export, and the feeding activities of the 'Crown of Thorns' starfish (*Acanthaster planci*), a notorious predator of corals. Although no quantitative estimates are available, most known coral reef areas around Sri Lanka have deteriorated rapidly during the past two decades. Increasing tourism also is taking its toll on some reef areas such as Hikkaduwa. Existing legislation (the Crown Lands Act, the Fauna and Flora Protection Ordinance, and the Fisheries Ordinance) could prevent the degradation of coral resources, but enforcement and political considerations have been largely ineffective.

BANGLADESH

The only known coral reef areas in Bangladesh occur around St. Martin's Island (20°31'N and 92°18'E) in the Bay of Bengal. FATTAH (1979) stated that "It is believed that a submerged reef is present on the south and southeast of St. Martin's Island and in all probability this reef is the western extension of the one of Malaysian sea coast. This area appears to be rich in biological zones, but adequate survey work is yet to be carried out on this region". He further stated that "... over the last 20 years in the author's experience, the damage to the reef surrounding the island has increased yearly at an alarming rate. It is known that a portion of the area has now been totally destroyed".

There is no information on the types of corals or their distribution at St. Martin's Island, but FATTAH (1979) summarized some of the human activities responsible for their deterioration, particularly the commercial collection of shells and corals. He also indicated that commercial collection of aquarium fish, recreational pressures, and use of explosives to catch fish could become future problems if adequate precautionary measures are not taken.

RESEARCH NEEDS

Traditionally coral reef resources have been exploited to cater to the needs of small communities. The methods used for such exploitation were largely non-destructive and highly selective. The Maximum Sustainable Yield (MSY) was thus not exceeded and the reef resources in most instances were able to replenish themselves. Even in instances where the MSY was exceeded alternative sites could be used, thus allowing for the regeneration of the over-harvested area. Today, in addition to the threat of overharvesting of coral reef resources, the spectre of pollution, excessive recreational and other human pressures pose real threats to coral reefs of the Central Indian Ocean.

To counter these pressures there is an immediate need for the following:

1) Further surveying and mapping of coral reef areas;
2) Assessing coral reef quality in terms of species diversity, percentage live/dead coral cover and the economic value as indicated by its exploitation for fisheries, tourism, etc.;
3) Determining the causes and the rates of coral reef degradation;
4) Calculating the economic value of coral reefs for fisheries, as breeding and nursery grounds, etc.;
5) Determining the Maximum Sustainable Yield (MSY) of exploitable coral reef resources:
6) Standardizing coral reef research methods to make results comparable;
7) Making a concerted effort at rationally managing and conserving coral reefs.

Coral reef research in many of the Central Indian Ocean countries, however, has remained at a low level of priority because of the lack of proper understanding by the policy makers of the value and vulnerability of coral reefs, as well as the lack of sufficient funds, scientists and expertise to carry out coral reef research.

The development of a central pool of Indian Ocean coral reef research and management expertise in collaboration with other relevant marine desciplines, with the blessings of the governments concerned, could provide the nucleus for much needed co-operative and comparative coral reef research in the region. There is also a need in the Central Indian Ocean and the adjacent region for a coral reef research data collection and dissemination service as well as a regional base for concerted efforts for rationally managing and conserving coral reefs.

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Table 1. Extent of damage to coral reefs of Maldives, Sri Lanka, India and Bangladesh.

<u>Code:</u> 0 = no problem; 1 = minor problems; and 2 = major problem

No. 1/	Cause of Damage	MALDIVES	SRI LANKA	INDIA	BANGLADESH
1.	a. Collection of shells and corals by tourists	0	1-2	0-1	2
	b. Collection of shells and corals for commercial purposes:				
	(i) souvenirs	1-2	1-2	0-2	2
	(ii) coral mining	1-2	2	2	0-1
2.	Spearfishing	0	1-2	0	0-1
3.	Collection of aquarium reef fishes	0	2	0	0-1
5.	Commercial fishing of reef fishes	1-2	1	0-1	0
6.	Explosives used for fishing and for public works	0-1	2	0-1	0
7.	Poisons used for fishing	?	?	0-1	0
8.	Other fishing methods destructive to coral	0-1	?	0	0
9.	Pesticides and detergents	0	0-1	0-1	0
10.	Sedimentation from fresh water runoff	0	0-1	1-2	1-2
11.	a. Domestic sewage and eutrophication	0-1	0-1	0-1	0
14.	Industrial wastes	0	1-2	0	0
20.	Dredging activities	1-2	?	0-1	0
21.	Construction activities on reefs	0	?	0	0
22.	Recreational impacts (scuba, snorkelling, boat and anchor damage)	0-1	1-2	0-1	0
24.	*Acanthaster* problems	0-1	0-2	0-1	0
	Most damage caused by:	1b	1,3,6,22	1,6,10	1,10

<u>1/</u> Human activities that cause damage to coral reefs, numbered according to the list adopted by the Coral Reef Working Group of IUCN (Coral Reef Newsletter No. 2).

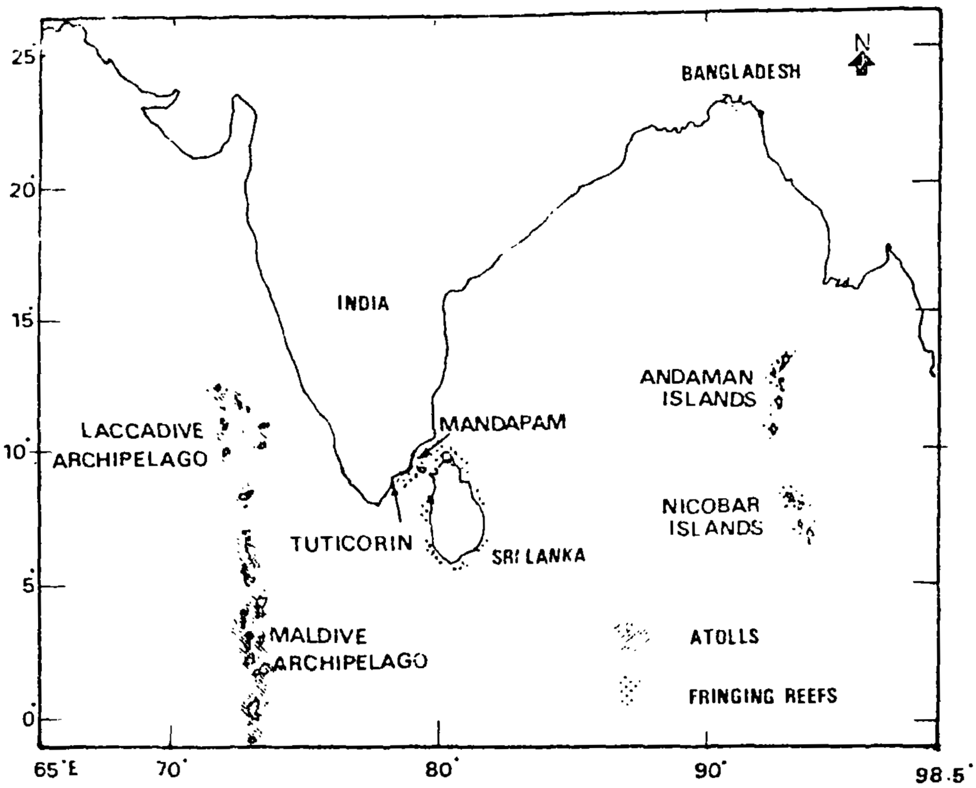

Figure 1. The major types of coral reefs in the Central Indian Ocean.

STATUS OF CRITICAL MARINE HABITATS IN THE INDIAN OCEAN

R. ORMOND

TMRU, Department of Biology
York University
York YO1 5DD
United Kingdom

INTRODUCTION

This paper summarises the nature, occurrence and status within the Indian Ocean of several types of habitat which, because of their significance in supporting marine life have, in recent years, been considered as *Critical Marine Habitats* (RAY, 1976). Critical marine habitats include the feeding, resting, breeding or nursery areas of marine animals, or are major sources of nutrients for feeding areas elsewhere (for example, mangroves), or are particularly rich in species (such as coral reefs), or are highly productive (such as seagrass beds) or of special scientific interest (such as coral islands). The contribution of such habitats is frequently essential for the survival and productivity of commercial species, as well as of rare and threatened ones.

Thus, for example, although commercial shrimp may be caught over a wide area of sea bed, these shrimp may depend to a very large extent on food originating from more limited areas of highly productive seagrass bed and mangrove; even smaller spawning areas and nursery areas may be essential to the survival of the stock (see FAO, 1981; PRICE, 1982). All these latter areas are critical to the maintenance of the resource, in a way that the much larger areas of open seabed are not.

In the Indian Ocean three types of critical habitat are especially important — coral reefs, mangrove forests and seagrass beds. Their social and economic importance arises in particular because of the very high primary productivity which these habitats are able to sustain: 500 to 5,000 g C m^{-2} yr^{-1}. By contrast, the open ocean in the tropics generally has a very low productivity, around 20-50 g C m^{-2} yr^{-1} (STEEMAN-NIELSEN, 1954; DOTI and OGURI, 1956; PLATT and SUBBA RAO, 1975). This is largely because of the way in which, in tropical seas, essential nutrients tend to become trapped at depth below a permanent body of warmer surface water. Thus, although most fish may be caught in the open sea offshore of reefs or mangroves, frequently a significant proportion of the food that sustains the fish stock will have originated directly or indirectly from one or more critical habitats.

In addition to these three highly productive marine habitats, three other habitats, small islands, kelp forests and enclosed soft-bottomed lagoons and mud-flats, also occur locally within the Indian Ocean, and may be regarded as critical in some degree. Each of these is considered more briefly.

CORAL REEFS

SIGNIFICANCE

Coral reefs and communities are perhaps the most economically significant coastal marine habitat over large parts of the Indian Ocean region, notably within the Red Sea and along the East African Coast. Handlining, principally for demersal and deep water reef species, is the most common method of fishing in these areas, and a majority of the species caught are mainly or frequently associated with the reef environment (see, e.g. WRAY, 1979). In the Red Sea, for example, five of the six most important groups of commercial fish, as judged by percentage contribution by weight of the total fish catch, are principally associated with the coral and reef environments. These are snappers (Lutjanidae), groupers (Serranidae), jacks (Carangidae), parrotfish (Scaridae), and emperors (Lethrinidae).

The significance of the reefs for fisheries relates in part to their very high primary productivity. Gross primary productivities of between 1,800 g C m^{-2} yr^{-1} (SARGENT and AUSTIN,

1954) and 7,300 g C m^{-2} yr^{-1} (GORDON and KELLEY, 1962) have been determined for coral reef areas in the Indian and Pacific Oceans, but generally the gross productivity appears to fall within the range of 2,000-5,000 g C m^{-2} yr^{-1} (see LEWIS, 1977). Estimates vary with the type of reef area under investigation, the methods used, and the assumptions made in the calculations.

The corals themselves are partly responsible for the high primary productivity of reefs due to photosynthesis occurring in their symbiotic zooxanthellae (YONGE, 1940, 1963). Corals have been found to fix carbon at rates equal to between 800 and 4,000 g C m^{-2} yr^{-1} (KANWISHER and WAINWRIGHT, 1967). But also important, and perhaps frequently of greater magnitude, is the productivity of the algal "lawn", composed of diatoms, filamentous green algae and small red algae, that colonises most of the rocky surfaces of the reef in shallow water (DAHL, 1974; WALKER, 1981). Such algal turf communities frequently occupy as much of the surface area of the reef as do the corals, and it has been found (JOHANNES et al., 1972; SMITH and MARSH, 1973) that the primary productivity of reef areas occupied by algal turf is higher than that of adjacent areas dominated by corals.

The algal turf is intensively grazed by various types of fish, especially by surgeonfishes (Acanthuridae), parrotfishes (Scaridae), damselfishes (Pomacentridae) and blennies (Blennidae), as well as by echinoids (sea-urchins) (RANDALL, 1961; BAKUS, 1966; DART, 1972; VINE, 1974; OGDEN and LOBEL, 1978), and this accounts both for the low standing crop in relation to turnover and productivity, and the relatively high biomass of grazing species on the reef (RANDALL, 1963; HOBSON, 1974). The coral, however, remains critical to the fish production of the community both in being responsible for the formation and maintenance of the shallow water substrata on which the algal turf develops, and in providing the habitat complexity which may allow dense populations of primary and secondary consumers to survive despite an abundance of predatory species (COLLETTE and TALBOT, 1972; SMITH and TYLER, 1972; CROWDER and COOPER, 1979; SAVINO and STEIN, 1982). Thus both coral and algal components contribute to the high primary productivity.

As a consequence of this high primary productivity the standing crop of fish on reefs may reach 5-15 times that in productive North Atlantic fishing grounds, and twice that for managed temperate lakes (STEVENSON and MARSHALL, 1974). It is suggested that maximum fish harvests that can be sustained from coral reefs are probably 10-20 mt km^{-2}. However, because of the patchy nature of the reef habitat and its tight recycling of nutrients, total catches may not be as high as in major fisheries in temperature and upwelling regions of the world's oceans. Nevertheless, as pointed out by GOMEZ (1980), the importance of coral reefs as providers of fish and shellfish for home consumption and local markets is probably underestimated.

Besides their significance to fisheries, reefs are of considerable actual or potential value in a number of other ways. They have proved invaluable for education and science, providing a suitable habitat in which to study scientific problems of much wider significance. As a result of their high diversity — individual reef sites may harbour as many as 3,000 different species of animal and plant — reefs represent a major genetic resource. Many reef animals have been found to contain pharmacologically active compounds which may be of medical value or lead to advances in medical research (CIERESZKO and KARNS, 1973; RUGGIERI, 1976). And other reef species may prove valuable for transplantation, exploitation or use in the development of mariculture (HESLINGA, 1980).

Also, coral reefs are generally regarded as being of considerable economic and social value for recreation and tourism. The need to protect reef areas used by a developing tourist industry has generally been accepted as a major impetus towards the establishment of marine parks and reserves, and such reserves have now been established in a wide variety of countries (see BJORKLUNG, 1974; OLINDO and ASAVA, 1975; ROBINSON et al., 1981; PERTET, 1982). Clear benefits have been recorded in, for example, the Netherlands Antilles (Bonaire Marine Park), Seychelles (Anne National Marine Park), Fiji (Tai Island) and Kenya (Malindi and Watamu Marine National Parks) (SALM, 1983). Benefits are not easily quantified, and relevant data not generally available, but as an example, in 1981 there were 1,322,300 visitors to the Buck Island Reef National Monument in the U.S. Virgin Islands (U.S. NATIONAL PARK SERVICE, 1981). Many Indian Ocean protected reef areas must also receive tens of thousands of visitors each year. Even where international tourism for leisure or water sports may not be significant, reefs clearly have considerable recreational potential for the local people, and as standards of living continue to rise, demand for more sophisticated pastimes, such as snorkelling and diving will probably increase.

OCCURRENCE

The distribution of reefs through the Indian Ocean region has been summarised recently by UNEP (UNEP, 1985). In the eastern Indian Ocean most of the reefs occur as fringing and patch reefs associated with offshore islands, there being relatively little development along the mainland coastline, presumably because of the sometimes heavy runoff of fresh water of which corals are intolerant. Similarly, Bangladesh and Pakistan have no coral reefs due to the high turbidity and soft substrate; one exception is the corals found on rocky substrate west of the Indus River.

India has two widely separated areas containing reefs, the Gulf of Kutch in the north-west and Palk Bay and the Gulf of Mannar in the south-east. Sri Lanka has shallow reefs especially on its south and east coasts. And the Indian islands of Lakshadweep (Laccadives) are entirely composed of atolls and therefore have extensive sublittoral reefs, as do the Nicobars and Andamans.

In East Africa well-developed fringing reefs and patch reef complexes occur along major sections of the coast of Somalia, Kenya and Tanzania, and around their offshore islands, while Madagascar has numerous inshore reefs.

Among the island groups of the Indian Ocean the Seychelles have varying reef development; they are generally absent from the west coast of Mahe, but those of Aldabra atoll have been especially well studied. In the Mascarenes, reef development generally diminishes from east to west, Rodriguez and Mauritius having well-developed fringing reefs although located far offshore.

The Red Sea in particular is dominated by coral reef habitats. Although (for zoogeographic reasons) species diversity is lower than in the Indian Ocean proper, the lack of fresh water input along most of the coastline results in vigorous inshore coral growth and formation of well-developed fringing reefs around almost the whole of the northern two-thirds of the sea. Around much of the Red Sea there are also extensive groups of offshore reefs, and in some regions complexes of reefs and islands extend as far as 50 km offshore, notably in the Wejh Bank (northern Saudi Arabia), the Farasan Bank and Farasan Archipelago (southern Saudi Arabia), the Dahlak Archipelago (Ethiopia), the Suakin Archipelago (Sudan), and in the mouth of the Gulf of Suez (Egypt).

In the Arabian Gulf reefs are less well-developed, but they are nevertheless more extensive and better developed than generally acknowledged (BASSON et al., 1977). Particularly well-developed fringing reefs surround Karan, Jana, Jurayd and three other offshore islands in the Saudi Gulf, Kubbar Island (Kuwait) and several islands between the Straits of Hormuz and Bandar Asalu (Iran). Some coral reefs are also present in the coastal waters of the United Arab Emirates and along the eastern coasts of Bahrain and Qatar, as well as in Oman.

In general, the reefs of the Indian Ocean are formed and/or dominated by a variety of species of the genera *Porites, Acropora, Goniastrea, Favia, Pocillopora, Stylophora, Millepora* and *Platygyra*. However, many other species and genera occur. Southeast Asia is generally considered as the faunistic centre of the Indo-Pacific region, and over 400 species of hard coral are thought to occur in Philippine waters alone. Moving westwards and northwards through the Indian Ocean the diversity of the coral fauna gradually declines. 174 species are recorded from the east coast of the Malaysian peninsula (DE SILVA et al., 1980), about 60 from the Gulf of Thailand, about 200 from Madagascar and from Chagos (SHEPPARD et al., 1984), 117 from south-east India, over 150 from the northern and central Red Sea (SHEPPARD, 1983), and about 40 from the Arabian Gulf (BURCHARD, 1979).

STATUS

Despite the acknowledged ecological significance and economic value of the reef habitat, degradation and even total destruction of reef communities is becoming common throughout the tropical regions in which reefs occur. Many factors are now recorded as having caused damage to coral areas, and these have been reviewed in general terms by JOHANNES (1970, 1975). In the Indian Ocean four impacts appear to be of special concern (UNEP, 1985).

Sedimentation

Sedimentation occurs as a result of land infill and other coastal construction work, as a result of dredging, and as a result of increased sediment load being carried to the sea by rivers following upstream land and soil erosion. Increased sedimentation is probably the single most destructive influence on coral reefs in the Indian Ocean.

Corals are generally intolerant of very heavy sediment loads; not only are the low light levels in turbid waters insufficient for adequate photosynthesis and growth, but the excessive sediment eventually clogs or smothers the coral polyps, preventing feeding and interfering with respiration (MARSHALL and ORR, 1931). Corals are able to free themselves of some sediment by secretion of mucus and the action of cilia, and some coral assemblages are characteristic of turbid waters (CROSSLAND, 1907; MARAGOS, 1972). But they are able to survive only so long as sediment particles are small and relatively east to remove; corals die when sediment loads increase, and in extreme cases may be completely buried (HUBBARD and POCOCK, 1972; MARSH and GORDON, 1974).

Examples of reef damage from sedimentation have been described from every part of the Indian Ocean. Sedimentation as a result of coastal and marine tin mining is the major cause of damage to reefs in Thailand (CHARSANG et al., 1961). Reef damage off the east coast of Peninsular Malaysia is especially noticeable in bays adjacent to agricultural development (DE SILVA et al., 1980). In the Gulf of Kutch (India) marked deterioration of coral reefs has occurred as a result of dredging of sand for the cement industry (CHAVAN, 1983). High coral mortality has occurred on Minicoy atoll, the largest in Lakshadweep, from the dredging of the main shipping channel (PILLAI, 1981). In the Comoros siltation of reefs and lagoons has lowered fishery productivity (WORLD BANK, 1979). Sedimentation from inland deforestation is reported to be causing damage in the Andaman Islands (WHITAKER, 1984). In Kenya coral in the area of Malindi Marine Park has been more or less killed as a result of the great increase in sediment input to the sea by the River Sabaki following upriver soil erosion (FINN, 1983). In Bahrain a quarter or more of the main coral area has been lost, and much of the remainder noticeably impacted as a result of sedimentation caused by land fill and dredging of shipping channels (IUCN, 1983b). And in the Saudi Red Sea, fringing reefs along 30 km of coastline have been seriously impacted by sedimentation following infill operations to construct the Jeddah corniche road (IUCN, 1984).

Damage by Echinoderms

There is some evidence for an increasing incidence within the region of damage to reefs by endemic populations of sea urchins (echinoids) and of the Crown-of-Thorns starfish (*Acanthaster planci*). Crown-of-Thorns starfish, which feeds directly on coral colonies, became notorious in the late 1960s and early 1970s because of the widespread destruction caused to coral communities on the Great Barrier Reef and some other parts of the western Pacific Ocean (e.g. Guam). More recently within the Indian Ocean, large populations of *A. planci* appear to have caused high coral mortality in Minicoy atoll (Lakshadweep) (PILLAI, 1981), in Sri Lanka, in Mauritius, and at locations in Peninsular Malaysia (DE SILVA et al., 1980).

Reef urchins normally feed by grazing on the algal turf that colonises rocky surfaces of the reef. However, when the urchins *Diadema setosum* or *Echinometra mathaei* are present in large numbers, the rasping action of their teeth erodes the bases and branches of the corals, causing the corals to collapse and die. In some cases the urchins appear to erode the living coral tissue or attack it directly. Extensive damage of this type, first reported from the Caribbean (LEWIS, 1977; KRISTENSEN, 1978), has now been observed in the Egyptian Red Sea (ORMOND, 1980), in Bahrain (BARRATT and ORMOND, 1985), in Kuwait (DOWNING, pers. comm.) and in Kenya (KENDALL, pers. comm.).

Whether such population explosions of reef echinoderms are an entirely natural phenomenon or are at least in part attributable to man's impact on the environment is as yet uncertain (see ENDEAN, 1976; FRANKEL, 1977, for opposing views). However, there is now good evidence that in the Red Sea at least high populations of urchins or Crown-of-Thorns are associated with reduced numbers of their principal predators, the large trigger fishes (e.g. *Balistoides viridescens*), the large puffer fishes (e.g. *Arothron hispidus*), and some of the emperors (*Lethrinus* spp.) (IUCN, 1984). It seems quite possible that throughout the western Indo-Pacific increased frequency and size of echinoid and *A. planci* population outbreaks may be attributable to the reduction of key predators from the marked increase in commercial and recreational fishing.

Mining of Reef Rock

Mining of reef rock from active reefs for road building, construction and lime production has caused extensive damage to reefs, especially in India, Lakshadweep, Sri Lanka, and to a lesser extent in countries of the eastern Indian Ocean. For example, in India fifty boats were engaged in quarrying the reef in Palk Bay, and by 1973 much of the reef had already been destroyed (PILLAI, 1973); extensive quarrying occurs at many sites in the Gulf of Mannar, at one site alone 30 boats removing 30,000 m^3 of reef per year (VENKATARAMANUJAM et al., 1981).

Dynamiting and Other Destructive Fishing Methods

The use of dynamite to kill and collect reef fish is not uncommon in countries from Egypt to Kenya and Tanzania to Sri Lanka. It is even prevalent in various protected areas; for example, damage to the Tarutai National Park in Thailand (ALEXANDER, 1983) is attributed in part to fish blasting. Dynamiting not only kills numerous non-commercial fish, but destroys the coral environment which provides food and shelter for the fish, thus reducing or eliminating the possibility of subsequent catches.

Six further impacts may be noted as having caused significant damage to some Indian Ocean reefs, and may also be emphasized because of the damage they are known to have caused in other coral regions.

Tourists and Visitors

The extent of damage to reefs and reef communities by souvenir collection and other tourist activities is increasingly appreciated, and within the Indian Ocean has been especially noted in Kenya, Sri Landa and Egypt. Such damage has been a major impetus for the establishment of marine parks and reserves (see BJORKLUND, 1974), since without control tourists or souvenir merchants collect corals and shells (JOHANNES, 1975; WELLS, 1981) and corals may be damaged by looking for shells or by trampling by fishermen or visitors. Significant coral damage also may be caused by the anchors of visiting boats (DAVIS, 1977). Visitors also may affect the abundance of important fish species by excessive spearfishing, thus competing with local fishermen, as well as potentially altering the ecology of the reef. Similar in nature are the effects of excessive collection of aquarium fish, recognised as a particular problem in Sri Lanka. Such over-exploitation can deplete considerably the populations of a variety of fish species (LULOFS, 1973; LUBBOCK and POLUNIN, 1975).

Over-Exploitation of Food Species

Increasingly reef fish and reef animals in many areas are being over-fished and over-exploited. Without controls the number of commercial fish species taken from reefs on over-populated tropical shores has led to a decrease in yield below that which, in theory, could be sustained (JOHANNES, 1981). Many other edible or semi-edible reef animals are being collected to near extinction, including many molluscs, especially giant clam (*Tridacna* spp.) and conch (*Lambis* spp.), echinoids and holothurians, crabs, crayfish and other crustacea.

Oil Pollution

The effects of oil pollution on corals was recently reviewed by LOYA and RINKEVITCH (1980). It was initially thought that corals were relatively unaffected by oil pollution, and, following large oil spills reefs, they were found to be significantly less impacted than other tropical marine communities (RUTZLER and STERRER, 1970; SPOONER, 1970). However, it appears that the extent of damage depends on the circumstances. Large spills can float over sublittoral reef communities and any toxic chemicals diffusing below the surface layer are soon dispersed.

However, corals that become coated with oil when exposed at low tide can be seriously damaged (JOHANNES et al., 1972), while serious effects also appear to result from chronic long-term exposure to even relatively low levels of oil.

In general, impacts on corals in the Indian Ocean from oil pollution appear much less frequent than might be supposed. LOYA (1975, 1976) found an almost complete lack of recolonisation by corals on reefs at Eilat, chronically polluted by oil. Field and laboratory studies show a decreased production of ova, premature discharge of eggs, damage to larvae, and a reduction in successful settlement by larvae when corals are exposed to long-term oil pollution (RINKEVITCH and LOYA, 1977, 1979). Parts of the Gulf of Suez (Egypt) have been subject to even more serious chronic oil pollution (WENNINK and NELSON-SMITH, 1979), and ORMOND (1980) has suggested this as one factor responsible for an apparent decline in coral cover on reefs in parts of this area.

Eutrophication and Sewage

Sewage discharge can result in severe damage to coral communities. The best studied case is of the killing of corals over most of Kanehoe Bay, Hawaii (BANNER and BAILEY, 1970; MARAGOS, 1972). Coral death has been demonstrated to correlate with nutrient concentrations and turbidity (MARAGOS, 1972), but neither are thought to be directly responsible. The mechanism may vary with different circumstances in different places; at Aqaba (Red Sea) WALKER and ORMOND (1982) have concluded that sewage discharge killed corals by stimulating the growth of algae, which in turn increase the rate of sedimentation affecting the coral.

Other wastes may cause similar effects. In both the Jordanian and Egyptian Red Sea local eutrophication leading to algal growth and coral damage has resulted from the escape of phosphate dust during loading of ships for export of crushed phosphate rock (FISHELSON, 1973; WALKER and ORMOND, 1982).

Thermal Pollution

Death of corals resulting from discharge of heated effluents has been observed, for example, in Florida (ROESSLER and ZIEMAN, 1969), Guam (JONES and RANDALL, 1973) and Hawaii (JOKIEL and COLES, 1974). The sensitivity of coral and reef animals to modest temperature increase should be especially noted here, since in parts of the Indian Ocean, with higher open water temperatures, coral and reef animals are likely to be living very close to their absolute limit of tolerance.

Altered Salinity

Although some corals live in salinities up to about 50 ppt (KINSMAN, 1964), most corals are killed by long-term exposure to salinities much above normal (EDMONDSON, 1928). Hypersaline effluents may be discharged from power plants or desalination plants, or result from restriction of water circulation in lagoons or bays. Corals also are affected by a reduction in salinity, and most are killed in less than a week by exposure to 75% seawater (EDMONDSON, 1928). At various locations in the Indian Ocean small areas of coral probably have been affected by such changed conditions.

MANGROVE FORESTS

SIGNIFICANCE

The term mangrove covers a wide variety (60 or so) of species of trees that belong to a number of essentially unrelated families, but which, often together, alive long the edge of brackish or sea water shores in many parts of the tropics. Here they can form zones of dense forest, up to several kilometers wide and containing as many as 20 or so different species. Such diverse and expansive communities are found only in a few areas; in other areas mangrove stands are much narrower and may contain only a few species.

In recent years it has come to be appreciated, as a result of many investigations in different parts of the world, that mangroves constitute a highly significant natural habitat and a very valuable resource. Their significance derives from the fact that mangrove communities, like coral reefs, are often very productive (in terms of the plant material produced by unit area), especially by comparison with the productivity of neighbouring areas or of the open ocean. A figure of 350-500 g C m^{-2} yr^{-1} seems typical for mangrove primary production (GOLLEY et al., 1962; MILLER, 1972; LUGO and SNEDAKER, 1974).

Man exploits this productivity, both directly and indirectly. There is now heavy direct utilisation of mangroves in many parts of the Indian Ocean. Coastal people use the wood for fuel, fishing stakes and floats, house posts and thatch, and boat building. The leaves are used for fodder and in the past the bark of Indian mangals has been used for tannin production. One or two species produce edible leaves or fruit. IUCN's review (1983a) on the global status of mangrove ecosystems lists nearly 50 products in all that are obtained from mangrove forest. Most recently mangals have been used heavily for chipboard production. In Indonesia, for example, the combined export and domestic value of all mangrove forestry products was estimated in 1978 at about US$ 26 million. Managed and cropped on a sustainable basis, mangroves can continue to provide such raw materials for man's use.

But the indirect benefits to be obtained from mangal communities may be more significant. Mangrove is frequently the basis for important marine food chains; the leaves fall into the water where they decompose, and the resulting detritus and bacteria provide food for meiofauna and larger molluscs and crustaceans, including some commercial species of shrimp. Food is also provided indirectly for various fish, including commercial species such as snapper (*Lutjanus* spp.) which feed in turn on the invertebrate fauna (see ODUM and HEALD, 1972; ODUM and JOHANNES, 1975). The roots of the mangrove can also provide shelter for the invertebrates and fish which depend on their productivity. And organic and inorganic nutrients may also be exported in the form of leaf litter to adjacent habitats in deeper water to support yet other animal species (ODUM and HEALD, 1972).

Thus the productivity of the marine prawns *Penaeus merguiensis, Metapenaeus monoceros* and *Metapenaeus brevicornis* has been clearly related to existing mangrove areas (UNAR and NAAMIN, 1984). In the Philippines the major fishing grounds are generally located near areas bordered by mangrove (GOMEZ, 1980). It was estimated in 1970 (ROBAS, cited in WALSH, 1977) that one acre (0.405 ha) of undisturbed mangal estuary in Florida yields $7,980 worth of commercial fish products in 20 years. And in Indonesia, the combined exports and domestic value of mangrove-linked fisheries products in 1978 was at least US$ 194 million (SALM and HALIM, 1984).

In addition, mangroves are significant as a habitat or refuge for a wide variety of other species of animals and plants of scientific interest or ecological value, including some that are relatively uncommon or rare (see MacNAE, 1968). For example, IUCN (1983a) indicates that over 300 species of plants and over a thousand species of marine invertebrates and vertebrates have been recorded from Asian mangrove areas. In addition, 177 birds and 36 species of mammals have been reported in association with mangroves in this area. More conspicuous species of scientific interest and conservational value include a variety of birds, such as herons and egrets, spoonbills and pelicans, frigate birds and boobies, the Estuarine Crocodile, and the rare Royal Bengal Tiger, found in the Sundarban mangroves of Bangladesh and India.

Mangroves also have an indirect value in tending to control both coastal erosion and coastal flooding (DAVIS, 1940; CARLTON, 1974). By buffering fresh water run-off into inshore waters, they also protect coastal reefs against variations in salinity to which they are particularly sensitive. And finally, mangroves also have aesthetic and landscape value, frequently forming the only dense plant growth on many otherwise sparsely vegetated shores around the Indian Ocean.

OCCURRENCE

The extent and development of mangrove stands and forests in the Indian Ocean have recently been summarised by UNEP (1985). They reach their fullest extent in southeast Asia, where Indonesia possesses 3,806,119 hectares (SALM and HALIM, 1984), Malaysia 652,219 hectares (SASEKUMAR, 1980), and Thailand 312,714 hectares (PIYAKARNCHANA, 1980). Mangroves are moderately well-developed in more restricted areas of the Indian sub-continent and East Africa. In the northern Bay of Bengal, the Sundarbans, in the southern Ganges delta, support over 500,000 hectares (MUKHARAJEE, 1984), and in Pakistan and western India the Indus delta supports 250,000 hectares (SNEDAKER, 1984). Mangrove is less developed in Sri Lanka and patchy or thin in the island nations and groups of the Indian Ocean, although mangroves in the Andaman and Nicobar Islands appear to be among the least disturbed.

In the Arabian Gulf relatively few stands of mangrove survive, but in the Red Sea patches of mangrove are scattered along both coasts, being more abundant in the south and thinning out to the north. The Saudi Red Sea coast, for example, is thought to support approximately 5000 hectares of mangrove (IUCN, 1983a).

The character and development of the mangal changes across the region in a similar way (UNEP, 1983a). In the east many species contribute to the mangal, for example 38 in Indonesia (SOEGIARTO and POLUNIN, 1982), and well-developed forests incorporate four or more zones dominated by different genera, particularly *Avicennia, Sonneratia, Rhizophora* and *Bruguiera* (GONG et al., 1980). The mangroves of eastern Bangladesh are essentially similar to those of Malaysia (SNEDAKER, 1984), whereas on the west coast of India the mangrove areas are dominated by *Avicennia marina* and *Rhizophora mucronata*, although a total of about twenty genera are reputed to be found in the well developed mangals of the Gulfs of Cambay and Kutch (UNTAWALE, 1984). In Pakistan the Indus River delta is dominated by *Avicennia officinalis*, often in poor stands, with only occasional trees of *Ceriops* (SALM, 1975).

Avicennia marina is the only mangrove recorded from the Arabian Gulf and the Gulf of Oman (BASSON et al., 1977), and this is also the principal species in the Red Sea, although very occasional stands of *Rhizophora mucronata* also occur (IUCN, 1984), mainly towards the south.

STATUS

Despite their enormous value as a renewable resource, mangrove forests within the Indian Ocean are being destroyed at an alarming and accelerating rate. They are being lost by increased pressure of cutting for fuel and timber, related to the rapid growth of human populations. They are cleared, drained and/or felled for agriculture, residential or commercial development. They are cleared for conversion into fishponds. And in particular in recent years large areas of mangrove forest have been felled for conversion to chipboard or paper, generally for export to industrial nations outside the region (MacNAE, 1974; SOEGIARTO, 1980). In the past, forest managers frequently have cropped mangroves on a 16 to 30 year silvicultural cycle, but increasingly timber companies (particularly Japanese) appear to be felling large areas with no regard for either traditional or recommended silvicultural practices.

Thus in Indonesia an estimated 700,000 hectares were converted to agricultural land between 1969 and 1979 (SOEGIARTO, 1980). In India, BLASCO (1977) estimated that only 365,000 hectares remained, roughly half the official estimate for 1963. Total destruction is even reported over large areas of the massive and relatively inaccessible stands in the Bay of Bengal (MUKHERJEE and TIWARI, 1984). In western and southern India, much of the originally extensive mangrove has been removed (UNTAWALE, 1984; KRISHNAMURTHY and JEYASEELAN, 1984). And even where mangroves represent a scarce resource they are being lost without regard to their value; over half of the mangrove stands on the Saudi Arabian Gulf coast have been infilled for housing or industrial development in the past ten to fifteen years (IUCN, 1982).

In addition to such direct destruction, mangrove forests are particularly susceptible to impacts that affect their leaves or aerial roots, since both organs are essential to the plants' survival, even for short periods (see ODUM and JOHANNES, 1975). The leaves, in addition to meeting the energy requirements of the plant through photosynthesis, are the main organs of salt secretion that allow the plants to survive in saline waters. And the aerial roots are essential in allowing oxygen access to the underground system which is frequently growing in anaerobic conditions; respiration takes place through small pores (lenticels) which are susceptible to smothering and clogging. The most significant additional impacts from the point of view of Indian Ocean mangroves may be summarised as follows:

Coating of Aerial Roots by Fine Sediment/Erosion

This can occur though natural sediments released near mangrove stands by dredging or landfill activities, through excessive organic matter input or through the introduction of chemicals into mangrove waters. In St. Croix (U.S. Virgin Islands) release of NaOH (used in bauxite extraction) into mangrove waters caused precipitation of hydroxides of various metals, including magnesium, onto aerial roots, and thus the death of mangrove plants (ODUM and JOHANNES, 1975).

Oil Pollution

The serious effect of oil pollution on mangrove stands has been demonstrated on a number of occasions. For example, SPOONER (1970) observed that an oil spill in Tarut Bay, Saudi Arabia, defoliated mangroves although the mangrove plants survived. And within the Red Sea major oil spills already have occurred on the Egyptian Coast, severely impacting several stands of *Avicennia marina*, particularly on Abu Minqar Island, near Hurghada (BARRATT, 1982).

Oil pollution damages mangroves by an immediate smothering effect on aerial roots, and under circumstances of heavy oiling, mangrove plants can be killed in 48-72 hours (IUCN, 1983a). In addition, oiling can cause chronic poisoning as a result of the uptake of volatile petrochemical fractions through aerial root and leaf lenticels. More detailed discussion of the effects of oil on mangrove can be found in BAKER (1982).

Fresh Water Flooding

As a result of inland construction works or coastal engineering interfering with land drainage, mangroves can be flooded with fresh water and subsequently die (see ODUM and JOHANNES, 1975). Thus the Sundarbans mangrove ecosystem has been severely affected by the Farakka Barrage, which has diverted waters of the Ganges; as a result floods, soil toxification, land instability and pest infestations have increased, while fertility, water quality, forests and fisheries have declined (IUCN, 1983a).

Herbicide Damage

Mangroves are unusually susceptible to herbicide damage because of the critical role of the foliage in salt secretion. Moreover it appears that herbicidal damage can prevent re-establishment of any new mangrove community for at least six years and possibly for much longer. This type of damage could become increasingly significant in the region because of the rapidly increasing amounts of pesticides being used by many of the countries bordering the Indian Ocean (WHITMORE, in UNEP, 1985). In addition, the potential for such damage also should be taken into account when considering the use of potentially toxic oil dispersants in clearing up oil spills in mangrove areas.

Thus it is appropriate to note here that decimation of mangrove resources has stimulated international concern and activity. In 1978 SCOR, in collaboration with UNESCO's division of Marine Sciences, established a working group on Mangrove Ecology, and in 1980 IUCN established a working group on Mangrove Ecosystems. UNESCO continues to be active in promoting activities related to mangrove research and management, including mangrove mapping by remote sensing, a UNEP/UNESCO regional project on sedimentology in mangrove areas, and the development of a network of mangrove managers and scientists in the Indian Ocean.

SEAGRASS BEDS

SIGNIFICANCE

Seagrasses are true flowering plants which propagate both vegetatively and by the dispersal of seeds. They occur in soft-bottom, shallow-water marine environments and are usually permanently submerged, although they occasionally are found intertidally. About 50 species in twelve genera are recognised; as is common with many marine groups, the highest diversity of seagrasses is found in the Indo-West Pacific where there are seven genera.

The principal ecological significance of seagrasses lies in the fact that the seagrass beds they form constitute an important shallow water habitat. Within the dense array of seagrass blades a large variety of fish and invertebrates obtain shelter and food, much of which derives directly or indirectly from the seagrasses themselves (DEN HARTOG, 1970; OGDEN, 1980). An important feature of the seagrass plant is its extensive root and rhizome system which enables it to colonise various soft-bottom substrates, provides a semi-permanence to the seagrass community, and also enables the plants to draw on mineral nutrients from the seabed.

Principal features of the ecology of seagrasses and seagrass communities may be summarised as follows (IUCN, 1984; see also WOOD et al., 1969; ZIEMAN, 1975):

High Productivity

Tropical seagrass beds are among the most productive of natural ecosystems. Productivities of 0.5-16 g C m^{-2} day^{-1} and 500-3,000 g C m^{-2} yr^{-1} have been recorded (DREW, 1971; PATRIQUIN, 1973; McROY, 1974; BUESA, 1975; ZIEMAN, 1975). A recent review of the topic is provided by ZIEMAN and WETZEL (1980). Unfortunately few studies have yet been made of the productivity of the species occurring in the Indian Ocean, although WAHBEH (1980) found an *Halophila stipulacea* stand at Aqaba (Jordan) to have an estimated productivity of 617 g C m^{-2} yr^{-1}.

Direct Grazing

Their cellulose walls make seagrasses difficult for most fish and invertebrates to ingest directly or to digest (by comparison with algae). Nevertheless many species do feed on seagrasses, and seagrasses are a major source of food for adult Green turtles (CARR, 1952; HIRTH et al., 1973; ROSS, 1979), for juvenile Hawksbill turtles (PHILLIPS and McROY, 1980), and for the Dugong (LIPKIN, 1975). A summary of other species reported feeding directly on seagrass is given by OGDEN (1980), who estimates that in the Caribbean, for which most the data are available, about thirty species of fish and five of sea urchins feed directly on seagrass. In the Red Sea, WAHBEH (1980) found that the sea urchin *Tripneustes gratilla*, the surgeonfishes *Zebrasoma xanthurum* and *Ctenochaetus striatus*, and rabbitfishes *Siganus* spp. were the principal direct consumers of a *Halophila stipulacea* seagrass bed near Aqaba (Jordan). The urchins accounted for 33.5% and the fish for 4.9% of the total loss of leaf and rhizome biomass. The bulk, however (61.6%), entered the decomposer food chain.

Production of Detritus

It has generally been concluded that most of the primary productivity of seagrass beds enters the decomposer food chain (WOOD et al., 1969). The senescent blades break off and decompose among the seagrass, on the surrounding seabed, or on an adjacent shore. The resulting detritus and bacteria (see FENCHEL, 1970) are consumed by an abundant meiofauna, by sponges, sessile polychaetes, many deposit- and suspension-feeding bivalves, a wide variety of crustacea, including commercial shrimp species such as *Penaeus semisulcatus*, and by some fish, including commercial species such as *Gerres oyena* and mullet (Mugilidae) (O'GOWER and WACASEY, 1967; ORTH, 1971; BROOK, 1975).

Substrate for Epiphytes

Well-developed seagrass beds generally support a dense epiphytic community that often makes a major contribution to the primary productivity of seagrass beds (WOOD, 1959; JONES, 1968; PENHALE, 1977). Particularly dense epiphytic turfs of sciaphilic algae are found on the stems and roots of *Thalassodendron ciliatum* (LIPKIN, 1976; DEN HARTOG, 1970). Many fish and herbivorous molluscs probably graze on the epiphytic algae rather than on the seagrass itself (see, for example, LIPKIN, 1976; PENHALE, 1977).

In addition to their contribution to primary productivity, these epiphytes may play an important role in supplying mineral nutrients to the seagrass bed, especially in nutrient-poor waters. Very high rates of nitrogen fixation have been reported for epiphytes on *Thalassia testudinum* and for bacteria associated with the roots (e.g. GOERING and PARKER, 1972). However, other workers have reported much lower levels of nitrogen fixation in seagrasses (e.g. McROY et al., 1973; ZIEMAN, 1975).

Sulphate Reduction

ZIEMAN (1975) reported that seagrasses may also be important in the sulphate cycle. There are no reports of this application by seagrasses in the Indian Ocean. However, the roots of seagrasses often are found in reducing conditions.

Reduction in Current Velocity and Sediment Binding

Dense stands of seagrass often reduce open water current velocity almost to near zero close to the substrate. This allows increased sedimentation and the development of an infauna and inflora that would not survive otherwise. The seagrass roots and rhizomes further bind the sediment that has settled. Of the species of seagrass reported from the Indian Ocean, only *H. ovalis* and *H. ovata* are not considered to be significant sediment binders (DEN HARTOG, 1970).

Refuge for Commercially Important Species

A variety of commercially important species of fish, mollusc and crustacea use seagrass beds as a refuge and frequently as a source of food as well. Pearl oysters are closely associated with seagrass beds in the Arabian Gulf (BASSON et al., 1977), and seagrass beds are also important areas for the juveniles of penaeid shrimps, including the commercially important *P. semisulcatus* (PRICE, 1982). There is evidence to suggest that the collapse of the commercial shrimp fishery around Kuwait (D.A. JONES, personal communication) is closely associated with the extensive destruction of the seagrass beds during industrial development in that area.

Given their significance to the various fisheries, seagrass beds are considered to be a resource of considerable economic importance. The approximate economic value of the fishery generated by a seagrass dominated bay in Saudi Arabia (Tarut Bay) has been estimated by BASSON et al., (1977). If seagrass in the bay were incorporated into food chains, the estimated value of the fish yield would be $8 million. If instead, grass were converted at the same efficiency (1% overall) into shrimps, the calculated value would be nearly $12 million. On the other hand, if the seagrass were grazed directly by Green turtles, at an efficiency of 10%, the turtle yield would be $46 million. The value of these natural resources is based on 1977 prices, and the calculations are very approximate and largely theoretical. Nevertheless, they deomonstrate the considerable value of local biological resources which, if managed correctly, are renewable.

OCCURRENCE

In general (see UNEP, 1985) knowledge of the occurrence and extent of seagrass beds in the Indian Ocean is very incomplete, save in two or three countries, notably Indonesia and Saudi Arabia, where recent surveys specifically have examined the question. In the eastern Indian Ocean (Indonesia) 13 species, of which *Thalassia hemprichii* is the most widespread, have been recorded (SOEGIARTO and POLUNIN, 1982). In the western Indian Ocean (East Africa) nine species have been recorded (ALEEM, 1984) and in the Red Sea (Saudi Arabia) ten species (IUCN, 1984).

Fewer species are recorded from any one of the island groups of the western Indian Ocean; six have been recorded in the Seychelles (TAYLOR, 1968), and six in Mauritius, although some of these species are abundant around parts of many of the higher granitic islands (i.e., Seychelles, the Mascarenes, the Andamans and the Nicobar Islands).

Around the periphery of the region, under cooler or more saline conditions, the number of species may be reduced. Only three species, *Halophila stipulacea, H. ovalis* and *Halodule uninervis*, occur within the Arabian Gulf, but there, as indicated above, they nevertheless form an extensive and economically significant habitat.

Along the coasts of the Indian sub-continent there are extensive seagrass beds in southern India and in the numerous estuaries and embayments of Sri Lanka. In other areas, however, dense seagrass beds are uncommon or not extensive. This seems to be true of at least some of the coasts of Pakistan and of much of the coasts of western India, probably because the coastal waters are exposed and turbid so that seagrasses can not easily find a foothold or grow well.

STATUS

Because their distribution and significance is incompletely known, the extent of damage to seagrass beds within the Indian Ocean can not be assessed accurately. Some impacts, however, have been reported. Industrial and agricultural runoff have damaged seagrass beds in Sri Lanka (SALM, 1975), and industrial wastes, sewage discharges and overfishing have been reported to have led to the destruction of a seagrass bed in Mauritius (PROCTER and SALM, 1974). However, damage to and loss of seagrass beds have been most extensive, and most serious in terms of economic consequences, in Saudi Arabia, Kuwait and Bahrain. There large areas of seagrass beds have been lost by extensive land reclamation and infilling, by dredging to form shipping and boat channels, and by excessive sedimentation resulting from upcurrent infilling or dredging activities.

In view of the limited information available on the status of seagrass beds within the Indian Ocean, it is perhaps most relevant to summarise (following IUCN, 1984) the range of impacts to which they, like coral reefs and mangroves, have been found to be susceptible. It should, however, be noted that different species can cope with different stresses to varying degrees, so that generalisations are not easily made, except where the impacts are extremely severe (see ZIEMAN, 1975). In general, the most susceptible habitats are the shallow subtidal and intertidal ones, and the most susceptible species are those that lack a significant starch reserve in their rhizomes.

Dredging and Filling

Because seagrass beds occur in shallow coastal habitats, they lie in the prime area for coastal development and are frequently destroyed either by infilling to produce reclaimed land or, alternatively, by dredging for the production of channels for boats, for the exposure of bedrock to provide a more secure substrate for construction, or for the provision of infilling materials for construction elsewhere. Similar damage also can occur by bottom trawling or where boats or walkers regularly scour seagrass-covered shallows. Finally, construction associated with dredging and landfill indirectly can influence seagrass beds by altering water current patterns and producing new water-quality conditions.

Infilling causes direct death of seagrass beds through burial, and both infilling and dredging can cause deterioration and death of seagrass beds in neighbouring areas because of increased sediment loads and increased water turbidity. Heavy sediment loads may suffocate the plants directly, while increased turbidity reduces light levels in the water column and on the seagrass, thus reducing or preventing photosynthesis and growth.

Sedimentation also will smother and kill the flora and fauna associated with the beds. Even where protective curtains or screens are used around sites of dredging and infilling, turbidity may still be sufficient to cause damage (ZIEMAN, 1975). These turbidity problems can also affect associated coral reefs, leading to their death.

Eutrophication

Eutrophication is of only limited benefit to seagrass communities. Seagrasses can take up extra nutrients, leading to stimulated growth. Enrichment also fosters the development of epiphytic algae and phytoplankton. However, seagrasses generally require high levels of incident light energy so that increased epiphytic growth and high water turbidity by reducing light levels may actually decrease growth of the seagrass.

Temperature and Salinity

The effects of alterations in temperature and salinity, such as might arise from coastal construction work or effluent discharge, depend upon the species concerned. For example, *Syringodium* is reported to die at temperatures below 20°C (ZIEMAN, 1975), whilst *Halophila ovalis* can tolerate seawater temperatures lower than 10°C. Conversely, *Halophila stipulacea* is very salinity-adaptable whilst *Cymodocea serrulata* is very sensitive to a decrease in salinity (DEN HARTOG, 1970).

Overgrazing

As already discussed, although many herbivores do not eat seagrass, a variety of species, particularly of fish and sea urchin, feed directly on the plant, and where these grazers are present in very large numbers, overgrazing and destruction of the grass bed can occur. Such overgrazing by reef-associated fish frequently results in a bare halo around patch reefs in the Caribbean (RANDALL, 1965; OGDEN, 1972), and in some localities epidemic overgrazing by sea urchins is reported (CAMP et al., 1973). In the Red Sea recently there has been loss of seagrass at Aqaba (Jordan) through overgrazing by large numbers of the urchins *Diadema setosum* and *Tripneustes gratilla* (BENAYAHU and LOYA, 1977; MASTALLER, 1979; WAHBEH, 1980), and such population outbreaks of echinoids may occur as a result of the loss of predators taken by man (LOWRY and PEARSE, 1973; BREEN and MANN, 1976). Thus intensification of some fisheries can indirectly result in the loss of seagrass beds.

Oil Pollution

As seagrasses are generally subtidal, they are less susceptible to damage from oil than organisms in the intertidal zone. However, in some regions large areas of reef flat, sometimes containing seagrasses, are exposed at low tides and these seagrasses probably would be killed by direct contact with oil. Also surface floating rafts of oil, if present for long enough, can affect significantly seagrasses by reducing light penetration, by increasing water temperature through absorption of light energy, and by reducing oxygen exchange at the air/oil/water interface.

In addition to direct effects on the seagrass, more serious and possibly longer-term damage may occur through interaction between the oil and the sediments in which the seagrass is growing. Much of the oil and tar from oil spills eventually sinks to the seabed, and here oil and sediment can agglomerate into more buoyant lumps and pellets which, in relatively shallow water, can be removed by wave and current action. DIAZ-PIFFERER (1962) recorded the loss of 3,000 m^3 of sand from a Puerto Rican beach in less than a week due to this effect. Where oil settles over seagrass beds, loss of sediment can lead to uprooting of the grass and damage to or destruction of the bed.

In this connection it should be noted that a number of oil spill dispersants cause floating oil to sink so that seagrass beds that might otherwise be safe from oil could be directly contacted by it, or impacted by destabilisation of sediments as just described.

Metals

Seagrasses are reported to have a very high uptake of metals (e.g. PARKER et al., 1963), despite the fact that the roots are often in reducing conditions that limit the lability of metal ions. Not only can excess metals be concentrated in the seagrass but they become available for further accumulation in the food chain. This has been of particular concern in some tropical regions where high concentrations of metals have been released into the environment from desalination plants (see CHESHER, 1975).

Disease

In the 1930's a large-scale and so far permanent loss of beds of the temperate seagrass *Zostera marina* occurred on both sides of the Atlantic (see STEVENS, 1936) that was attributed to a fungal disease (TUTIN, 1934). So far as is known, disease has not caused extensive damage to seagrass beds in the tropics. However, susceptibility to disease is frequently a response to increased environmental stress, and subsequent loss of *Zostera* beds in Japan was linked to industrialisation and increasing water pollution along the coast (NAIKAI REG. FISH. RES. LAB., 1967; NASEI REG. FISH. RES. LAB., 1974).

OTHER CRITICAL HABITATS

SMALL ISLANDS

Small islands, offshore of the mainland coast or oceanic islands in deeper water, are numerous in many parts of the Indian Ocean. For example there are 200 or more small coral cays and islands along the eastern seaboard of the Red Sea alone. Island groups of the Indian Ocean, the Maldives, Seychelles, Lakshadweep, Nocibars and Andamans, include many small coralline or volcanic islets, besides the larger islands that support human population.

The smaller islands constitute significant critical habitats. Not only are they generally surrounded by shallow water marine habitats, coral reefs, seagrass beds and mangrove stands, which are considered critical in their own right, but the islands themselves are vital breeding and nursery areas for animals that are of significant economic value and particular scientific and popular interest — namely seabirds and turtles.

Forty-two species of seabirds breed in the Indian Ocean, almost all of these invariably on small offshore or oceanic islands. Of these 42, fourteen are endemic to the region (UNEP, 1985); these are:

Mascarene petrel	*Pterodroma aterrima*
Barau's petrel	*Pterodroma baraui*
Jouanin's petrel	*Bulweria fallax*
Abbott's booby	*Sula abbotti*
Indian cormorant	*Phalacrocorax fuscicollis*
Socotra cormorant	*Phalacrocorax nigrogularis*
Javanese cormorant	*Halietor niger*
Christmas Island frigatebird	*Fregata andrewsi*
White-eyed gull	*Larus leucophthalmus*
Sooty gull	*Larus hemprichii*
White-cheeked tern	*Sterna repressa*
Saunder's tern	*Sterna saundersi*
Lesser crested tern	*Sterna bengalensis*
Lesser noddy	*Anous tenuirostris*

In the Indian Ocean the breeding of all seabirds has become increasingly impacted by man's activities, largely by general disturbance and habitat changes (linked to the spread of villages, holiday homes, navigational and military installations to offshore islands), by accidental introduction of predators (e.g. cats and rats), and also by egg collection and taking of chicks. In the past eggs have been collected in small quantities by fishermen and coastal villages, and generally in a sustainable manner; but in recent years, as a result of human population expansion and the increase in range of motorised fishing craft, taking of eggs and chicks has greatly intensified. The Pelicaniformes (boobies, frigatebirds, etc.) appear to have been most affected and to be in serious decline: They and their young are taken for food; they are especially vulnerable to human disturbance because gulls take their eggs when they leave the nest; and they have been excluded either by the destruction of trees used for nesting, or by the planting of previously treeless islands with coconuts. In Indonesia (DE KORTE, 1983) and in the central/western Indian Ocean (FEARE, 1984), many islands that supported breeding colonies earlier in this century now either lack any breeding birds or support only greatly reduced numbers.

Small offshore and oceanic islands also provide important nesting grounds for all five species of marine turtles found in the Indian Ocean: the Green turtle (*Chelonia mydas*), the Hawksbill (*Eretmochelys imbricata*), the Loggerhead (*Caretta caretta*), the Leatherback (*Dermochelys coriacea*), and the Olive Ridley (*Lepidochelys olivacea*). Of these, the Loggerhead may nest predominantly on mainland beaches, but the others nest frequently, and the Hawksbill almost exclusively, on small islands. Green and Hawksbill are the most widespread species and are the only ones breeding in significant numbers in the Red Sea, in the Arabian Gulf, and on the island groups of the western and central Indian Ocean.

As is well-known, sea turtles are threatened worldwide, and all of the above species are now regarded as endangered in the IUCN Amphibia and Reptile Red Data Book (HONNEGER, 1975) and also are listed in the CITES Convention of Trade in Endangered Species. The precipitous decline of the turtle in the Indian Ocean (as elsewhere) stems in part from increasing human access to and exploitation of small remote islands that previously had harboured unimpacted breeding colonies. The adults are taken on the beaches for their meat, or in the case of the Hawksbill for their shell (tortoiseshell). At an increasing number of sites eggs are collected to excess; at some sites in India and Malaysia over 90% of eggs are thought to be taken. The situation is further exacerbated by the general disturbance caused by visitors and by residential and other developments. There is often an increased number of feral or domestic dogs digging up the nests. And the adults also may be hunted by net or harpoon in the immediate neighbourhood of their breeding grounds.

Such a decline is continuing despite an appreciation that the region's turtle population represents a valuable renewable resource. For example, the value of turtle products exported from Indonesia in 1980 was US$ 928,539 and in 1981 US$ 407,542 (SALM and HALIM, 1984). Groombridge (1985) suggests that the Olive Ridley in India could be sustainably exploited, given proper management, and, once they have been allowed to recover, the same might be said of other turtle populations in the region. The decline, however, continues despite the laws and regulations which in various countries of the region should give adequate protection to all species. Enforcement is difficult, and exploitation and impact appear continuous and very largely uncontrolled.

KELP FORESTS

Reasonable growth of macroalgae occurs intertidally and subtidally on rocky shores and reefs around various parts of the Indian Ocean (see, for example, JAASUND, 1976). While these algae may be relatively productive, they do not form a conspicuous critical habitat except within one area of the Arabian Sea, where an unusual and highly productive community recently has been described (IUCN, 1983c).

Along parts of the Dhofar coast of Southern Oman a dense algal community is dominated intertidally by thick growth of the large green *Ulva fasciata*, and subtidally by the kelp *Ecklonia radiata*, together with a similar hitherto undescribed plant, probably from a new genus allied to *Sargassum*. Associated with these larger algae are ninety or more 'understorey' species, of which the most conspicuous is *Amphiroa anceps*. It is possible that other areas of kelp forest, or at least of dense algal growth, may occur on parts of the coast of Somalia or of South Yemen, as these also lie adjacent to the southern Arabian upwelling.

The occurrence of the kelp forest in Dhofar is associated with the intense upwelling of water in southern Arabia. This upwelling occurs seasonally during the southwest monsoon when, as a result of the southwest winds, the surface waters are carried in a southeasterly direction, away from the coast, and are replaced by the upwelling of water from below (SMITH, 1968; BOTTERO, 1969; CURRIE et al., 1973). According to calculations by BOTTERO (1969), the upwelling extends at least 400 kilometers offshore and parallels the coastline for a distance of over 1,000 kilometers; because of the breadth of the upwelling zone, water is supplied to the surface from greater depths than is usual in other coastal upwelling regions. However, as the upwelling is seasonal, luxuriant growth of the kelp and of the other algae occurs during the monsoon period (June to September), but decreases rapidly following the end of the monsoon.

Based on studies in other regions, when kelp forests occur in situations where nutrient supply, illumination and temperature are optimal, they can maintain high productivities of 1,500-3,000 g C m^{-2} yr^{-1}, comparable with the most productive of aquatic or terrestrial ecosystems. Not only does this productivity sustain large populations of a great variety of invertebrates and fishes, but the plants themselves create an environment within which many find shelter. In Oman over 200 species of molluscs and 120 species of crustacea have been recorded in association with the algal communities (IUCN, 1983c). Some of the plant material is consumed directly by herbivores, notably the sea urchins *Echinometra mathaei* and *Stomopneustes variolaris*, and the abalone, *Haliotis mariae*. But the bulk of the kelp production enters the food web via the detritovore food chain. The fronds of the kelp and other algae are continually eroded by wave action, and the detritus thus formed is utilised by bacteria, zooplankton and filter feeding invertebrates, which in turn support larger invertebrates and fish, including many of the species of local commercial significance.

LAGOONS AND MUD FLATS

A third additional habitat which also may be comparatively productive, and thus possibly critical to the maintenance of coastal stocks of fish and other marine life, occurs in very shallow soft-bottom areas, in coastal lagoons or in sheltered areas, even in the absence of seagrass beds or mangrove. In such areas, sand-mud bottoms and mud flats can have a high secondary productivity. The finer particles of silt and detritus that settle out in such areas may support large populations of detritivorous crustacea and gastropods, and suspension and deposit feeding bivalves and polychaetes. This is particularly so on the estuarine mud flats which occur around the mouths of many temperate rivers. Frequently in the tropics the secondary productivity of such sand-mud bottoms and mud flats is supported or enhanced by the primary productivity of adjacent seagrass beds or mangrove detritus that is deposited in low wave-energy areas. Such areas with a very high secondary productivity may be critical in that they are frequently the nursery grounds for various fish and crustacea; in addition they often provide vital feeding grounds for migrant or over-wintering populations of wading and shore birds. The extensive mud flats of Kuwait, for example, provide a critical feeding and staging site for hundreds of thousands of waders migrating between their breeding grounds in northern Eurasia and wintering areas in southern and eastern Africa.

It is less widely appreciated, however, that in the tropics in particular intertidal mud and shallow water sand-mud may show a significant *primary* productivity as a result of the growth of an epibenthic algal lawn, i.e. a thin film of diatoms and blue-green and filamentous green algae. For example, in a recent study in Bahrain (IUCN, 1983b), an intertidal mud flat on the east coast had a chlorophyll concentration due to the epibenthic algal film of 0.16-0.24 g m^{-2}, a value that approaches that typical of oceanic upwelling zones (typically about 0.3 g m^{-2}) and is much higher than that of most tropical ocean areas (of the order of 0.03 g m^{-2}). In fact the primary productivity per unit area of the mud flat was estimated to be approximately fifty times higher than that of the surrounding sea where phytoplankton growth was limited by the high turbidity, high salinity and limited depth. This algal film on the mud flat help support large numbers of amphipods and a variety of gastropods, such as *Monilea obscura, Rhinoclavis sordidula* and *Mitrella blanda.*

Such epibenthic algal films also can occur over mixed coarser sand and/or mud-sand bottoms in shallow subtidal areas up to 1 to 2 m deep, that are sufficiently protected for the bottom not to be disturbed by wave action. Thus such films are especially observed in coastal lagoons behind bars and barrier islands, in fringing reef lagoons, and within suitable areas of the long creeks (known locally as *sharms* or *mersas*) that are common on both sides of the Red Sea.

However, the primary and secondary productivity of coastal lagoons and enclosed bays and mersas throughout the Indian Ocean is extremely variable. This is so because of the extremes of salinity which may occur along many more enclosed shorelines. In some areas rivers and wadis may flow seasonally into semi-enclosed bays or lagoons, suddenly converting a marine habitat into a brackish or fresh-water one; few organisms can withstand such complete changes in salinity. And, especially within the more arid regions of the Gulf and Red Sea, many semi-enclosed bays and mersas have elevated salinities and temperatures as a result of a hot climate and high rates of thermo-evaporation.

Thus a recent study in the Red Sea (IUCN, 1984) found that in moderately shallow water (1-3 m deep) to the sides of broader mersas, but not particularly enclosed and with some circulation of water, salinities can reach 45-48 ppt, compared to 39-40 ppt in the open sea. In very shallow, more enclosed areas at the innermost parts of mersas salinities up to 50-60 ppt were recorded; the highest figure obtained was 79 ppt in a rapidly evaporating pond almost isolated from the main part of a mersa. The most saline lagoon areas were completely devoid of marine life but in more equitable areas a variety of fish and invertebrates were common. The invertebrates included mostly crustacea and molluscs, among which cerithid gastropods of six or more different species of *Cerithium* and *Rhinoclavis* were the most abundant, occurring in densities of up to 100 m^{-2}, and feeding directly on the algal film. Other common invertebrates included herbivorous gastropods of the genera *Littorina* and *Strombus*, several predatory gastropods, various crabs, and most significantly juveniles of three species of commercial shrimp, of which the most abundant was *Penaeus semisulcatus*. Such bays and mersas are known to serve as nursery areas for these species of shrimp (BRANFORD, 1981), and probably do so in many areas of the Indian Ocean.

Amongst fish the tiny *Aphanias dispar* is ubiquitous and very abundant in such lagoons in many parts of the Indian Ocean, while commercial species characteristic of such habitats include especially mojarra (Gerridae), silversides (Atherinidae), grey mullet (Mugilidae) and juvenile emperors (Lethrinidae). As concluded by BARNES (1980), many of the most abundant lagoonal species appear to consume detritus, benthic algae and epiphytes rather indiscriminately, a generalisation that applies not only to the fish but also to most molluscs, annelids and nektonic prawns.

The significance of hypersaline conditions in limiting both primary and secondary productivity also has been demonstrated in several other recent studies. PRICE (1982) described that within the Arabian Gulf, the abundance of zooplankton is lower within embayments and is inversely correlated with salinity, and KAPETSKY (1981) has described how in Bardawil lagoon (on the Egyptian Mediterranean coast) amelioration of hypersaline conditions, through maintaining continuous connections with the sea, led to an increase in fish stocks.

In general, however, there appear to have been few intensive studies of such enclosed bays and lagoons within the Indian Ocean, and in particular no extensive survey of their occurrence and significance throughout the region. Such survey work may be an important priority in view of the fact that such lagoon areas are especially subject to impact from various developments. They are especially liable to be affected by construction work because they can be potential harbours or boat channels, or are easily infilled; they are also attractive locations for the siting of private residences or commercial buildings, and serve as a natural focus for recreational developments. They are also particularly susceptible to pollution because of their reduced water circulation, so that pollutants discharged into them tend to concentrate there. Besides industrial discharges, private houses clustered around a creek tend to be built, in the absence of any regulations, so as to discharge their sewage and waste into the creek, a practice that easily can result in local eutrophication. The many impacts that can affect a single creek or lagoon have been documented in various studies (e.g. BELLAN, 1972).

CONCLUDING REMARKS

It is very apparent that critical habitats within the Indian Ocean are being subjected to an increasing variety and severity of impacts as a result of the increased pace of industrial and municipal development and the pressures of a greatly increased human population. Much of the damage to these habitats is direct damage or destruction to which these habitats are especially susceptible because of their coastal and near-shore shallow water locations. In fact, a recent review has concluded (UNEP, 1985) that at the present time the impact on Indian Ocean habitats and resources due to pollution is as yet generally limited and confined to local areas, whereas, by contrast, the loss and destruction of critical habitats is now so extensive that, combined with widespread overfishing, during the last ten to fifteen years the potential fisheries yield of the region may have decreased by as much as 20%. Given the increasing food shortages in countries bordering the Indian Ocean, it must be considered an urgent task to halt and reverse this acceleration in destruction of the region's renewable resources.

A first step in this direction, save in countries such as Indonesia and Saudi Arabia where such action is already being taken, must be a survey of the occurrence, extent and significance of each of these critical marine habitats. This may be accomplished for some habitats in association with regional projects sponsored by the international agencies (UNESCO, UNEP and IUCN). Such survey work must then lead to the development of an integrated coastal zone management plan. Through suitable coastal zoning, areas of productive habitat can be guarded against unnecessary destruction by planners or developers who may not realise their critical role in maintaining their country's renewable resources.

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RISING SEA LEVEL AND CHANGING SEDIMENT INFLUXES REAL AND FUTURE PROBLEM FOR INDIAN OCEAN COASTAL NATIONS

JOHN D. MILLIMAN
Woods Hole Oceanographic Institution
Woods Hole, MA 02543 USA

ABSTRACT

Within the next 100 years the combined effects of sea level rise, subsidence and the damming and diverting of major rivers can effect a shoreline regression of 5 to more than 20 km in many deltaic areas bordering the Indian Ocean. Oceanographers may have little control in many of the policy decisions regarding the rising sea level or the damming of rivers, but by documenting the coastal environments we can at least estimate the potential damage as well as suggest ways in which the impact could be lessened.

INTRODUCTION

Few areas are as dependent upon the sea and their coastal areas as the countries bordering the Indian Ocean. In addition to the transportation and mariculture normally associated with coastal states, these deltas often house large human populations. A quick look at a map of western Asia shows that the large cities of Karachi, Calcutta, Dacca, Rangoon and Bangkok all lie on or near major deltas. Moreover, the continually renewed river sediments on the delta form the source for particularly productive soils. The Bengal Delta, for example, is one of the most productive agricultural areas in Asia. Yet because of its low elevations, it also is one of the most exposed to the whims of the ocean. A slight rise in sea level an increased rate of coastal erosion could have a major adverse effect on the peoples in northwest India and Bangladesh. Similar arguments can be made for other deltas, both large and small, that border the rim of the Indian Ocean.

In the next century three events will occur, one natural and two man-made, which will affect many coastal areas around the world, but perhaps particularly the deltaic areas around the Indian Ocean. Indian Ocean nations will have no control over one of these processes (Basin Subsidence), some control over another (Rising Sea Level), and essentially complete control over the third (Damming and Diverting of Rivers). If all three events occur unchecked, the result will almost certainly be disastrous for many coastal areas. Research into these problems should begin immediately, so that the potential impact can be predicted and minimized.

BASIN SUBSIDENCE

Deltas subside at rates ranging from 1 mm to several cm per year; large deltas often subsiding at faster rates than smaller ones. This subsidence results partly from crustal processes (often caused by relative movement along continent-ocean crust boundaries) and partly reflects isostatic compensation and consolidation of the sediment deposited by rivers. Assuming that the delta subsides at a rate of 3 mm/yr, over the next 100 years the delta would subside 30 cm; that is, relative sea level would rise by that amount if accumulation of incoming river sediment did not offset the subsidence. Removal of ground water or petroleum from the underlying strata can increase local subsidence by as much as an order of magnitude.

In deltatic areas, river-borne sediment will generally deposit at a rate equal to or greater than the rate of subsidence, such that the delta either maintains its position relative to sea level or progrades seaward. However, over the next century this natural pattern will change through the increasing damming and diverting of the rivers draining into the Indian Ocean. This problem is of particular concern for southern Asian and the adjacent oceanic islands: 75 percent of the river-derived sediment reaching the ocean on an annual basis comes from southern Asia and the island nations of Indonesia, Philippines, Taiwan, New Zealand and New Guinea (Fig. 1).

Southern Asia, with its mountainous terraine, torrential rainfall, geologically young rocks, and often poor efforts in soil conservation, contributes more sediment to the world ocean than the rest of the world combined. And given the long period of time in which there has been organized agriculture in Asia, such conditions probably have existed for thousands of years. While only one of the eleven largest drainage basins is located in southern Asia, 3 of the largest in terms of water discharge occur there, and 7 of the 11 largest in terms of sediment discharge (Table 1). The island of Taiwan, for example, contributes about 300 million tons of sediment annually to the ocean, not much less than the entire North American continent, which is three orders of magnitude larger in area. Extensive deltas have formed off many of large rivers — such as the Indus, Ganges/Brahmaputra, Irrawaddy, Pearl, and Yellow — and numerous smaller deltas have formed off smaller rivers.

With increased economic and political pressure, rivers in Asia and Africa have been and are being dammed for many purposes: hydroelectric energy, irrigation, flood control, etc. The Nassar Dam in Egypt, for example, increased the arable land along the Nile River by nearly 100 percent, increased dramatically the amount of available electric power, and practically eliminated the disastrous effects of spring floods on the lower portions of the river valley. Diverting or eliminating water and sediment discharge of a river, however, also can result in tremendous changes to the coastal environment. Yet as many tropical countires are seeking ways to utilize the hydroelectric and irrigation potential of their rivers, scientists and engineers from these countries often are unable to predict accurately the possible effects upon the coastal environment. Moreover, the decisions to dam a river often is made from an inland capital, where the country's ruler(s) may be particularly influenced by the needs of the inland inhabitants. One wonders, for instance, if the Aswan Dam in Egypt would have been built in the same way if the nation's capital had been in Alexandria (on the Mediterranean coast) rather than in Cairo, 200 km inland.

Some downstream and marine problems become apparent as soon as a river is dammed. For example, because Lake Nassar traps most of the Nile's sediment and regulates river flow, spring floods are smaller and little sediment is deposited on the flood plains, a process that for many millenia had renewed the Nile delta's agricultural lands. Perhaps less expected was the dramatic decrease in nearshore biological productivity after the damming of the nutrient-rich Nile river water, and with it, the collapse of some coastal marine fisheries. The sardine industry in Egypt, for example, reported a more than 95 percent decrease in catches within a few years after completion of the Aswan Dam (WAHBY and BISHARA, 1981).

The decrease in freshwater input also can drastically decrease both aquatic and terrestial productivity within estuarines and mangrove swamps. Mangrove swamps in coastal Pakistan have been used for many hundreds of years as a source of food and fuel by local communities. With the decreased flow of the Indus River following construction of upstream dams, the mangrove forests have mostly died, resulting in a corresponding demise of the wood-gathering culture.

In terms of biological impact, many of the negative effects (and the degree of their severity) may become apparent within a few years of river "shut-down". If the river begins to flow again (if the dam is modified or destroyed), the adverse effects *may* be corrected quickly. Such is not the case with geological effects, where the ultimate impact might not be felt until long after the initial damming, and the problem may become increasingly worse with time. The fate of the sediment brought to the ocean by a river depends on many factors, particularly the nature of the sediment (e.g., gravel, sand, silt, clay), the amount of sediment, and the geomorphic and oceanographic regime of the coastal environment. The amounts of sediment involved, particularly off large rivers, can be impressive. For instance, the annual load of the Ganges-Brahmaputra River (1.7 billion tons) equals more than 1 km^3

of unconsolidated sediment (assuming about 55 percent water) deposited yearly. Imagine what would happen to the nearshore environment if this volume of sediment were cut off?

One example of a river that has been recently diverted and dammed is the Indus River, the largest river emptying into the western part of the Indian Ocean, but the effects of its effects on the coastal environment are still only poorly understood. Until the mid-1940's, about 250 million tons of Indus River sediment reached the Indus Delta. Beginning with wide-spread construction of barrages in the lower parts of the river valley in the mid and late 1940's, the sediment load dropped to less than 50 million tons (Fig. 2). Since then, the annual load has averaged substantially less than 100 million tons except for several years of large floods. Water discharge from the Indus, in contrast, remained high until construction of upstream dams in the early 1960's, after which it dropped from more than 100 km^3/yr to less than 60 km^3/yr. As hydroelectric and irrigation projects increase in the coming years, sediment and water discharge will undoubtedly decrease even further. The impact of diminished river flow upon the mangrove ecosystem and the marine fisheries has been towards decreased fish yield, probably in part because of construction of the Kotri barrage in 1955 and in part because of dam construction in the early 1960's (Fig. 3). As of yet, the effect of decreased sediment load upon the coastal environment has not been documented, but the effects should be apparent and dramatic. The Indus Delta is exposed to extremely high wave energy during the summer monsoon season. This should cause large-scale erosion of the delta. The impact also will be felt ultimately in northwest India, where coastal currents historically have transported Indus-derived sediment. It is possible that the coast east of the Indus River is already eroding, but the area generally is unpopulated so that human impact so far probably has been small.

RISING SEA LEVEL

The first two events discussed are primarily local events, related to local basin subsidence and to river flow. The third event, and the one that may have the greatest impact upon coastal states, is the world-wide rise of sea level. With the end of the last glacial epoch about 15 thousand years ago, world sea level rose about 100 meters, in response to the influx of glacial melt water to the world ocean. Sea level more or less reached its modern level about 5000 years ago, and since then has risen only a fraction of a mm per year. This natural situation changed, however, with the increased burning of fossil fuels in the early part of this century — initiating the so-called greenhouse effect.

Presently sea level is (and has been) rising at about 1 mm/year, and various scenarios predict total rises ranging from 25 cm to more than 2 meters by the end of the next century (HOFFMAN et al., 1983). This effect will be heightened by the use of chlorofluorocarbons within the world community (J. TITUS, 1985, oral communication). When added to regional subsidence, it is possible that Indian Ocean deltaic areas may experience a *relative* subsidence of several meters or more. With the very low gradients present on many deltas this would mean that coastal regression could be extensive in many areas by the 22nd century. If one adds in the possible coastal erosion associated with the decreased sediment influx from dammed or diverted rivers, the coastal regression may exceed several tens of kilometers.

Coastal erosion and regression, of course, are not the only problems. Rising sea level, when combined with decreased river flow, will mean a landward migration of the salt water wedge within the ground water, which could severely affect the quality of drinking and agricultural waters. [Recent damming of the Ganges River, for instance, has meant more than an order of magnitude increase in salinity of groundwater within the Bangladesh delta (RAHMAN KHAN, 1983)]. Salt marshes and mangrove areas will be destroyed, thus adversely affecting the spawning areas of many fishes. Finally, the number of tropical storms may increase significantly with the slight climate change associated with rising sea level (D.G. AUBREY, 1985, oral communication), and thus furthering the possibility of catastrophic coastal flooding in low-lying areas.

FUTURE STUDIES

What can marine scientists do to document and hopefully change this possible disaster? In terms of altering the policy decisions regarding the influx of fossil CO_2 to the atmosphere, we probably can do little. We may have somewhat greater influence on the building of dams. However, given the nature of the problem, we can document factors which can ultimately quantify the effects that these human activities will have on the deltas. Such a program involves basic research with practical goals. Some of these are as follows:

1) Detailed analysis of tide gauge records can show the relative rise of sea level in the deltaic areas. If wave gauge stations do not exist in these areas, then they should be installed;
2) We need to document the flow patterns and sediment loads of the rivers (particularly before and after damming);
3) We need to understand more completely (and more quantitatively) the wave and current regimes in deltaic areas. By careful comparison of old maps and satellite photographs, we should be able to determine recent erosion and accretion patterns on the deltas, and therefore predict which areas might be impacted particularly severely in the next 50 years.

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Table 1. Ranking of the world's major rivers in terms of drainage area, discharge and sediment load. Data from MILLIMAN and MEADE (1983). Asterisks (*) denote Asian rivers.

A. Drainage Basin Area

River	x $10^6 km^2$
1. Amazon	6.15
2. Zaire	3.82
3. Mississippi	3.27
4. (Nile)	(2.96)
5. La Plata	2.83
6. Yenissei	2.58
7. Ob	2.50
8. Lena	2.49
*9. Changjiang	1.94
10. Amur	1.85
11. MacKenzie	1.81
Total	29.24 (29% of world total)

B. Average Annual Water Discharge

River	km^3/yr
1. Amazon	6300
2. Zaire	1250
3. Orinoco	1100
* 4. Ganges/Brah.	971
* 5. Changjiang	900
6. Mississippi	580
7. Yenissei	560
8. Lena	514
9. Plata	470
* 10. Mekong	470
Total	13115 (32% of world total)

C. Average Annual Suspended Load

River	x $10^6 t/yr$
* 1. Ganges/Brah.	1670
* 2. Huanghe	1080
3. Amazon	900
* 4. Changjiang	478
* 5. Irrawaddy	285
* 6. (Indus)	(250)
7. Magdalena	220
8. Mississippi	210
9. Orinoco	210
* 10. Hunghe	160
* 11. Mekong	160
Total	5373 (40% of world total)

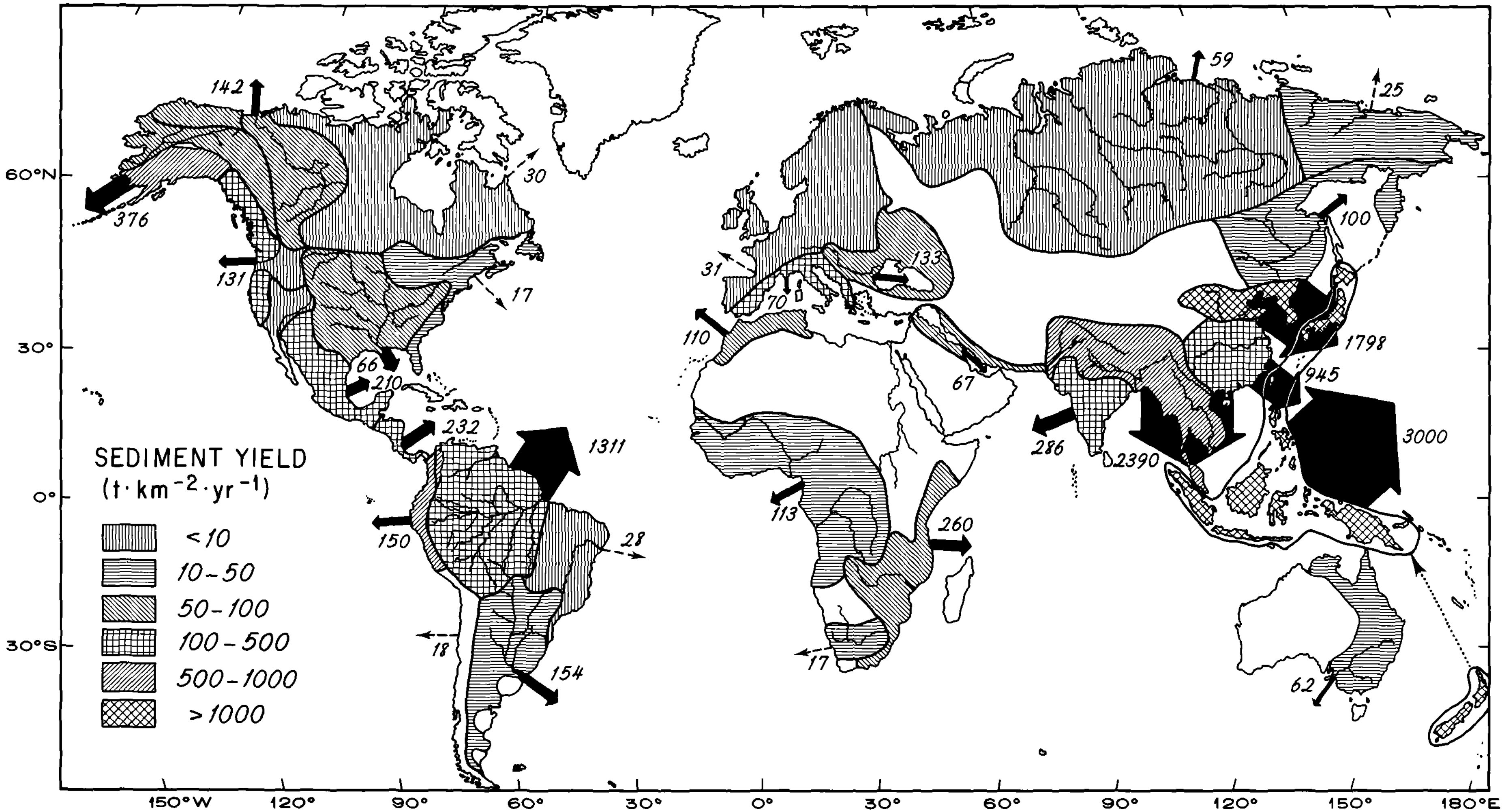

Figure 1. Average annual sediment influx to the oceans from the major river basins of the world. Numbers are in millions of tons per year, and the arrows are proportional to these numbers. From MILLIMAN and MEADE (1983).

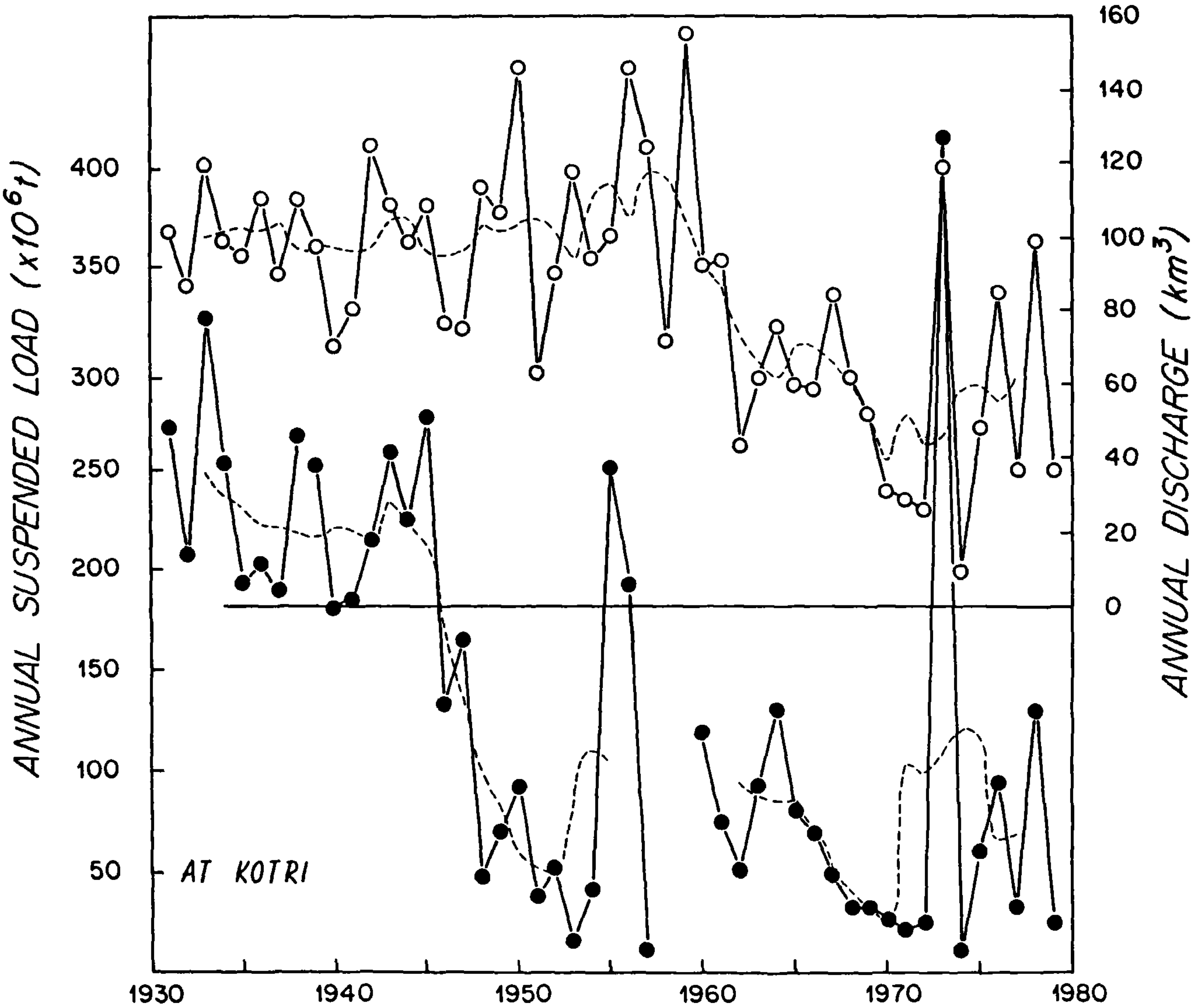

Figure 2. Discharge and sediment loads for the Indus River at Kotri (about 250 km from the Indus mouth) for the years 1931–1979. Dashed lines represent 5-year running mean averages. The decrease in sediment load in the late 1940's was in response to construction of irrigation barrages in the lower reaches of the river. The decrease in water discharge in the early 1960's resulted from construction of hydroelectric dams in the upper reaches of the river. From MILLIMAN *et al.* (1984).

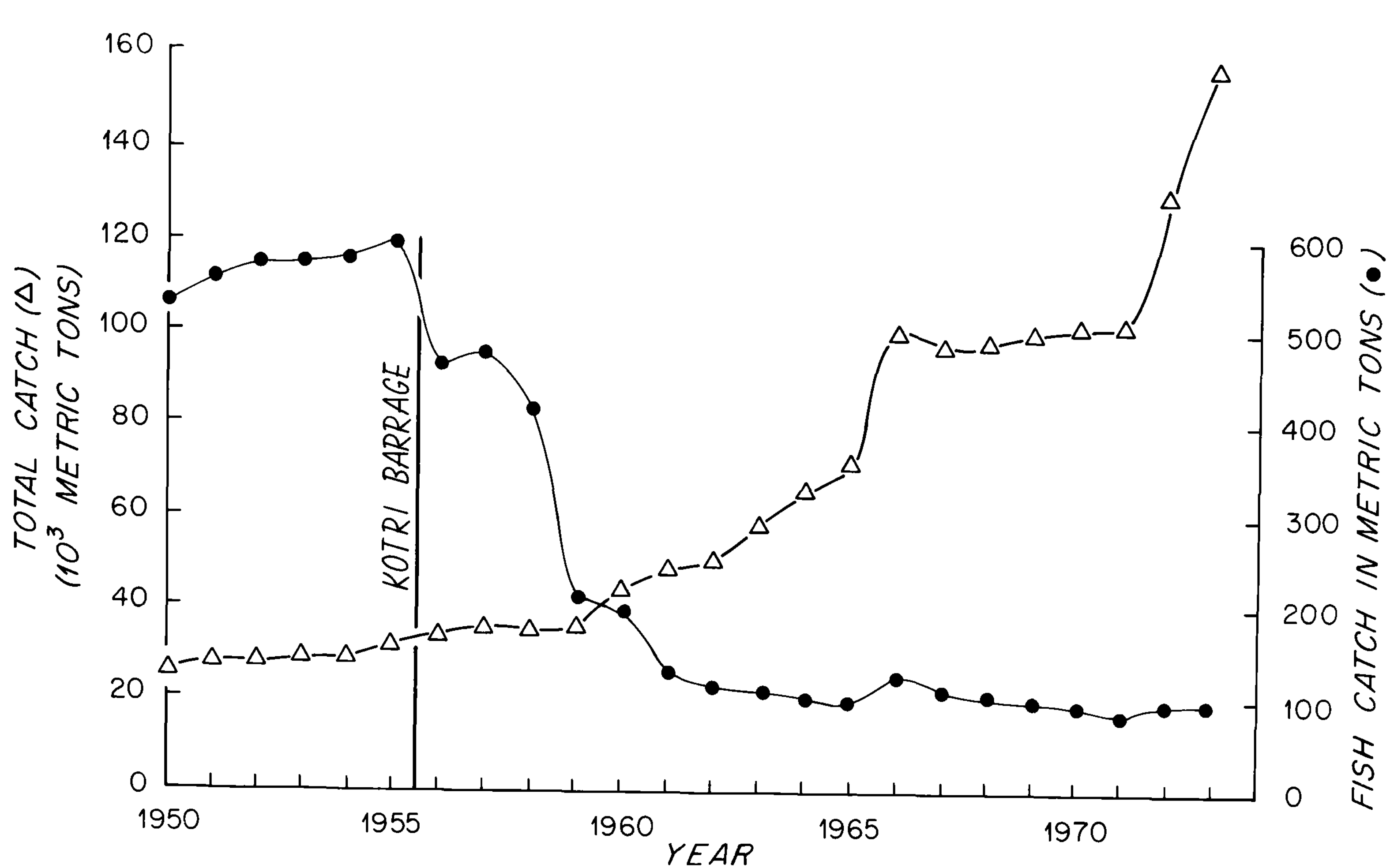

Figure 3. Yearly variations of total fish catch and fish catch per boat along the Sind coast. With the construction of the Kotri barrage (1955) there was a marked decrease in the catch per boat. The increase in total catch resulted from a increase in number of boats that fished the area. From MILLIMAN et al. (1984).

MARGINAL SEAS

STORM SURGES IN THE BAY OF BENGAL*

T.S. MURTY[1], R.A. FLATHER[2] AND R.F. HENRY[1]

[1]Institute of Ocean Sciences
Sidney, B.C., Canada

[2] Institute of Oceanographic Sciences
Bidston, U.K.

INTRODUCTION

Coastal floods associated with storm surges - the changes in water level generated by storms passing over the sea - surpass even earthquakes for loss of life and property damage (SHAH, 1983). Nowhere are these losses more serious than in the Bay of Bengal (Fig. 1). About 60% of all deaths due to storm surges have occurred in the low-lying arable coastal areas of the countries bordering the Bay and the adjoining Andaman Sea, with Bangladesh alone accounting for about 40% (MURTY, 1984). The steady increase of population continuously increases the potential for disaster.

The surges are caused by tropical cyclones which develop in the southeast part of the Bay or in the Andaman Sea at certain times of the year. The cyclones move along tracks that frequently cross the continental shelves bordering the Bay. A storm surge is generated partly by the resulting variations in atmospheric pressure, but the main contribution is produced directly by the winds, often of exceptional strength, acting over shallow water.

A comprehensive list of all Bay of Bengal surges for which some indication of surge severity is available is given in Table 2. By referring to Figure 1, it can be seen just how much coastline is vulnerable. Thailand is exceptional in that there is no record of its west coast, facing the Andaman Sea, having experienced a major surge. Storm surges are much less frequent in Sri Lanka than in Bangladesh, India or Burma.

While immediate loss of life is often the most dramatic result of a major surge, other effects also can be very serious. The most devastating surge in this century was that which struck Bangladesh in November 1970 (Surge No. 49 in Table 2). Human deaths caused directly by the surge were estimated at 200,000 to 300,000, but there was also enormous economic damage which must have led to more deaths later. Much of the coastal area of Bangladesh was flooded by sea water, almost all livestock in the affected area was drowned, and most of the fishing fleet was destroyed (FRANK and HUSSAIN, 1971). Damage due to storm surges in the Bay of Bengal region in the period 1945-75 has been estimated at 7000 million U.S. dollars (MURTY, 1984), but this scarcely expresses the impact of such disasters on developing countries.

A number of general reviews and descriptions of individual cyclones and the associated surges have been published previously; see, for example, ELIOT (1900), DUNN (1962), RATNAM and NAYAR (1966), and the Bay of Bengal Pilot (ANON, 1966). Notably, FRANK and HUSSAIN (1971) discussed the disastrous effects of the November 1970 storm surge in Bangladesh; COHEN and RAGHAVULU (1979) gave a vivid description of the impact of the 1977 Andhra surge; and DAS et al. (1978) provided a general review.

This paper is intended as an introduction to and a broad review of major aspects of the Bay of Bengal storm surge problem. An account of the factors affecting the generation of storm surges in the Bay and their consequences also is presented. During the last few years, efforts have concentrated in the dynamical approach, in which surges are computed by solving the governing equations to determine the response of the sea to a model cyclone moving over its surface. Advances made in such numerical models are discussed elsewhere (MURTY et al., 1986); the present paper, in fact, represents a condensation of that earlier paper.

* Modified from a paper in *Progress in Oceanography*, 16, 195-233.

FACTORS RELATING TO SURGE GENERATION

Storm surges are atmospherically-forced oscillations of water level in a coastal or inland water body, in the period range of a few minutes to a few days, driven by pressure gradients and wind stresses associated with travelling weather systems. By this definition, storm surges are distinct from wind waves and swell, which have periods of the order of a few to several seconds. Although storm surges belong to the same class of wave as tides and tsunamis, that is long gravity waves, there are at least two important differences. First, whereas tides and tsunamis can occur on an oceanic scale, storm surges are simply a coastal phenomenon. Second, significant tsunamis and tides do not occur in a completely closed small coastal or inland water body, but storm surges can occur even in completely enclosed lakes, canals or rivers.

One may ask why storm surges with different amplitudes occur at different locations in the same water body. The answer is that surge height depends on the topography of the water body near the location under discussion, and also the position of this place relative to the track of the storm that generates the surge. Shallow areas of water bodies generally experience surges with larger amplitudes.

THE BAY OF BENGAL

The Bay of Bengal is essentially a northward extension of the Indian Ocean, with depths decreasing from 4000 m in the south to 2000 m in the north. The Andaman and Nicobar Islands form an almost continuous barrier that divides the eastern part of the Bay (referred to as the Andaman Sea) into almost a separate water body, at least from the point of view of tides. Coastal geometry and bottom topography are important in determining the surge response on the open coast (FLIERL and ROBINSON, 1972). The width of the shelf is fairly small, typically 35 km for both eastern and western margins of the Bay. However, at the head of the Bay the Ganges delta forms a much wider shelf, typically 175 km wide. This shelf effect, combined with the funnel-like configuration of the coastline, amplifies the tides and surges at the head of the Bay, especially on the Bangladesh coast. The only significant morphological feature of the shelf at the head of the Bay is the "Swatch of No Ground". This canyon, about 150 km long, 22 km wide, over 1000 m deep at its southern end and some 270 m deep at its northern end, extends in a north-eastward direction to a point south of the Pussur River estuary, and obviously must influence tide and surge propagation into the Pussur River area.

Coastal orography is another factor which greatly increases the susceptibility of Bangladesh to storm surge inundation. Most of Bangladesh consists of delta lands built up by the distributaries of the Ganges, mostly less than 5 m above sea level. This very flat terrain allows rivers to change their course rapidly and frequently, and also allows coastal erosion to exceed 250 m per year in some places. Similarly rapid deposition can form new islands ("chars") almost overnight. Such areas are soon densely populated, posing a high disaster potential.

The large earthquake (Richter magnitude about 8.5) of August 15, 1950 in Assam state in northeast India caused major landslides. Millions of tons of soil clogged the tributaries of the Brahmaputra River and ultimately worked its way into other rivers and estuaries and helped to form new deltas. Water depths in some rivers were reduced by as much as 4.5 m, and consequently neighbouring areas became much more vulnerable to flooding from storm surges than before.

SURGE GENERATION

The coasts of Bangladesh and Burma and the Bay of Bengal coast of India have been subjected to frequent severe storm surges (Table 2). Meteorologically speaking, the Bay of Bengal is a breeding ground for tropical storms and depressions, the yearly average from 1890 to 1969 being 13 (RAGHAVENDRA, 1973). MOWLA (1968) studied 413 depressions and cyclones (see Table 3 for a definition of terms) in the Bay of Bengal for the period 1924 to 1952, and found that a warm pool of air usually forms at the 200 mb level, 3 to 4 days before the advent of cyclogenesis. The associated low pressure centres over the Bay intensify into depressions, some of which intensify further into cyclonic storms. A few of these develop into severe cyclonic storms of hurricane intensity. Tropical cyclones capable of generating surges usually occur during the pre-monsoon months (April to May) or the post-monsoon months (Sepember to December), but rarely during the monsoon seasons (June to August and January to March).

Low pressure areas mainly form in the Andaman Sea (Figs. 1-3). Occasionally lows that are remnants of typhoons from the China Sea move into the Bay from the east (MOOLEY, 1980).

MOOLEY (1980) estimated that the probability of a severe storm striking the coast of the Bay of Bengal in any year is 0.345. More recently MOOLEY and MOHILE (1983) identified significant trends in storm frequency, notably a significant increase in the mean annual frequency of storms on the Bangladesh coast during the period 1965-1980. RAO (1968) suggested that for the pre-monsoon storms (April to May) the strongest winds are usually in the southeast or eastern sector, whereas for the post-monsoon storms (Sepember to December) the region of maximum winds is to the north of the storm centre.

In Burma, storms occur mainly in the pre-monsoon season and affect the Arakan coast. If the storm occurs during 1-10 May, the landfall is north of 18°N. During the post-monsoon season the storms landfall north of 18°N.

Consider now the sequence of events as a typical storm approaches the coastline. In view of the water depth, atmospheric pressure effects can dominate over the deeper part of the Bay. Disturbances generated there propagate much more quickly than the speed at which the cyclone moves, and therefore reach the coasts ahead of the storm. Wind-generated currents (ANON, 1966) can influence sea level where they impinge on the continental shelf. However, the main surge effect is generated by the wind acting over the continental shelves. RATNAM and NAYAR (1966) suggested that the resonant coupling mechanism, shown by PROUDMAN (1929) to occur when the speed of a travelling atmospheric disturbance matches that of free gravity waves, could be significant in the shallow coastal waters of the Bay of Bengal. However, wind stress acting over shallow water is the main generating mechanism.

The maximum onshore winds, and hence also the maximum surge levels, tend to occur to the right of the storm track as it crosses the coast. A negative surge associated with offshore winds on the left of the storm centre also may appear. The surge maximum occurs about the time the storm centre crosses the coast. The speed and angle of approach of the storm, relative to the coastline, influence the magnitude of surge, and convergence of coastline in a bay or estuary can produce substantial amplification.

The extreme shallowness of the water in some near-shore areas, the focussing effect of inlets and estuaries and the absence of shore defences to protect the low-lying coastal lands may lead to the formation of a 'storm wave', frequently described by eye-witnesses as a 'wall of water', which can travel several kilometres inland, flooding affected areas to a depth of as much as 13 m (see the account of the Backergunge cyclone of October 1876 in ELIOT, 1900). Some accounts suggest that the accumulating water associated with the surge may be held back by the river flow and ebbing tide in estuaries such as the Hooghly, eventually advancing upstream as a bore, overtopping the river banks and inundating low-lying land as a storm wave (e.g. the Calcutta cyclone of October 1864, ELIOT, 1900). Clearly, nonlinear effects play a vital part in this stage in the development of the surge.

In the northern reaches of the Bay of Bengal and in the Andaman Sea, the tidal range is comparable with probable surge amplitudes. The relative timing of high tide and peak surge then becomes very important. For instance, in the disastrous inundation of Bangladesh in November 1970, the surge peak occurred at high tide and killed more than 200,000 people. In contrast, a comparable surge on October 31, 1960 (No. 22, Table 2) occurred at low tide and caused far fewer casualties. The apparent importance of nonlinear effects suggests that dynamical tide-surge interaction must also play a part here.

RAO (1968) classified the Indian and Bangladesh coast along the Bay of Bengal coast into three categories based on the combined amplitude of the storm surge and wind waves (Table 4, Fig. 4). Rao's C-type belt, which is the most dangerous zone, occurs in three locations:

(1) The coastal belt around the head of the Bay of Bengal, approximately to the north of 20°N. The frequency of cyclones is high, and storm tracks are usually favorable for generating maximum surges, especially in the Sunderbans.

(2) The south Coromandel coast around Balk Bay. Although the frequency of storms striking this region is somewhat smaller than for the first belt, the major storms that strike this coast usually produce major surges.

(3) A short C-type belt occurs Nizampatnam Bay; November 1977 storm surge here killed over 10,000 people.

The east coast of India between 14 N and 16.5 N and the Coromandel coast between Point Calimere and Karikal fall into the B-category, the next most dangerous zones.

SUGGESTIONS FOR FUTURE WORK

Storm surges have killed hundreds of thousands of people and caused extensive economic damage around the Bay of Bengal in historic times. Similar unrecorded events must have happened earlier, but the losses probably have been highest in this century as increasing population pressure has forced cultivation and habitation of newly-formed low-lying delta land. This situation will continue and even greater storm surge disasters must be anticipated, since economic and technical considerations render the construction of adequate coastal defences unfeasible. The only practical counter-measures evident for the foreseeable future are improved surge warning systems and construction of raised refuges a few kilometers apart in vulnerable areas. Much more intense study of recent and future surges will be essential in order to develop reliable warning systems and to forecast flood heights for refuge design purposes.

The need for considerable accuracy in public warning systems is well established. Unnecessary evacuations following erroneous warnings are not only costly, but inevitably cause the public to ignore subsequent warnings of genuine danger (WHITE and HASS, 1975). Surges occurring in the Bay of Bengal often affect relatively short stretches of the coast, but the area likely to be most affected is hard to predict even a few hours in advance due to our present lack of knowledge of factors governing cyclone tracks, speed of cyclone movement and surface stress fields under cyclones, to name only a few of the meteorological variabilities. On the oceanographic side, a major difficulty lies in the interaction of surge and tide, a thorough understanding of both being necessary for accurate forecasting of the magnitude and time of peak water level.

Numerical models offer the best prediction methods in the near and medium term. Far too few past surges have been recorded adequately enough to permit statistical or empirical surge forecasting, particularly in cases where tidal state/stage is significant. At present, in fact, there are even too few records to allow verification of numerical models or other prediction techniques. It is this aspect of the problem in which progress could probably be made most rapidly at reasonable cost. MURTY and HENRY (1983) proposed that an international effort be organised to collect and publish complete dossiers of relevant data on future major surges in the Bay of Bengal. On average, about four large surges occur per year, but since each affects only a small portion of the vulnerable 4000 km of coastline, this data-gathering scheme would have to be pursued for at least one or two decades to accumulate an adequate body of verification data. For moderate annual cost, a program of this type could provide a better basis for improved surge-forecasting capability than any short-term effort, no matter how heavily funded.

Collecting data on a storm surge requires a fair degree of prior organisation, since most local services are disrupted or destroyed in the inundated area. Complete meteorological records are required for several days before and sometimes after the time of cyclone landfall. Aerial or satellite photographs can show the extent of inundation together with confirmatory ground observations. In this connection, valid topographic maps also should be collected, and new surveys may have to be conducted if a surge causes significant changes. Water level records during surge episodes pose one of the most serious difficulties, due to the relatively sparse network of permanent tide gauges around the Bay and the fact that the types of gauges in use often fail to record water levels much beyond their normal working range. Accurate observation of flood marks can nevertheless give highly useful information. For instance, flood marks inside buildings should give a good estimate of water levels exclusive of wind waves.

All available relevant data for a surge should be assembled on microfilm or some other medium that permits convenient distribution to interested researchers and ensures safe preservation of the data. It is not extravagant to think that records collected carefully in our time could be immensely useful to surge forecasters a century or more from now.

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Table 1. Latitude and longitude of places mentioned in the text but not shown in Figure 1.

Name		Latitude (deg North)	Longitude (deg East)
	BANGLADESH		
BARISAL		22°42'	90°22'
CHANDPUR		23°13'	90°39'
COMAPANIGANJ		23°53'	91°17'
COX' BAZAR		21°26'	91°59'
GALACHIPA		22°07'	90°24'
HATIA		22°30'	91°15'
NOAKHALI		22°49'	91°06'
SANDWIP		22°30'	91°25'
	BURMA		
AKYAB		20°08'	92°54'
BASSEIN		16°47'	94°47'
GWA		17°35'	94°34'
KYAUKPYU		19°26'	93°33'
SANDOWAY		18°25'	94°28'
	INDIA		
ADIRAMPATNAM		10°23'	79°23'
BARUVA		18°52'	84°36'
CHANDBALI		20°47'	86°44'
CHIRALA		15°49'	80°21'
CONTAI		21°47'	87°45'
DEVIPATNAM		09°29'	78°55'
DHANUSHKODI		09°11'	79°24'
FALSE POINT		20°25'	86°47'
COPALPUR		19°16'	84°55'
KAKINADA		16°56'	82°15'
KALINGAPATNAM		18°21'	84°08'
KARIKAL		10°50'	79°50'
KAVALI		14°55'	79°59'
KORINGA		18°50'	80°30'
MASULIPATNAM		16°08'	81°08'
MIDNAPORE		22°25'	87°20'
NAGAPATTINAM		10°46'	79°51'
NARSAPUR		16°27'	81``45'
NELLORE		14°25'	80°00'
ONGOLE		15°31'	80°04'
PAMBAN		09°16'	79°12'
PARADIP		20°16'	86°41'
POINT CALIMERE		10°18'	79°52'
PURI		19°45'	85°50'
RAMESWARAM		09°17'	79°18'
SACRAMENTO SHOAL		16°36'	82°19'
VISAKHAPATNAM		17°41'	83°17'

Table 2 Partial List of Storm Surges in the Bay of Bengal

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
1	7 to 11 October 1737	Landfall at mouth of Hoogly River, India	Area Affected: Sunderbans area of Bangladesh, west Bengal state of India. Total water level 12 m (probably including tide and wind waves).	300,000 deaths in India and Bangladesh
2	1779	Masulipatnam, India		20,000 people killed
3	December 1789	Landfall at Kakinada, Andhrapradesh, India		20,000 deaths in India
4	1789	Bengal		20,000 killed
5	1822			40,000 deaths in Bangladesh
6	1833			50,000 deaths in India
7	1839			20,000 deaths in India
8	October 1854	Landfall at mouth of Hoogly River, India	Area Affected: Calcutta and environs. Total water level of 12 m.	50,000 deaths in India
9	November 1864	Landfall near Masulipatnam, Andhrapradesh, India.	Area Affected: Masulipatnam and environs (Andhrapradesh)	40,000 deaths in India
10	1864	Bengal		100,000 killed
11	27 October to 1 November 1876		This was the infamous Bakergunj Cyclone. Peak water levels in Bangladesh varied from 3 to 15 m, including tide and wind waves.	100,000 deaths in Bangladesh (then part of India). The India Meteorological Dept. was expanded following this surge. Some estimates, for example, Gill (1975) put the death toll at 400,000, but this could be an overestimate.

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
12	September 1895	Landfall at False Point, India.	Area Affected: North Orissa Coast of India. Peak water levels of 7 m, including tide.	More than 5,000 deaths in India.
13	1897			175,000 deaths in Bangladesh
14	1912			40,000 deaths in Bangladesh
15	20 to 25 September 1919			40,000 deaths in Bangladesh
16	May 1926			More than 5,000 deaths in India
17	1 November 1927	Landfall at Nellore, Andhrapradesh, India	Area Affected: Andrapradesh, India The town of Nellore destroyed.	300 deaths in India
18	May 1941			More than 5,000 deaths in India
19	October 1942	Landfall in West Sunderbans, India.	Area Affected: West Bengal state of India. 5 m surge at Midnapore	40,000 deaths in West Bengal State of India
20	28 October 1949	Landfall at Masulipatnam, Andhrapradesh, India. Central pressure - 955 mb; pressure deficit - 55 mb; max. wind speed - 185 kph.	Peak surge at Masulipatnam was 2.5 m. Estimated tide at time of surge was 0.5 m. Other places varied from 3.0 to 5.0 m. Surge penetrated 14.5 km inland.	800 deaths (Cohen and Raghavulu, 1979)

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
21	November 1952	Landfall at Nagapattinam, Tamilnadu, India.	Area Affected: Coromandel coast and Northern shores of Palk Bay. Peak surge of 3 m	Several thousand deaths in India.
22	21 to 24 October 1958	Winds of 89 kph	Peak surge of 2 m in India	
23	29 May 1960	Landfall at Sunderban area of Bangladesh	Peak surge of 3.2 m. Estimated tide at time of peak surge - 0 m	
24	7 to 10 October 1960	Landfall at Noakhali (Meghna Estuary) Max. wind - 161 kph	Observed peak water level of 6.6 m and of this 1.5 m was tide and the rest was surge.	5,000 deaths in Bangladesh
25	30 to 31 October 1960	Landfall at Chittagong, Bangladesh Max. wind - 210 kph	Observed peak surge at Chittagong was 7.1 m (probably including tide). Peak water levels up to 8.8 m.	15,000 deaths in Bangladesh
26	6 to 9 May 1961	Landfall at Cox's Bazar, Bangladesh. Max. wind - 149 kph	Peak level of 7.5 m at Galachipa. (including tide and wind waves). (Estimated tide at time of peak surge was 1.2 m).	1,000 deaths in Bangladesh
27	27 to 30 May 1961	Max winds up to 145 kph	Peak water levels up to 9 m (including tide and wind waves).	10,466 deaths in Bangladesh
28	30 October 1962	Landfall near Chittagong	Peak surge of 5.8 m. Estimated tide at time of peak surge was 0 m	50,000 deaths in Bangladesh
29	25 to 29 May 1963	Landfall near Chittagong Max. wind - 201 kph	Peak water levels up to 8.1 m (including tide and wind waves). Estimated tide at time of peak surge was 0.3 m. Peak surge about 5 m	50,000 deaths in Bangladesh

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
30	8 June 1963	Landfall at Sunderbans, Bangladesh	Peak surge of 3.1 m. Estimated tide at time of peak surge was 0 m	
31	20 October 1963	Landfall at Chittagong, Bangladesh	Peak surge of 2.2 m near Chittagong. Estimated tide at time of peak surge was 0 m.	
32	11 April 1964			196 deaths in Bangladesh
33	17 to 24 December 1964	Landfall at Adiramapatnam, Tamilnadu, India. Max. winds - 193 kph; Min. central pressure - 970 mb.	Area Affected: South Coromandel coast and west shores of Palk Bay. 5 m surges at Dhanuskodi, India. Peak surges up to 6 m.	1,000 deaths in India
34	10 to 12 May 1965	Landfall between Barisal and Noakhali, Bangladesh Max. wind - 162 kph	At Sandwip tide plus surge was 2.9 m. At Noakhali tide plus surge was 5.2 m (of this 1.2 m was tide). Max. observed total water levels were 6.0 m (including tide and wind waves).	12,000 deaths in Bangladesh
35	31 May to 1 June 1965	Landfall near Chittagong	Tide plus surge was 7.1 m at Comapaniganj. At Chittagong 1.6 m surge on tide. Total observed water levels up to 8 m.	19,279 deaths in Bangladesh
36	11 to 15 December 1965	Landfall near Cox's Bazar. Max. wind - 209 kph: central pressure - 950 mb; pressure deficit of 60 mb.	At Chittagong peak surge 4.0 m, tide 0.2 m, observed total level 5 m. Observed level at Cox's Bazar was 8.8 m.	1,000 deaths in Bangladesh

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
37	1 October 1966	Landfall near Chittagong Max. wind - 145 kph	Peak surges up to 7.2 m and peak water levels up to 10 m (including wind waves).	850 deaths in Bangladesh
38	27 October to 1 November 1966	Max. wind - 145 kph	Peak water levels up to 10 m in Bangladesh.	
39	15 to 18 May 1967	Landfall between Akyab and Kyaukpyu, Burma. Pressure fell by 16 mb in one hour at Akyab. Central pressure at Akyab - 983 mb; max wind - 161 kph		100 deaths in Burma. Damage to property was 30 million kyat.
40	18 May 1967	Landfall near Cox's Bazar, Bangladesh.	Peak surge of 0.9 m at Chittagong.	
41	9 to 11 October 1967	Landfall at Puri, Orissa State, India. Central pressure - 970 mb; pressure deficit - 40 mb; max. wind - 167 kph.	Peak surge of 2.5 m; tide: 0 m. Total water level (including wind waves) 9 m.	
42	11 October 1967	Landfall at Noakhali, Bangladesh. Central pressure - 975 mb; pressure deficit - 35 mb; max. wind - 160 kph.	Peak water levels up to 9.3 m in Bangladesh. Tide plus surge at Comapaniganj was 8.7 m. At Noakhali surge was 3.0 m, tide: 0 m.	
43	20 to 24 October 1967	Landfall between Akyab and Chittagong; pressure fell by 29 mb in 1 hour at Akyab. Central pressure at Akyab - 986 mb; max. wind speed - 145 kph.	Peak water levels up to 7.5 m in Burma.	200 deaths in Burma. Damage of 100 million Kyat.

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
44	23 October 1967	Landfall near Chittagong; central pressure - 983 mb; pressure deficit - 27 mb; max. wind - 130 kph.	Tide plus surge at Comapaniganj was 7.6 m. At Cox's Bazar: surge: 2.2 m; tide: 0 m; at Chittagong, 2.0 m surge, 0 m tide.	
45	4 to 8 December 1967	Landfall near Nagpattinam, Tamilnadu, India. Central pressure - 988 mb; max. wind - 130 kph.	Moderate surges on the Tamilnadu coast of India.	7 deaths in India.
46	17 to 10 May 1968	Landfall at Akyab, Burma; pressure fell by 42 mb in 1.5 hours at Akyab. Lowest pressure 958 mb at Akyab; max. wind - 209 kph from NE. Central pressure of 940 mb in Bangladesh; pressure deficit of 70 mb; max. wind in Bangladesh - 222 kph.	Water level at Comapaniganj in Bangladesh was 4.7 m. Tide at time of peak surge: 0 m.	1,037 deaths in Burma. Damage was 20 million Kyat.
47	17 April 1969			75 deaths in Bangladesh.
48	10 October 1969		Peak water levels up to 8 m in Bangladesh; tide plus surge reached 7.8 m at Sandwip.	
49	4 to 9 November 1969	Landfall at Kakinada, Andhrapradesh, India. Central pressure - 968 mb; pressure deficit - 42 mb; max. wind - 176 kph.	Storm surges at Visakhapatnam and Koringa, India. Peak surge: 2.6 m; surge occurred at low tide; peak water level: 3.1 m (including wind waves).	200 deaths in India.

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
50	5 to 7 May 1970	Landfall at Cox's Bazar, Bangladesh. Central pressure - 977 mb; pressure deficit - 33 mb; wind - 148 kph.	Peak surge: 2.3 m; tide: 0.2 m; peak water level: 5.0 m (including wind waves).	
51	22 to 23 October 1970	Max wind - 118 kph	Tide plus surge at Chandpur: 4.7 m; Total water level was about 5.5 m.	300 deaths in Bangladesh
52	8 to 13 November 1970	Landfall at Hatia, Bangladesh. Central pressure - 940 mb; pressure deficit - 70 mb; max. wind - 222 kph.	Tide plus surge 5.6 m at Chittagong and Comapaniganj; total water levels (including wind waves) reached 10.7 m.	Between 200,000 and 300,000 deaths in Bangladesh.
53	7 to 8 May 1971	Landfall in Bangladesh	Total water level (including tide surges and wind waves) about 5 m.	
54	28 to 30 September 1971	Landfall in Bangladesh	Tide plus surge 5.0 m at Chandpur	
55	30 October 1971	Landfall at Paradeep, Orissa State, India. Central pressure - 960 mb; pressure deficit - 40 mb; max. wind - 167 kph.	Area Affected: Orissa Coast of India; peak surge: 4.0 m; tide: 0.9 m; peak water levels of 6.0 m.	10,000 deaths in India.
56	5 to 6 November 1971	Landfall near Chittagong, Bangladesh.	Peak surge at Chittagong: 2.1 m; tide: 0 m; total water levels up to 5.5 m (including wind waves).	
57	7 to 14 September 1972	Landfall near Baruva (between Kalingapatnam and Gopalpur), India. Central pressure - 945 mb; pressure deficit - 65 mb; max. wind - 204 kph.	Area Affected: coast of India between Baruva and Chandbali. Storm surges up to 3.4 m; tide: 0.8 m.	

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
58	15 to 23 November 1972	Landfall south of Nellore in Andhrapradesh State, India. Central pressure - 983 mb; max. wind - 148 kph.	Minor surge near Nellore in India.	
59	16 to 18 November 1973	Landfall near Barisal, Bangladesh. Central pressure - 1000 mb.	Surge: 1.0 m; peak water levels (includes wind waves) 3.8 m.	
60	6 to 9 December 1973	Landfall at Sundarban, Bangladesh. Central pressure - 1002 mb; max. wind - 118 kph.	Peak water levels up to 6.2 m; peak surges up to 4.5 m.	183 deaths in Bangladesh
61	13 to 15 August 1974	Landfall in Bangladesh. Max. wind - 100 kph.	Peak water levels up to 6.5 m; water levels varied from 1.5 to 6.5 m along the Bangladesh coast.	
62	29 November 1974	Landfall at Chittagong, Bangladesh. Central pressure - 974 mb; pressure deficit - 36 mb; max. wind - 161 kph.	Surge at Chittagong: 3.1 m; tide: 0.2 m; peak surges in other areas of Bangladesh: 4.8 m; tide: 1.17 m; total water levels in excess of 6.0 m.	20 deaths in Bangladesh
63	5 to 7 May 1974	Landfall at GWA (between Sandoway and Bassein), Burma. Pressure fell 25 mb in 9 hours at Bassein. Central pressure - 965 mb; radius of maximum winds 48 to 64 km; max. wind - 161 kph.	Surge propagated 100 km inland into estuaries and rivers. Peak water level (surge plus tide) occurred at 48 km inland in a river. Near the shore, the duration of the surge was 12 hours. Some 48 km inland, the duration was 72 hours. Further inland, the duration decreased to 6 hours.	303 deaths in Burma. Damage of 776.5 million Kyat.

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
64	5 to 7 June 1975	Landfall near Chittagong, Bangladesh.	Peak surge of 4.0 m in Bangladesh	
65	24 to 28 June 1975	Landfall in Bangladesh. Max. wind - 161 kph.	Maximum value of surge plus tide in Bangladesh was 4.8 m.	
66	8 to 12 November 1975	Landfall between Barisal and Noakhali in Bangladesh. Central pressure - 979 mb; pressure deficit - 31 mb; max. wind - 143 kph.	At Chittagong, surge of 2.2 m; tide: 0 m; maximum surge on Bangladesh coast: 3.1 m.	
67	29 April to 2 May 1976	Landfall at Sandoway, Burma. Max. wind speed - 129 kph.	Storm surge on Burma coast.	
68	11 September 1976	Landfall at Contai, India. Central pressure - 971 mb; pressure deficit - 38 mb; max. wind - 148 kph.	Surge on India coast: 2.5 m; tide: 1.4 m.	40 deaths in India.
69	18 to 20 October 1976	Landfall in Bangladesh. Max. wind - 105 kph.	tide plus surge 5.0 m at Comapaniganj in Bangladesh.	
70	20 November 1976	Landfall at Chittagong, Bangladesh. Central pressure - 990 mb; pressure deficit - 20 mb; max. wind - 111 kph.	Surge on Bangladesh coast: 1.0 m; tide: 2.1 m.	
71	12 to 13 May 1977	Landfall at Sundarban, Bangladesh. Max. wind - 121 kph.	Surge on Bangladesh coast: 0.6 m; tide: 0.7 m.	

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
72	19 November 1977	Landfall between Nizampatnam and Chirala on the Andhrapredesh coast of India. Central pressure - 909 mb; pressure deficit - 101 mb; max. wind - 259 kph.	Peak surge: 5.0 m; tide: 0.3 m; Divi and surroundings on the Andhrapradesh coast of India affected by surge. Peak water levels exceeding 6 m occurred on a coastal stretch 80 km long and penetrated 8 to 15 km inland on low-lying deltas.	Between 10,000 and 20,000 deaths in India.
73	12 to 17 May 1978	Landfall at Myebon (between Akyab and Kyaukpyu). Pressure fell by 21 mb in 6 hours at Kyaukpyu. Central pressure - 974 mb at Kyaukpyu. Max. wind - 161 kph.		90 - 95% of the city of Kyaukpyu damaged.
74	November 1978	Landfall at Pamban, Tamilnadu coast, India. Central pressure - 938 mb; max. wind - 212 kph.	Peak surges of 4 m in Sri Lanka and on the Tamilnadu coast of India.	373 deaths in Sri Lanka. Extensive damage on the north-east coast of Sri Lanka. 10 deaths in India.
75	12 May 1979	Landfall at Kavali (between Nellore and Ongole), Andhrapradesh coast of India. Central pressure - 936 mb; pressure deficit - 56 mb; max. wind - 189 kph.	Peak surge of 3 m; tide: 0.6 m on the Andrapradesh coast of India.	700 deaths in India.
76	10 December 1981	Landfall in Bangladesh. Max. wind - 80 kph.	Surge of 2 m on Bangladesh coast.	15 deaths in Bangladesh.

Table 2. (continued)

Serial Number	Dates of Storm	Meteorological Aspects	Storm Surge Aspects	Casualties and Damage
77	2 to 5 May 1982	Landfall south of GWA, Burma Central pressure - 950 mb; pressure deficit - 58 mb; wind speed - 226 kph; speed of storm movement - 24 kph; radius of maximum winds - 40 to 48 km.	Peak surge of 4 m along the south Arakan coast of Burma.	31 deaths in Burma. Damage of 38 million Kyat.
78	1 to 4 June 1982	Landfall between Paradip and Chandbali on the coast of India.	2 m surges on the Orissa and West Bengal coasts.	245 deaths in India.
79	15 October 1983	Landfall near Chittagong Severe cyclonic storm		43 deaths in Bangladesh. Substantial damage.
80	9 November 1983	Landfall betwen Chittagong and Cox's Bazar. Max. wind - 136 kph.	Storm surge over 1.7 m along 12 km stretch of coast.	300 fishermen missing in Bangladesh. Substantial damage.

Table 3. Nomenclature used by India Meteorological Department

Nomenclature	Wind Speed in Knots
Low	17
Depression	17 to 33
Cyclone	34 to 63
Severe Cyclone*	64 to 89
Very Severe Cyclone*	90 to 119
Catastrophic Cyclone*	120 or above

* The word cyclone is sometimes replaced with storm

Table 4. Maximum possible storm surge amplitudes and total water levels (storm surge plus wind waves) at selected locations on the east coast of India (for comparison a few stations outside India are included). The hypothetical storm has a wind speed of 40 m sec^{-1}. The classification in the last column is based on the total water level. A: 0 to 2 m, B: 2 to 5 m, C: >5 m.

Location	Favorable Wind Direction	Storm Surge (m)	Storm Surge + Wind Wave (total level) in meters	Classification
Dhanushkodi	NNE	4.8	8.2	C
Rameswaram	SE	6.8	11.3	C
Pamban	NNW	4.4	7.3	C
Devipatnam	E	4.5	7.5	C
Adirampatnam	SSE	5.1	8.5	C
Point Calimere	SSE	4.2	7.0	C
Nagapattinam	E	1.5	2.5	B
Karikal	E	0.3	1.3	A
Madras	ENE	1.5	2.5	B
Nizampatnam	SW	4.5	7.4	C
Mouth of Krishna River	SE	1.6	2.7	B
Narasapur	S	1.7	2.9	B
Sacramento Shoals (outer sand banks)	SSE	1.4	2.3	B
Kakinada (outer sand banks)	E	0.6	1.0	A
Visakhapatnam	SE	0.7	1.2	A
Kalingapatnam	E	1.1	1.8	A
Gopalpur	SE	0.9	1.5	A
Mouth of Devi River	SE	0.8	1.3	A
False Point	SE	1.9	3.2	B
Balasore	SE	3.0	5.0	C
Mouth of Hoogly River	S	6.5	10.8	C
Mouth of Matla River	S	5.0	8.4	C
Mouth of Baleswar River (Bangladesh)	S	6.9	11.5	C
Mouth of Meghna River (Lakhichar Island, Bangladesh)	SSE	8.0	13.4	C
Cox Bazar (Bangladesh)	WSW	3.2	6.3	C
Mouth of Faaf River (Burma)	SW	3.2	5.3	C

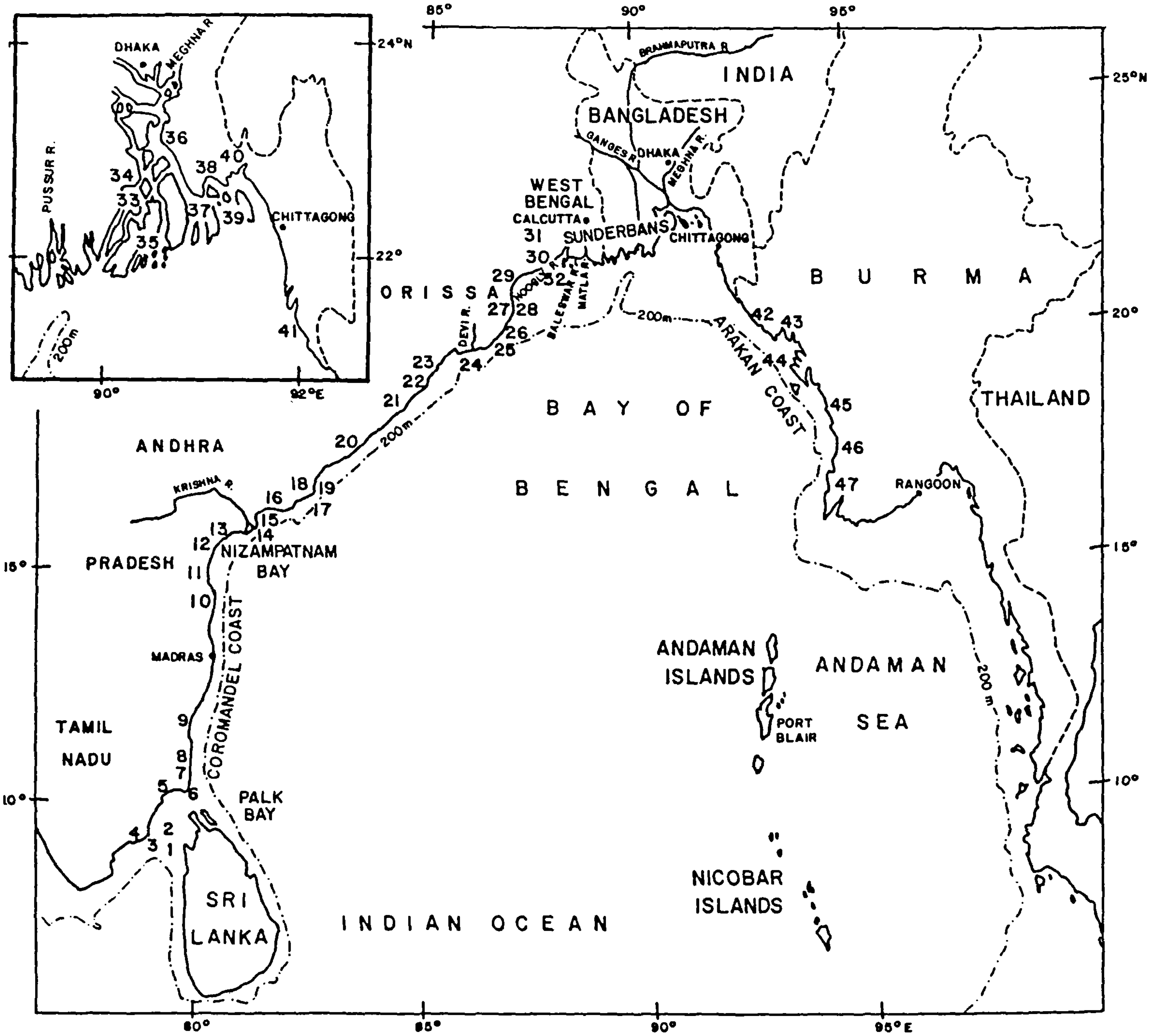

Figure 1. Map of the Bay of Bengal Region. See Table 2 for the names and locations of places identified by numbers.

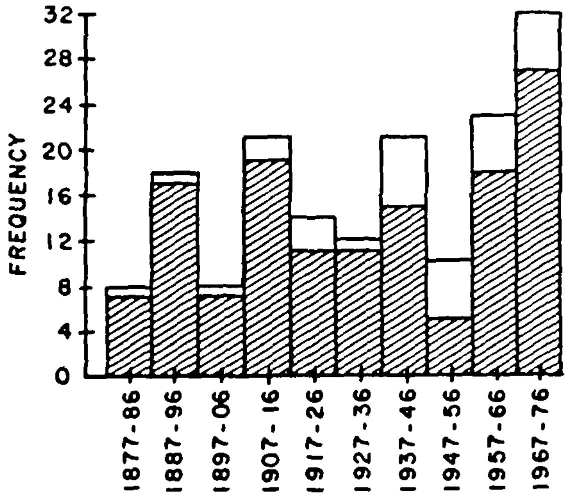

Figure 2. Frequency of severe cyclonic storms which formed over the Bay of Bengal (full rectangles) and those that struck the coast (hatched rectangles) in 10-year periods: 1877 to 1976 (from MOOLEY, 1980).

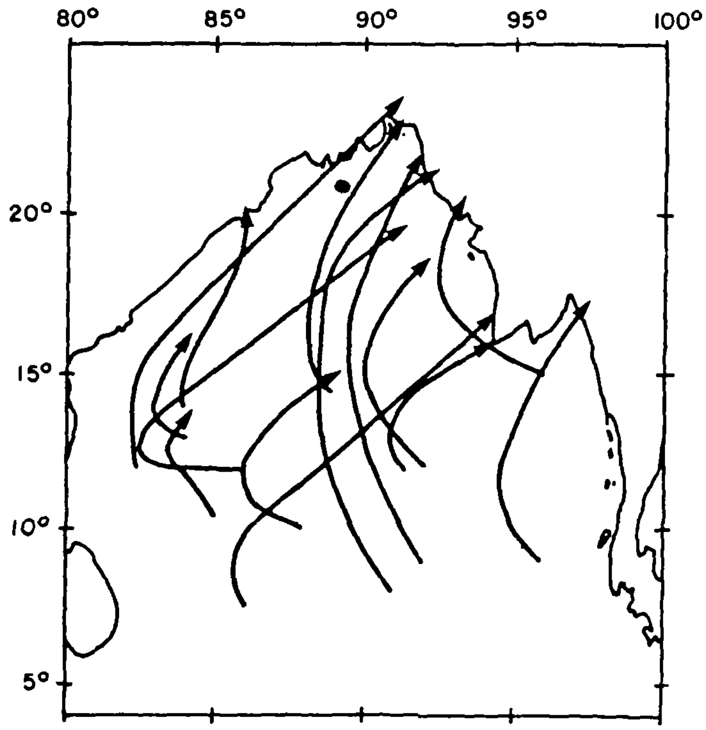

Figure 3. Tracks of some recurring storms in the Bay of Bengal during the period 1945 to 1954 (from CHAKRAVORTTY and BASU, 1956).

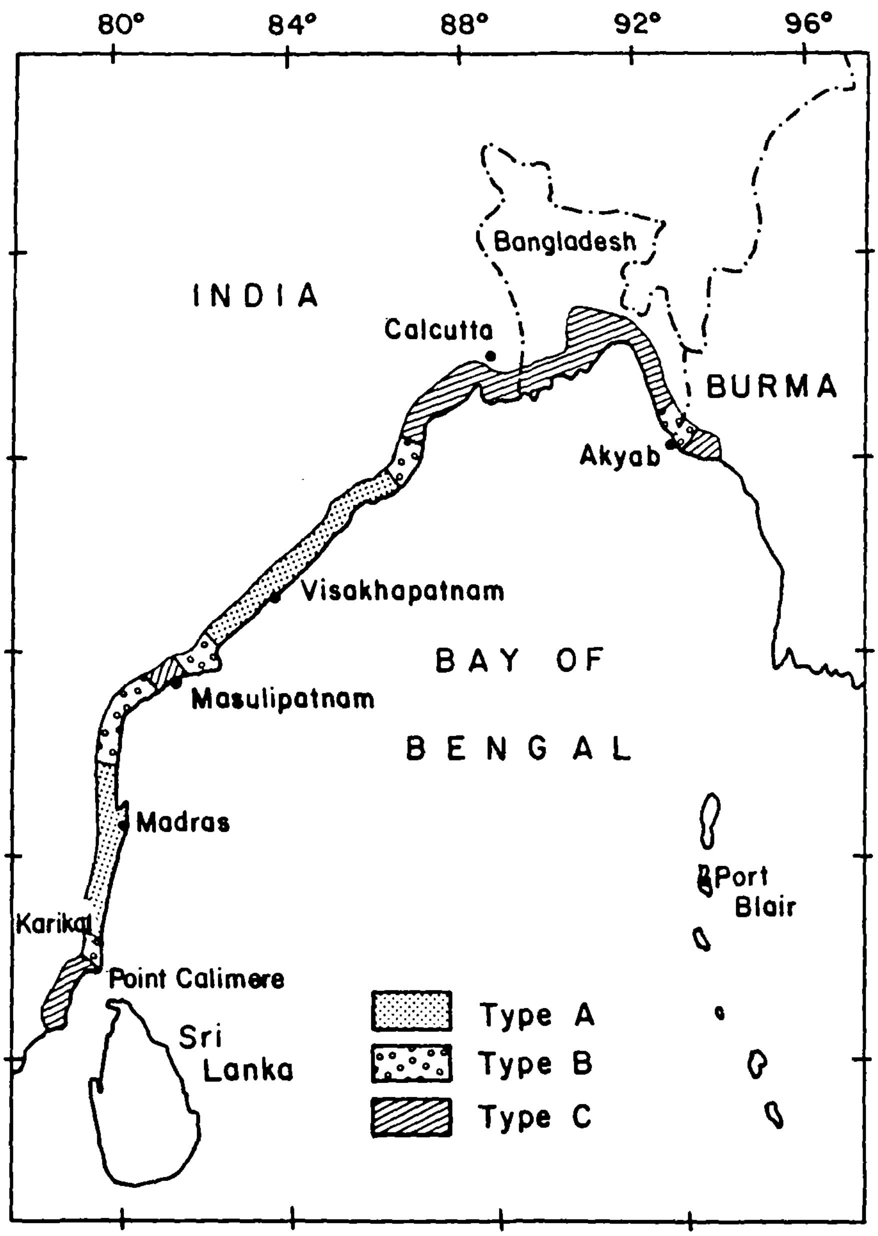

Figure 4. Classification of part of the Bay of Bengal coast according to the maximum possible water level (storm surge plus wind waves) resulting from a hypothetical storm with wind speed of 40 m sec^{-1} (from RAO, 1968).

CIRCULATION AND PHYSICAL PROCESSES OF THE ARABIAN SEA AND THEIR RELATIONSHIP WITH S.W. MONSOON AND PRODUCTIVITY

G.S. QURAISHEE

National Institute of Oceanography
P.E.C.H.S.
Karachi 29, Pakistan

INTRODUCTION

Circulation of the Arabian Sea as well as physical processes and related phenomena have direct impact on the economic development of the countries in the region. Seasonal changes in the Arabian Sea and reversal of monsoons influence various oceanographic and atmospheric processes. The southwest monsoon varies from year to year and causes significant influence on the economy of Pakistan and adjoining countries. The timing of the monsoon and its severity have direct bearing on the agricultural economy, irrigation, flooding of rivers and even on breeding of marine organisms in the coastal region. The observations so far obtained and inferences derived from them suggest some possibility of predicting SW monsoon and other oceanographic parameters.

MONSOON AND DYNAMIC HEIGHTS

The atmospheric forcing of the ocean during SW monsoon has been noted from the wind stresses during summer months (May - September), when positive stresses with strong gradients extend to 70°E, covering almost the entire North Arabian Sea (BRUCE, 1981), from the Somali coast to the coasts of Pakistan and India. In view of the pronounced seasonal cycle in the atmospheric and monsoon oceanic circulation in the Indian Ocean, beginning in April along the Somali coast, there is a possibility of forecasting the SW monsoon intensity if the events are well recorded from the African coast to the sub-continent. Such measurements would include dynamic heights, intensity of upwelling, records of temperature, thickness of the mixed layer, etc.

In April, the South Equatorial Current (SEC) along the African coast converges with the south- flowing Somali Current to form a loop-like feature which moves northward (MOLINARI, 1983). This early indication of the intensity of dynamic heights across the equator appears to have some relationship with the progressive development of monsoon intensity. One measure of intensity of the SW monsoon is reflected in Indus River discharge, where peak flow develops in August due to heavy rain and from large snowmelt due to persisting warm temperatures. Indus River discharge as a parameter for monsoon intensity has been used to correlate dynamic heights at the equator near Somali coast (Fig. 1). The variation of surface dynamic heights from 1°N to 1°S off Somali coast have some correlation with the intensity of monsoon in Pakistan (QURAISHEE, 1985). Steady and large variations in the dynamic heights are associated with high river discharges. These two events, i.e. dynamic heights and Indus discharges, have a lag of about 2 months which can be used to predict monsoon intensity two months hence.

MONSOON AND SOUTHERN OSCILLATION

Other physical processes, extended over larger oceanic areas, generally known as Southern Oscillations (SO) (PHILANDER, 1983), also appear to have some correlation with the SW monsoon variability (PANT and PARTHASARATHY, 1981). SO, in fact, represent large scale exchanges of atmospheric mass between the eastern and western hemispheres, visualised as a see-saw of high pressure in the South Pacific and low pressure in the Indian Ocean. These exchanges exhibit a cycle of 2 to 6 years. El-Niño has been found to be related with the SO. A long data base of 112 years (1866-1977) of monsoon intensity shows that wet monsoons (strong) and dry monsoons (weak) have correlation with El-Niño. The investigation on physical causes and relationship between SO and monsoon intensity can provide a good forecasting tool.

MONSOON AND THERMAL STRUCTURE

Monsoon intensity and thermal structure of the Arabian Sea also show reasonable correlation (SHUKLA, 1983). SST and temperatures at depth are affected by upwelling along the Somali and Arabian coasts and circulation of upwelled water (SWALLOW, 1985) outflow and mixing of Red Sea water at about 750 m depth and Persian Gulf water at about 250 m depth (MEERA PATHMARAJAH, 1982).

Cooling of the eastern Arabian Sea (SWALLOW, 1983) may be due to evaporation and mixing with deeper cool water, resulting in deepening of mixed layer. On the basis of the few available data, the depth of 20°C isotherm in the centre of North Arabian Sea in May appears to have some correlation with the August discharges of Indus River (QURAISHEE, 1985). The depth of 20°C isotherm (Table 1), which is considered to be almost in the middle of permanent thermocline, sinks to greater depths when large mixed layers are formed. The high floods of the River Indus in 1967, 1973, 1976, 1978 and 1983 are reasonably correlated to deeper depths of 20°C isotherm, i.e. 180 - 200 m. Thus there is a reasonable correlation between temperature structure, particularly the depth of thermocline, and the intensity of monsoon rainfall.

MONSOON AND EDDY CIRCULATION

Satellite imageries, using high resolution infrared radiometer, show that circulation of the Arabian Sea contains warm and cold core eddies (CAGLE and WHRITNER, 1981). The eddy circulation appears to become intensified and locally persist (QURAISHEE, 1984) during the SW monsoon (May - September). These months are dominated by upwelling along the Arabian coast, and the cold water plume and wedge extend eastward. Comparatively weak upwelling also appears along the Pakistan coast west of Karachi. In the middle of these two upwelled cold water areas an anticyclonic eddy is found with a warm core. Records generally show that this eddy circulation is repeated in the SW monsoon.

The warm core eddies appear to transfer energy from the ocean to the atmosphere. Consequently the rainfall on the sub-continent is affected. The thermal field in the warm core eddy of June 1983 has been investigated (QURAISHEE, 1984). The normal surface temperature in June is 29°C, based on WYRTKI's (1971) bimonthly charts, when the mixed layer depth is about 20-40 m. Eddy circulation, however, disturbs these normal conditions: cool water (26.0°C) prevails near the coast, and warm water (29.0°C) is trapped in the centre of the eddy. The trapping of warm water is reflected by the configuration of isotherms at 50 m and 100 m, and less prominently at 200 m and 500 m (Fig. 2). The 20°C isotherm lowers to a depth of about 230 m (Fig. 3) which is as much as 50 m deeper than normal (see above). Thus information about the thermal structure in an eddy field also can provide a useful parameter for prediction of monsoon intensity.

EDDY CIRCULATION AND PRODUCTIVITY

In 1983 high values of phosphate and nitrate were observed at the core of a warm-core eddy. Mixing of the nutrients from the adjoining cool upwelled water, particularly from deeper depths (50 - 300 m), is apparent. High phosphate concentrations (2.23 µg-at P $^{-1}$) have been observed at the surface.

In the Arabian Sea dissolved oxygen concentrations generally decline rapidly with depth, particularly north of 20°N; the dissolved oxygen content of the water in thermocline decreases at temperatures of 26°-27°C, signifying that the oxygen demands exceed the supply at 100 m. In the eddy field oxygen values of 3 ml l^{-1} at 100 m indicate that water trapped in the eddy field is well oxygenated. Even at the thermocline, where water temperatures are 26-27°C, oxygen concentrations are 3.5 ml l^{-1}, possibly because of the sinking of surface water as indicated by the salinity distribution (Fig. 4).

The whole eddy field is rich in both nutrients and dissolved oxygen, which would infer an increase in local biological productivity. Preliminary observations indicate that the zooplankton is indeed rich, and standing crops (measured by displacement volume) of 0.29 - 0.72 m m^{-3} have been observed. However, more detailed studies are required to correlate the intensity of eddy with productivity, and in turn to SW monsoon intensity.

CONCLUSION

It is possible to predict monsoon intensity in advance, from few weeks to months, and possibly even several years based on various physical oceanographic parameters and processes resulting from ocean and atmospheric interaction. However, detailed and systematic studies are required to understand scientific phenomena and refine the forecast.

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Table 1. **Distribution of the depth of 20°C isotherm in the Central North Arabian Sea and monsoon discharge in August in River Indus.**

Year	Depth of 20°C Isotherm (May) (Meters)	Location	Indus discharge in August (km^3)
1962	-	-	17.3
1963	-	-	18.9
1964	-	-	25.3
1965	-	-	17.3
1966	-	-	30.3
1967	210	22°N, 61°E	36.2
1968	-	-	27.7
1969	-	-	23.4
1970	-	-	9.2
1971	-	-	12.4
1972	-	-	7.6
1973	170	20°N, 65°E	39.1
1974	-	-	3.3
1975	-	-	21.4
1976	200	8°N, 10°N 50°E, 60°E	38.1
1977	150	61°N, 18°N 50°E, 60°E	18.2
1978	200	8°N, 9°N 50°E, 60°E	46.4
1979	200	19°N, 20°N 50°E, 60°E	16.4
1980	-	-	11.4
1981	-	-	17.9
1982	-	-	6.9
1983	180	21°N, 22°N 62°E	26.6

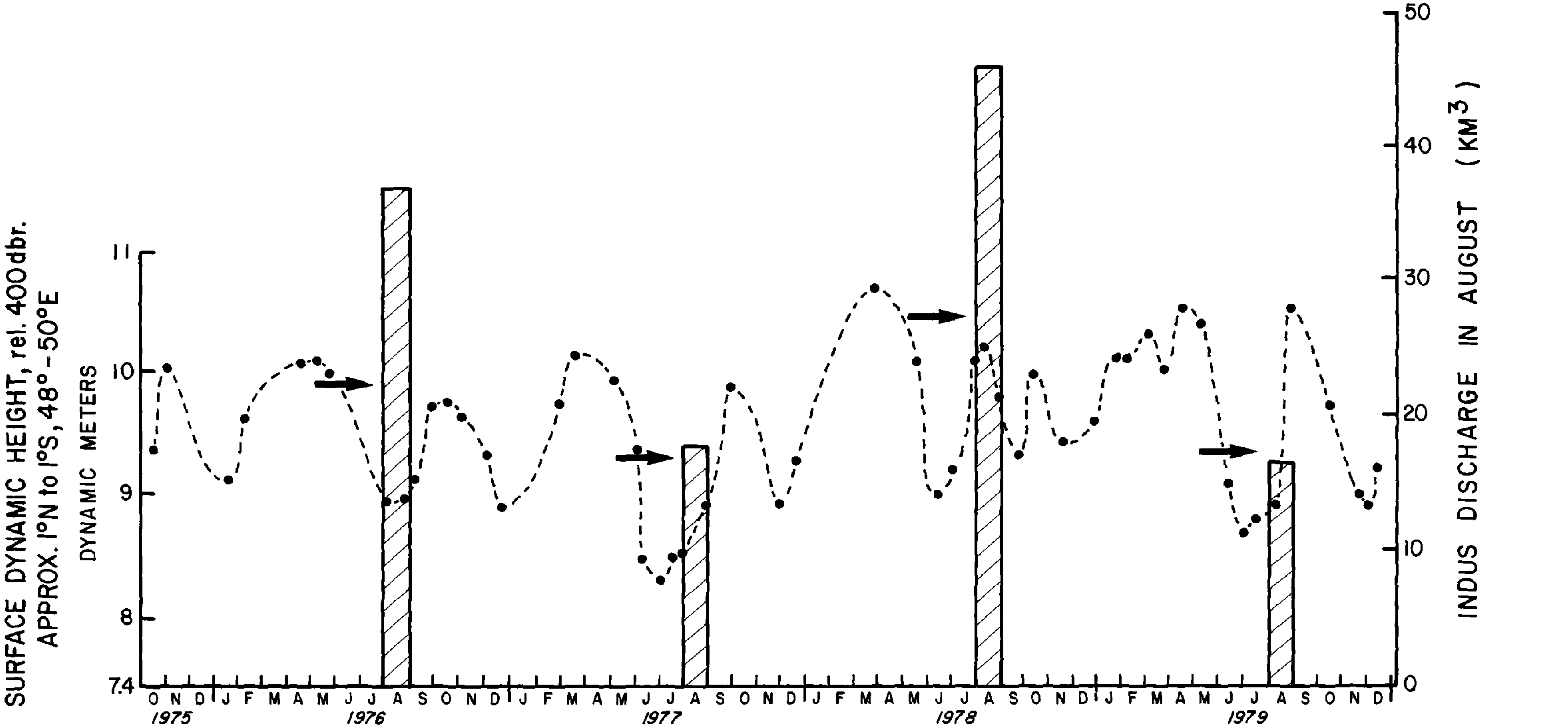

Figure 1. Variation in surface dynamic heights off Somali coast (BRUCE, 1981) and monsoon flooding in Indus River (Hatched). Arrows indicate time lag between peaks of dynamic heights and flooding.

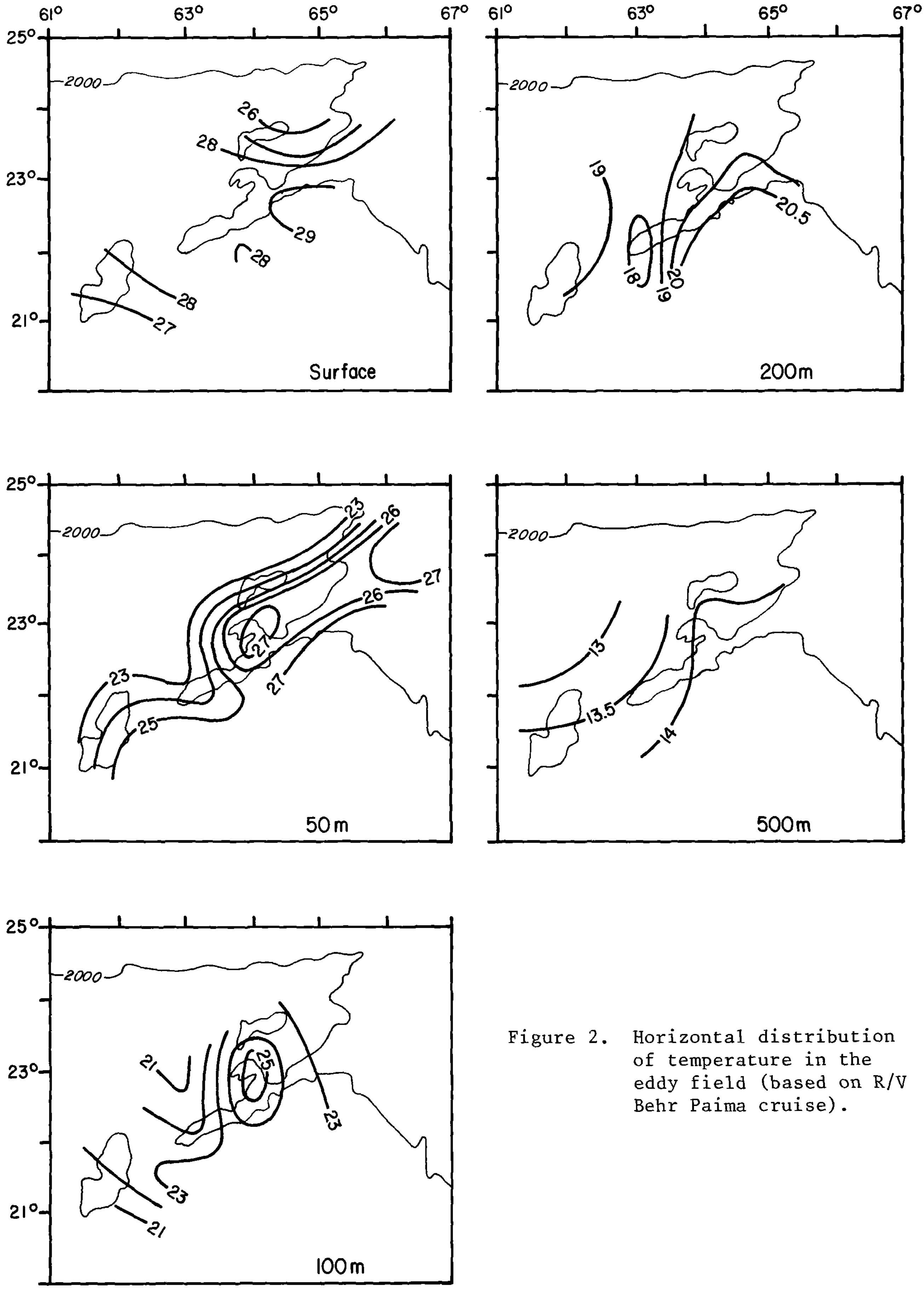

Figure 2. Horizontal distribution of temperature in the eddy field (based on R/V Behr Paima cruise).

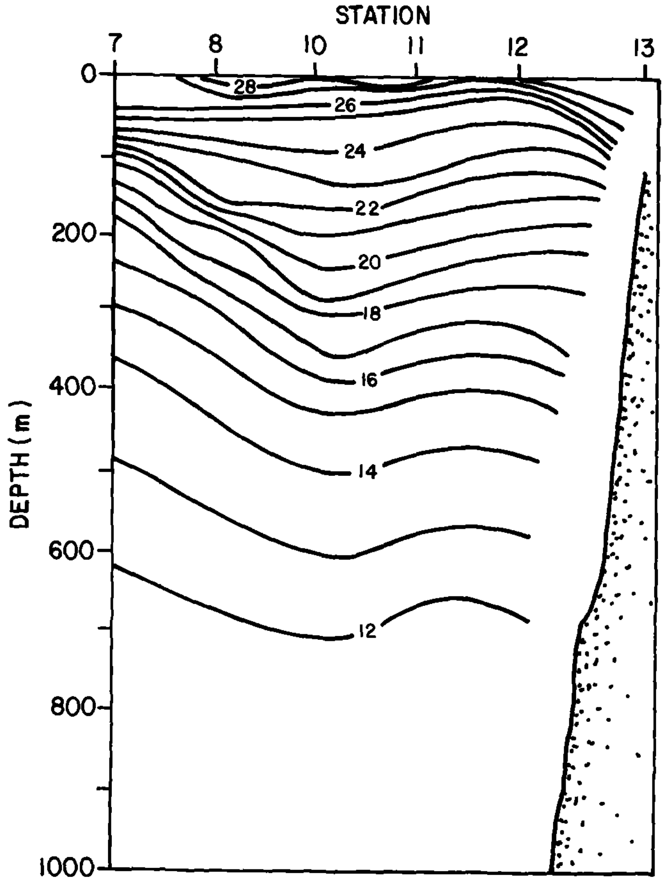

Figure 3. Vertical distribution of temperature in the eddy field (based on R/V Behr Paima cruise).

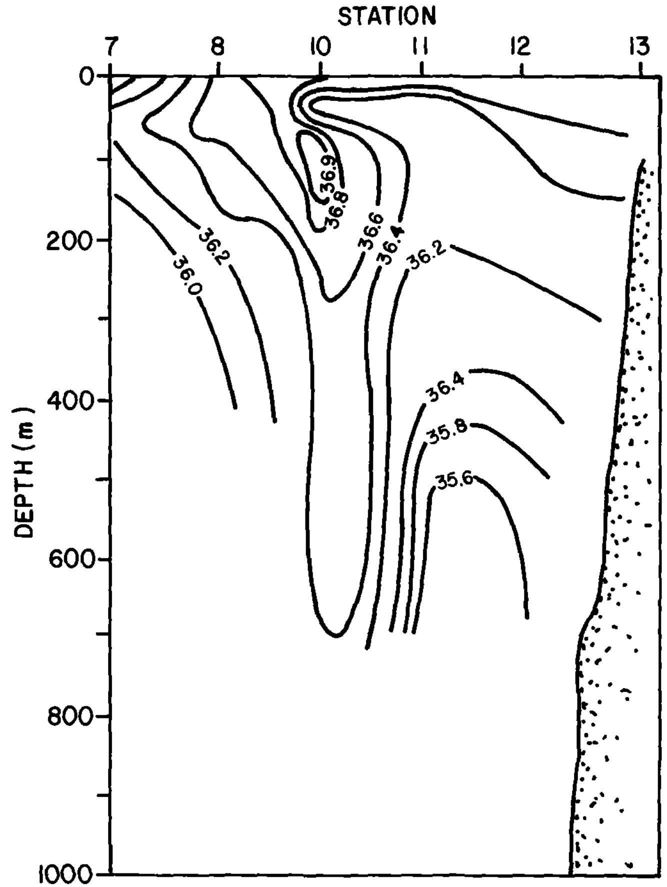

Figure 4. Sinking of surface water as indicated by vertical salinity distribution (based on R/V Behr Paima cruise).

A REVIEW OF THE PHYSICAL OCEANOGRAPHIC CHARACTERISTICS IN THE ARABIAN GULF

T.S. MURTY AND M.I. EL-SABH

Department of Fisheries and Oceans
Institute of Ocean Sciences
P.O. Box 6000
Sidney, B.C., Canada V8L 4B2

Département d'Océanographie
Université du Québec à Rimouski
300 avenue des Ursulines
Rimouski, Québec, Canada G5L 3A1

ABSTRACT

The meteorological and physical oceanographic systems that affect the Arabian Gulf (the "Inner Gulf" of the Kuwait Action Plan Region) are considered. The so-called "Shamal" phenomenon can cause significant positive and negative storm surges in the Gulf. Water balance is maintained by flow through the Strait of Hormuz, with surface inflow and deep outflow, resulting in a flushing time of a few years. Tidal height patterns and residual currents also are discussed.

INTRODUCTION

The Arabian Gulf (also termed the Persian Gulf and the "Inner Gulf" of the Kuwait Action Plan Region) is a semi-enclosed sea measuring 1000 km in length; it varies in width from a maximum of 340 km to 60 km in the Strait of Hormuz (Fig. 1). It has an area of 226,000 km^2 with an average depth of about 35 m. The Gulf is asymmetrical in profile with a more gentle gradient on the southwestern side. The northeastern coastline is marked by mountains and cliffs whereas the southwestern shore is often sandy.

The continental area surrounding the Gulf is very arid, with considerable evaporation. The result is a classical Mediterranean-type circulation of surface inflow and deep outflow, with salinity being the most important determinant of density. As will be seen later, water circulation is governed by three factors: excess of evaporation, tidal currents and wind-driven currents.

The physical oceanography of the Gulf has been reviewed by GRASSHOFF (1976), HUGHES and HUNTER (1979) and HUNTER (1982). A recent bibliography of the marine science of the region has been given by FARMER and DOCKSEY (1983). The purpose of this paper is to provide a "state of the art" review of the known physical characteristics of the Gulf, following the recent symposium/workshop held in Saudi Arabia in October 1983 (EL-SABH, 1984).

METEOROLOGICAL SYSTEMS

The Arabian Gulf is affected mainly by extra-tropical weather systems, whereas the Gulf of Oman is at the northern edge of tropical weather systems (MURTY and EL-SABH, 1984). Thus the Strait of Hormuz region forms the boundary between the generally west-to-east travelling extra-tropical weather systems and east-to-west travelling tropical weather systems. One of the interesting weather phenomenon in the region is the so-called "Shamal" (Arabic word for North) which is a sub-synoptic scale wind phenomenon that occurs with sufficient frequency and influence on the local weather (PERRONE, 1981). "Shamal" is used in the meteorological context to refer to seasonal northwesterly winds that occur during winter as well as summer.

The winter Shamal occurs mainly from November to March and is associated with mid-latitude disturbances travelling from west to east. It usually occurs following the passage of a cold front and is characterized by strong northwesterly winds (mainly during December to February). PERONNE (1981) mentions that the winds at most locations in the Gulf exceed 20 knots less than 5% of the winter. Although infrequent, the winter Shamal is not insignificant from an operational point of view, because it sets in with great abruptness and force; the March 1983 oil spill in the Arabian Gulf occurred during a winter Shamal. The summer Shamal occurs with practically no interruption from early June through July and is associated with the relative strengths of the Indian and Arabian thermal lows. When viewed from the point of wind strength and the associated weather conditions, however, the summer Shamal is not as important as the winter Shamal.

Although the winter Shamal is a relatively rare event, typically occurring only once or twice each winter, it brings some of the strongest winds and highest seas of the season to the Gulf region. Usually the Shamal occurs first in the northwest part of the Gulf and then spreads south and east behind the advancing cold front. It takes up to 12 to 24 hours for the Shamal to spread from the northwest corner of the Gulf to the southern part. The onset of the Shamal is difficult to predict, mainly because of the difficulty in forecasting the associated upper air pattern. Once the Shamal has begun, it may subside within 12 to 36 hours after the passage of the cold front or it may persist for three to five days. The relationship between the surface and upper air patterns determines which sequence is most likely to occur. Once the Shamal begins, the wind direction is strongly influenced by the coastal orography. In the northern part of the inner Gulf, the Shamal winds generally blow from north to west-northwest. In the middle parts of the Gulf, Shamal winds tend to be from west-northwest to northwest. On the southeast coast of the Gulf, the winds are westerly. In the Strait of Hormuz area, the Shamal winds are generally from the southwest. The speed of Shamal winds generally ranges from 20 to 40 knots.

The Shamal winds persist for three to five days with gale force and blow from northwest over the whole Gulf. The longer duration Shamals occur when the upper air trough stalls over the Strait of Hormuz. Because of the large pressure gradient between the low (over the Gulf of Oman) and the high (over Saudi Arabia), the Shamal winds are strongest in the southern and southeastern parts of the Gulf. Average wind speeds in the southern and southeastern parts of the Gulf range from 30 to 40 knots, with peak winds in excess of 50 knots not uncommon. Winds over the northern part of the Gulf, on average tend to be 5 to 15 knots less than the above values.

According to PERRONE (1981), two areas of the Gulf experience stronger than average Shamal conditions (Fig. 2). One area is near the Qatar Peninsula, and another is near Lavan Island.

STORM SURGES

Based on the linearized versions of the shallow water equations, MURTY and EL-SABH (1984) numerically simulated the storm surges in the Gulf using finite difference scheme. Figure 3 shows schematically the spatial structure of the storm as it advances one grid space in the x-direction. This storm has a life of about one day and is supposed to represent a short duration winter Shamal. Figure 4a shows for selected locations the computed profiles of tide (with three semi-diurnal and three diurnal constituents included), tide plus storm surge and storm surge alone.

MURTY and EL-SABH (1984) also made additional simulations with the same wind directions, but for a longer duration (4 to 5 days) Shamal. Figure 4b shows the results of the simulations when the wind constantly blows from the positive y-direction, while Figure 5 shows plots at four different times of the horizontal distribution of the storm surge heights. From these various types of plots, it can be seen that significant positive and negative storm surges can occur in the Arabian Gulf. The tide is also important.

PHYSICAL OCEANOGRAPHIC PROPERTIES

The temperature-salinity diagram for the Gulf winter waters (Fig. 6) clearly shows the T-S maximum associated with the outflowing water in the Gulf of Oman. Surface water of approximately 22°C and 36.5‰ enters the inner Gulf and flows northward, cooling to 17-18°C and being greatly modified in salinity to 40.5‰ as it flows out. The cluster of points of high salinity on the right of Figure 6 represents deeper outflowing water and also the cool very high salinity water found on the western and southern shelves. According to BREWER and DYRSSEN (1985), this line of points across the bottom of the diagram represents the intense (but low capacity) signal of low salinity water originating from the confluence of the Tigris, Euparates and Karun rivers at the delta of the Shatt Al-Arab.

The surface salinity distribution (Fig. 7a) shows Gulf of Oman water entering and tending to flow northwards along the Iranian Coast. The vertical salinity distribution along the major axis (Fig. 7b) shows the Mediterranean-type circulation clearly. The highest deep salinity along this section at Station 2368 is of particular interest. BREWER and DYRSSEN (1985) show that these high salinity waters form in the shallow area near United Arab Emirates' coast and migrate down slope to augment the outflowing waters.

The Gulf undergoes wide and rapid temperature changes in response to daily and seasonal cycles of heating and cooling. As with salinities, the highest water temperatures are found in the shallow bays and lagoons where annual temperatures can range from 15°C in winter to 40°C in summer. Even offshore, temperatures can range from 18°C in winter to 36°C in summer.

Mixing processes in the Gulf were reviewed by HUGHES and HUNTER (1979) who showed, from simple analyses of the limited oceanographic data available, that: a) The time to obtain 90% mixing of a contaminant over a water column in the Gulf would be around 16 days (longer in summer, due to the presence of vertical thermal density stratification), and b) Exchange of water in the horizontal direction is dominated by the residual circulation. Two time scales were estimated for the Gulf by HUGHES and HUNTER (1979) in relation to the overall mixing processes. The time for 90% flushing of the basin was estimated to be of 5.5 years. They also estimated the time it takes for all the water in the Gulf to come within the influence of the open sea to be days. However, if the interaction of vertical mixing is included with their estimated residual current, the turnover time increases to 2.4 years which corresponds to the value estimated by HARTMANN et al. (1971).

TIDES AND TIDAL CURRENTS

The tides in inner Gulf co-oscillate with tides in the narrow Strait of Hormuz, while tides in the Gulf of Oman co-oscillate with those in the Arabian Sea. Usually the Gulf and the Strait of Hormuz are treated as a coupled system for tidal computations. The range of tides in the Gulf is greater than 1 m (LEHR, 1984), and exceeds 3 m at Shatt-al-Arab. These large amplitudes cause strong tidal currents which commonly exceed 0.5 m s^{-1} at maximum ebb or flood. The tide appears to progress up along the Iranian coast from the Strait of Hormuz and down the coast of Saudi Arabia. The dimensions of the Gulf are such that resonance amplification of the tides can occur (HUGHES and HUNTER, 1979).

M_2, S_2 and K_1 tidal constituents (Fig. 8), together with O_1 tides, are the four important tidal constituents in the Gulf. As can be seen, the semi-diurnal constituents have two amphidromic points, one in the northwestern part of the Gulf and the other in the southwestern part. The K_2 and O_1 constituents have a single amphidromic point almost in the middle of the Gulf. In fact, the tides in the Gulf are complex, consisting of a variety of tidal types (Fig. 9), with areas semi-diurnal [$S = (M_2 + S_2)/(K_1 = O_1)<0.5$], predominantly semi-diurnal (S between 0.5 and 1.0), predominantly diurnal (S between 1.0 and 1.5) or diurnal (S>1.5). The diurnal character of the tide appears around the amphidromic points of the semi-diurnal cotidal maps. EVANS-ROBERTS (1979) showed that the diurnal characteristics for tidal currents arise in the central part of the Gulf, over an area where tidal elevations are semi-diurnal, and vice-versa for the semi-diurnal. This is evidently due to the amphidromic shape of these tides, where maximum amplitudes of currents are located in the areas of minimum amplitude of sea surface elevation.

Several two-dimensional models of tides in the Gulf already exist (for complete references, see EL-SABH, 1984; LE PROVOST, 1984). All of these models either have been used or are capable of predicting tidal currents as well as heights, but there is almost a complete absence of systematic current surveys to measure tidal current in detail to provide empirical verification. IMCOS Marine (HIBBERT, 1980) claims to have developed a verified model to predict currents from tide height predictions applicable to any part of the Gulf.

More recently, MURTY and EL-SABH (1985) studied the age of tides in the Gulf (the age of tides is a term used to denote the interval between the time of new or full moon and the time of the local spring tide). As can be seen in Figure 10, the age of the semi-diurnal tide in the Gulf varies between -35 to 76 hours, with the major variations occurring in areas where the semi-diurnal tide has an amphidromic point. Higher values are found in the shallow area between Bahrain and United Arab Emirates. On the other hand, all ages for the diurnal tide show negative values which vary between -4 and -92 hours; the lower value is found in the southern part of the Strait of Hormuz while higher values occur in the area where the amphidromic point for the diurnal tide occurs. In general, higher ages occur in shallow areas and near the amphidromic points, associated with a major topographic feature.

RESIDUAL CIRCULATION

Very little information exists describing the residual current field in the Gulf, the most recent effort are those by HUNTER (1984) and GALT et al. (1984). It has been known for a long time that evaporation exceeds precipitation in the Gulf, and so the more saline dense water should sink and pass out of the Strait of Hormuz, giving rise to a compensating surface flow of less dense water into the Gulf. The effect of the Earth's rotation would deflect these flows to the right, giving a surface flow west and northwest along the Iranian coast, and a deep flow to the southeast and east along the coasts of Saudi Arabia and the United Arab Emirates (the deep flow would be constrained further to these latter coastlines adjacent to the shallow sea areas of high evaporation). This circulation pattern undoubtedly is modified by forcing of wind and atmospheric pressure, but it generally has been supported by observations of ship drift. Evidence for this anti-clockwise residual circulation in the Gulf, driven predominantly by evaporation (giving rise to horizontal density and pressure gradients), has been described by several authors (see HUNTER, 1984, for details).

The northwestern part of the Gulf is undoubtedly influenced by the freshwater inflow of the Tigris, the Euphrates and the Karun (entering the Gulf via the Shatt Al-Arab). This inflow appears to be deflected to the right by Coriolis force to form a river plume of width approximately 20 km, flowing along the Iraqi coast into Kuwait waters (MATHEWS et al., 1979). Freshwater inflow has been estimated by GRASSHOFF (1976) to vary between 5 and 100 $km^3\ yr^{-1}$. A more realistic estimate made by HARTMANN et al. (1971) of 37 $km^3\ yr^{-1}$ represents only about 9% water loss from evaporation. Since precipitation is light in the region, consideration of the differential flow through the Strait of Hormuz is crucial to the water balance in the Gulf.

On the basis of a geostrophic force balance across the Gulf and a frictional balance along the Gulf, HUNTER (1982) indicated the circulation schematically by the diagram shown in Figure 11. Evaporation in the shallow areas of the southwestern and southern Gulf leads to a sinking of dense water that is deflected to the right by Coriolis force, to flow out at the bottom of the Strait of Hormuz. This flow is compensated by a surface inflow along the Iranian coast. This circulation will be modified by the effect of surface wind stress, especially in shallow areas where the effect of density forcing will be small.

GALT et al. (1984) numerically simulated the steady state surface circulation in the Gulf from the combined effects of wind forcing and excess of evaporation (Fig. 12). One major feature of the circulation is a current moving to the southeast along the western shore. This first develops in southern Kuwait and then moves south along Saudi Arabia, gaining strength as it moves. By the time it reaches Bahrain, some of it moves south along the western shore of Bahrain, but most of the flow is bathymetrically steered to the south and east towards the northern tip of Qatar where some small fraction of it rounds the Qatar coast and moves into the eastern basin of the Gulf. Some re-circulation north of Qatar also can be seen. The maximum speed associated with this current is about one-half knot along the southern coast of Saudi Arabia and north of Bahrain and Qatar. A second major feature of the derived current pattern is coastal flow along the eastern shore or the southwestern coast of Iran.

This current is weaker and narrower than the one along the Saudi Arabian coast. Throughout the center of the northwestern basin, there is a weak return flow that covers the deeper segments of the Gulf. This continues towards the northwest where it approaches the shoreline and bifurcates about midway down the Kuwait coast.

It is clear that the circulation of the Arabian Gulf is complex and three-dimensional. The only numerical three-dimensional model of the overall residual currents in the region is by HUNTER (1983). The model was used to simulate the steady-state response of the Gulf to a specified density distribution and a wind stress. The prediction (Fig. 13) indicates a surface inflow of strength around 10 cm s^{-1} along the Iranian coast, some evidence of river inflow at the northwestern end of the Gulf, and an outflow of water along the bottom from the coastal areas off Saudi Arabia, Qatar and the United Arab Emirates.

The inflow of water from the Gulf of Oman, near the Iranian Coast, and conditions in the Strait of Hormuz have been observed by SONU (1979). In the Strait of Hormuz, the inflow was estimated to occupy the top 30 m of the water column, a mixing layer in the middle 20 m and the outflow at the bottom 30 m. This deep saline outflow of water from the Strait of Hormuz into the Gulf of Oman has been observed by several authors (e.g. BREWER and DYRSSEN, 1985; EMERY, 1956; HARTMANN et al., 1971; LEVEAU and SZEKIELDA, 1967). It appears that a tongue of saline water flows out of the Strait of Hormuz at a depth of about 100 metres, sinks to about 200 metres in the Gulf of Oman and finally sinks to about 500 metres in the South Arabian Sea.

ACKNOWLEDGEMENTS

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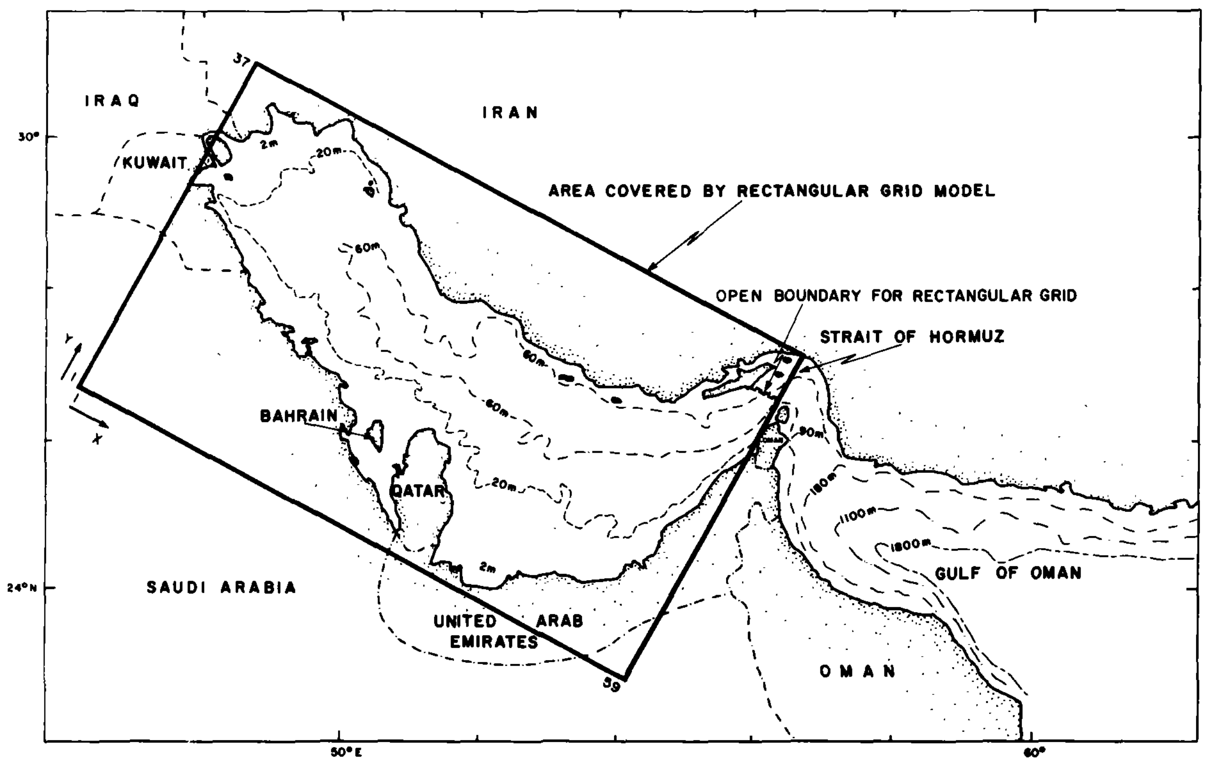

Figure 1. Geography and bathymetry of the Gulf. The area enclosed by the rectangle has been modeled numerically by the authors for tides and storm surges.

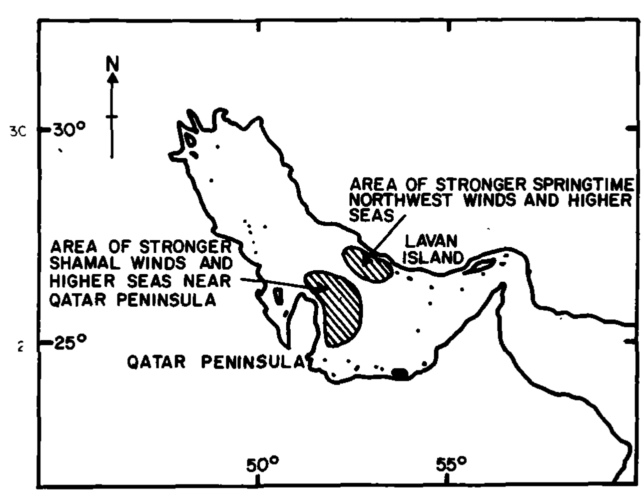

Figure 2. Areas of stronger than normal northwesterly winds and higher seas (from PERRONE, 1981).

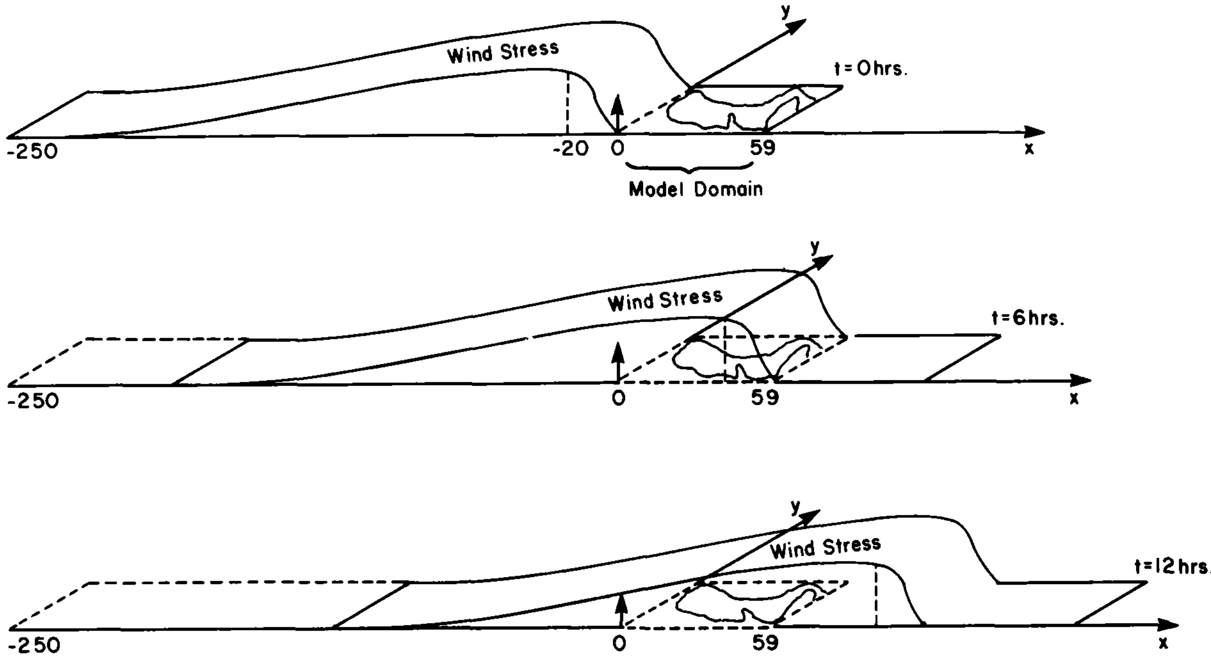

Figure 3. Schematic of the wind field used for storm surge computations.

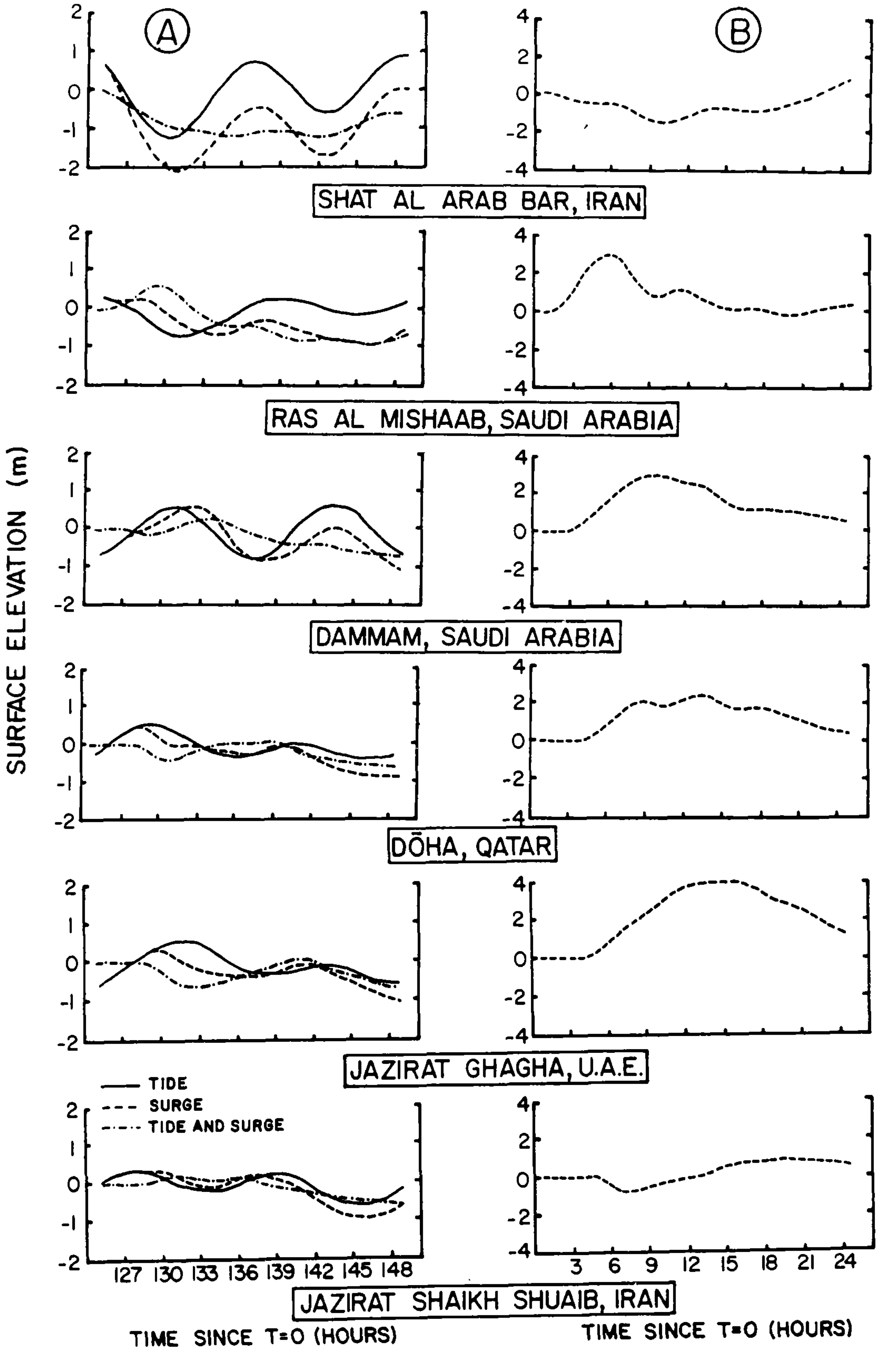

Figure 4. a) Computed profiles of the tide (six constituents M_2, S_2, N_2, K_1, O_1, P_1 are included), surge plus tide and surge alone at selected locations; b) Storm surge amplitudes computed for long duration Shamal. The wind blows constantly from the positive Y direction (from MURTY and EL-SABH, 1984).

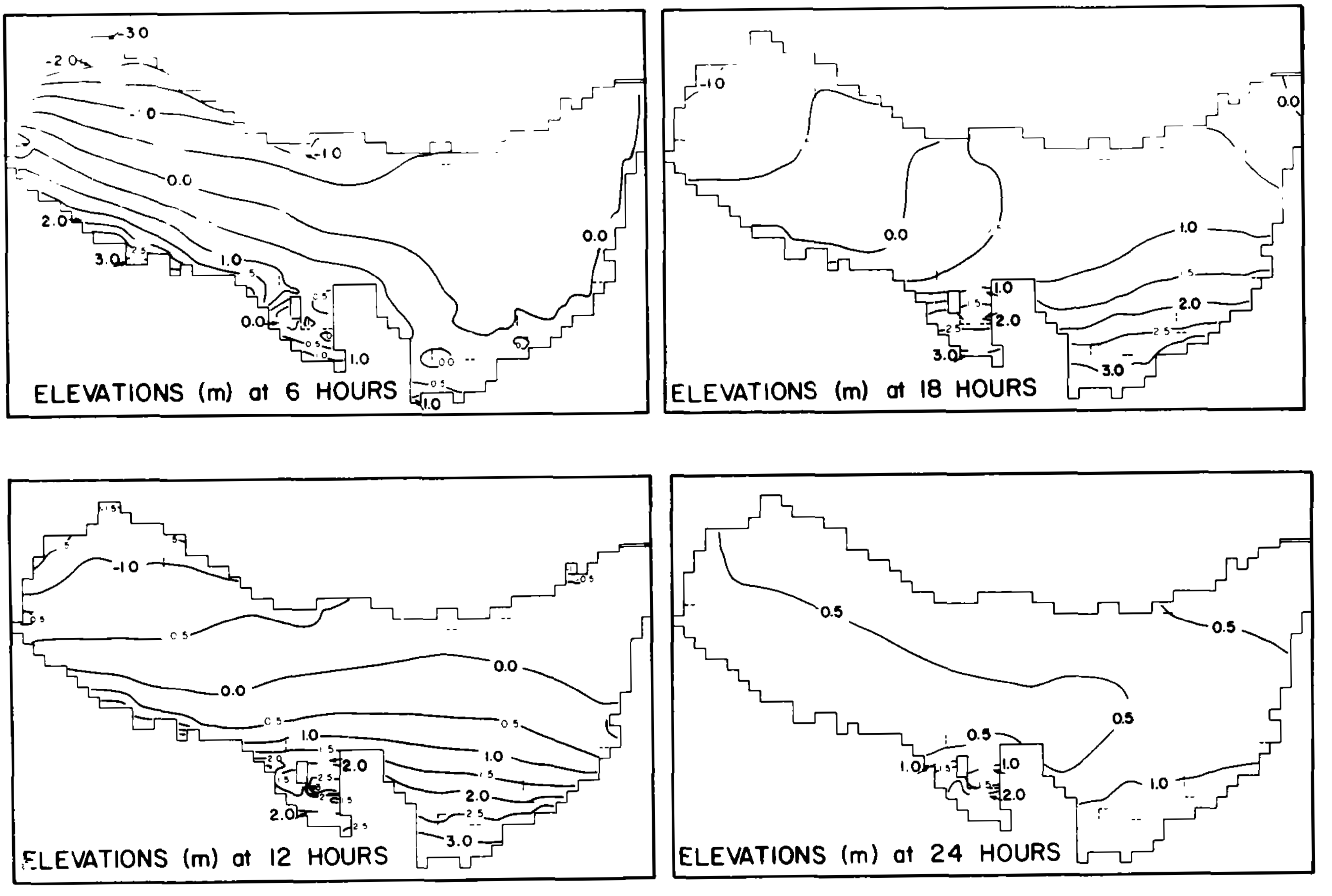

Figure 5. Distribution of storm surge heights (from MURTY and EL-SABH, 1984).

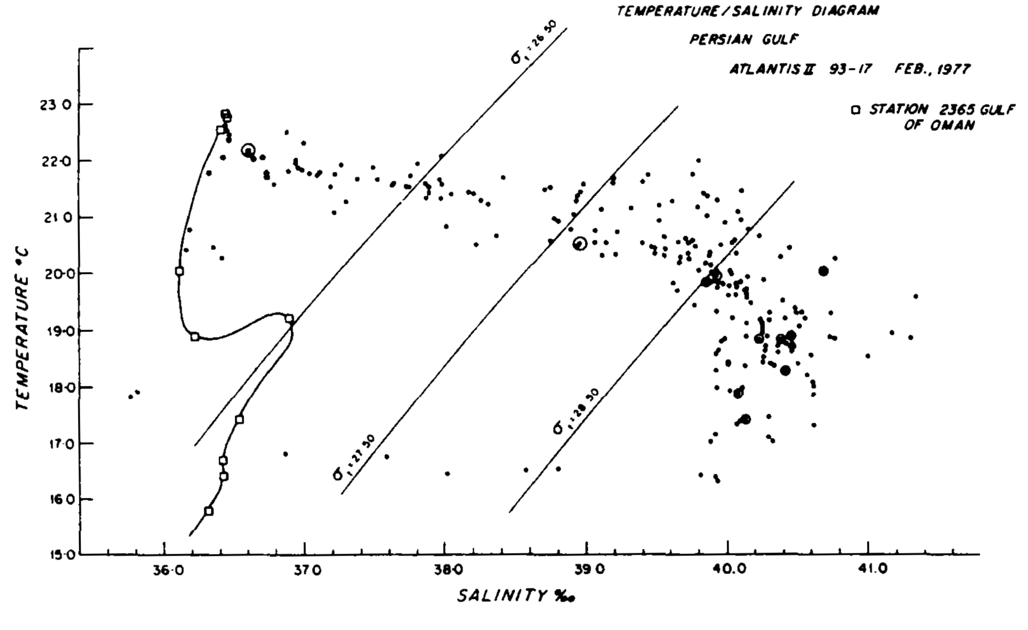

Figure 6. Temperature-salinity diagram for Inner Gulf Waters. The square symbols indicate Station 2365 in the Gulf of Oman; open circles indicate overlapping data points; February 1977 (from BREWER and DYRSSEN, 1985).

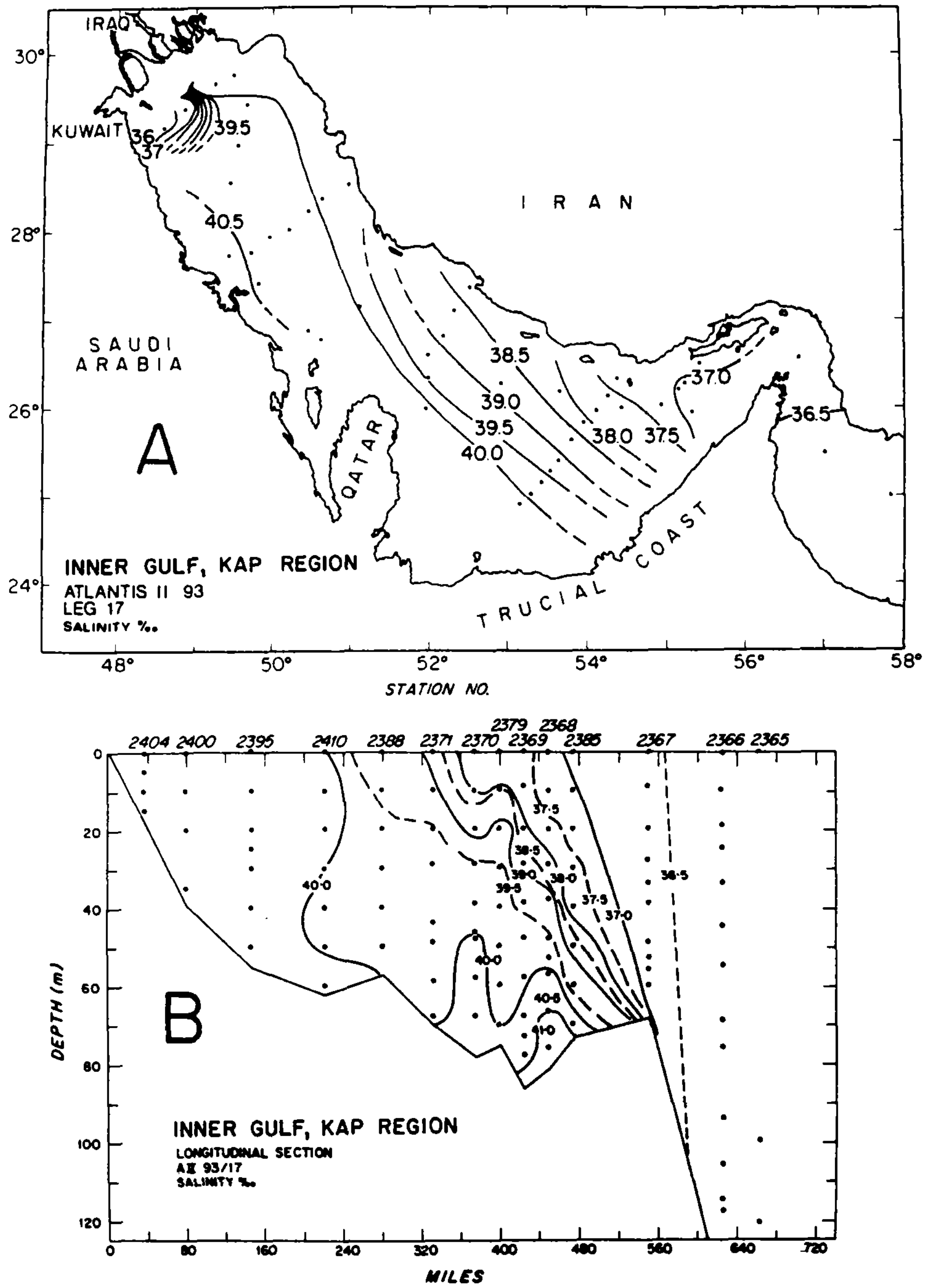

Figure 7. a) Map of surface salinity within the Gulf; b) Vertical salinity section along the major axis of the Gulf (from BREWER and DYRSSEN, 1985).

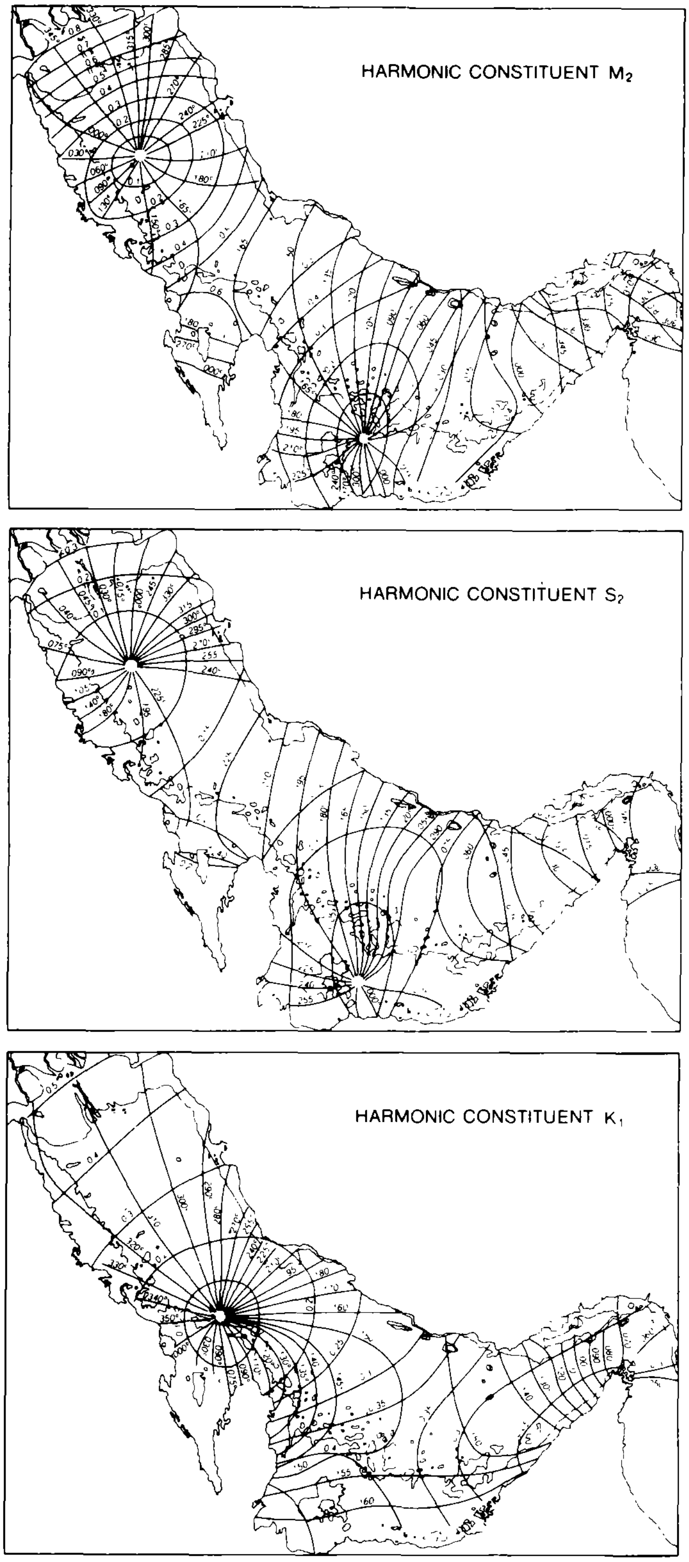

Figure 8. Observed M_2, S_2 and K_1 tide heights and phases in the Gulf.

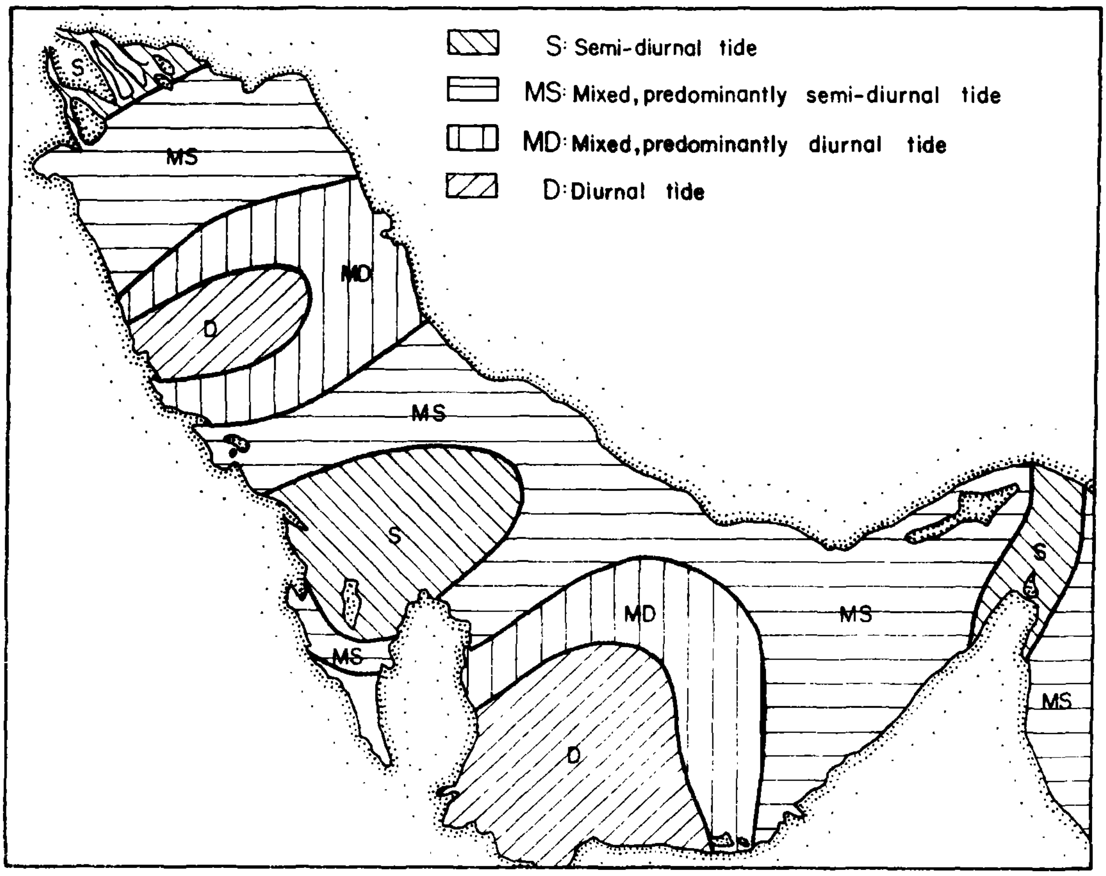

Figure 9. Classification of tidal types following EL-SABH (1982).

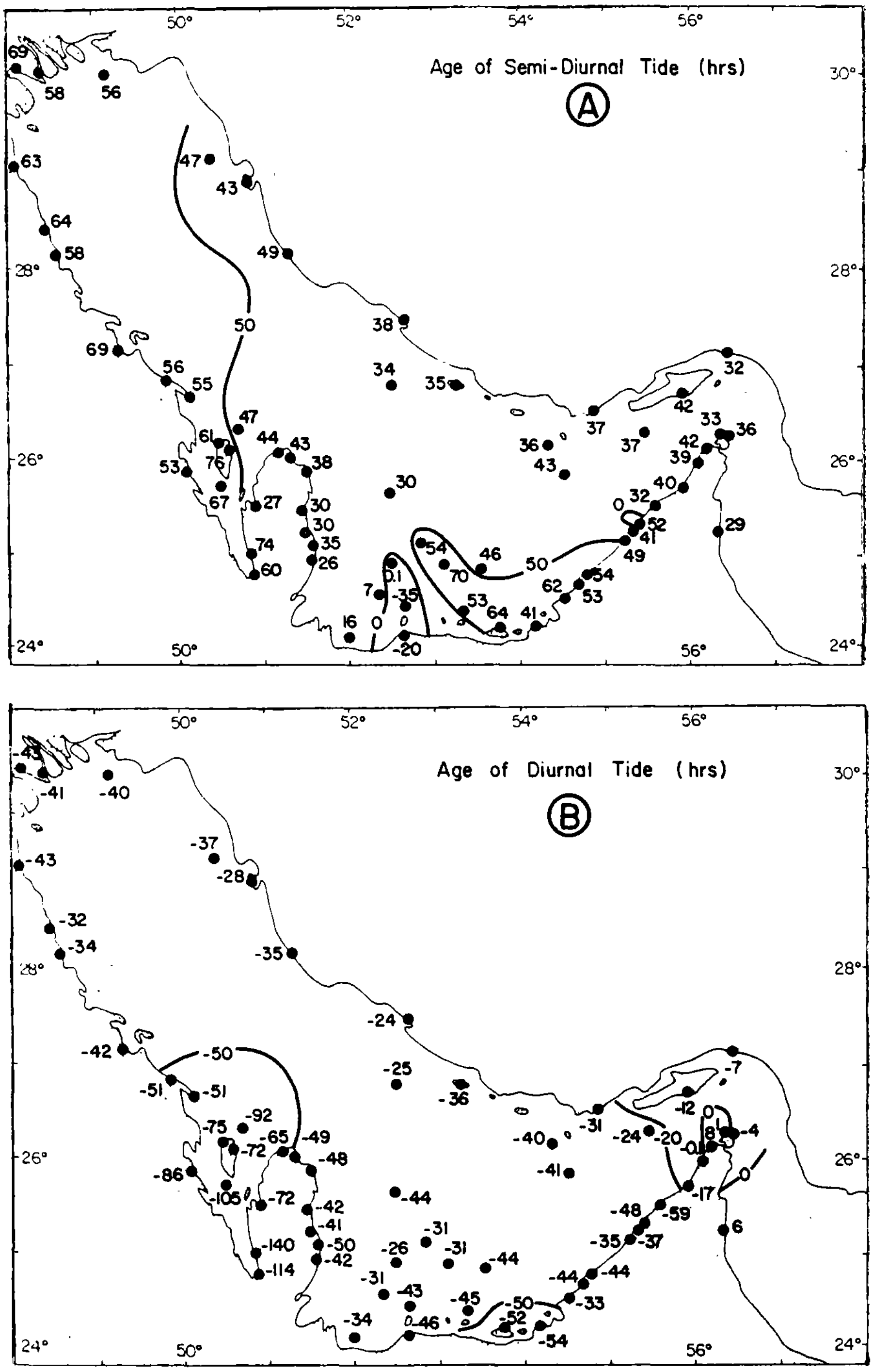

Figure 10. Age of the semi-diurnal and diurnal tides in the Gulf (from MURTY and EL-SABH, 1985).

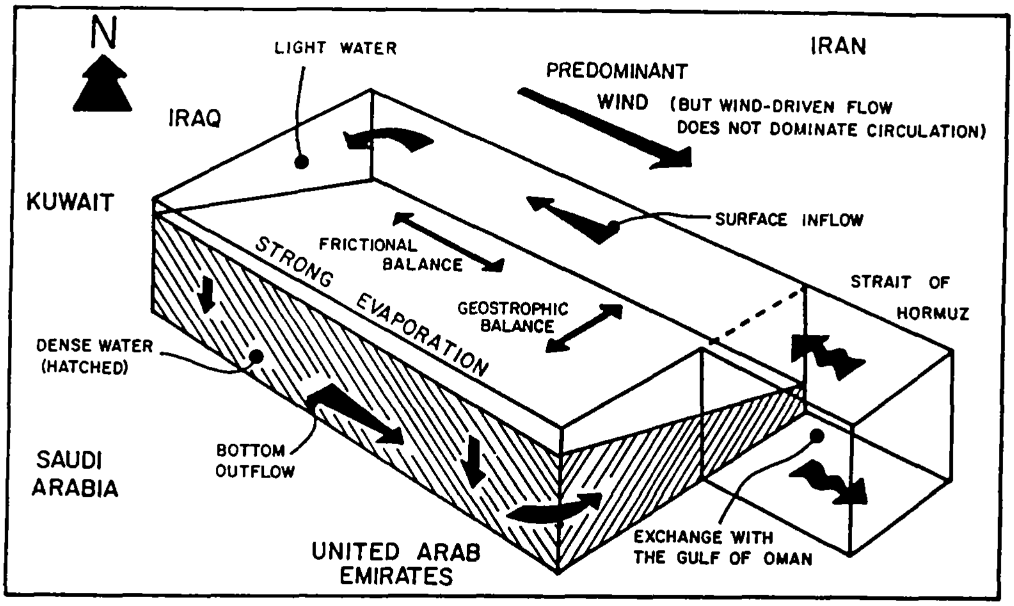

Figure 11. The probable circulation pattern in the Gulf (from HUNTER, 1982).

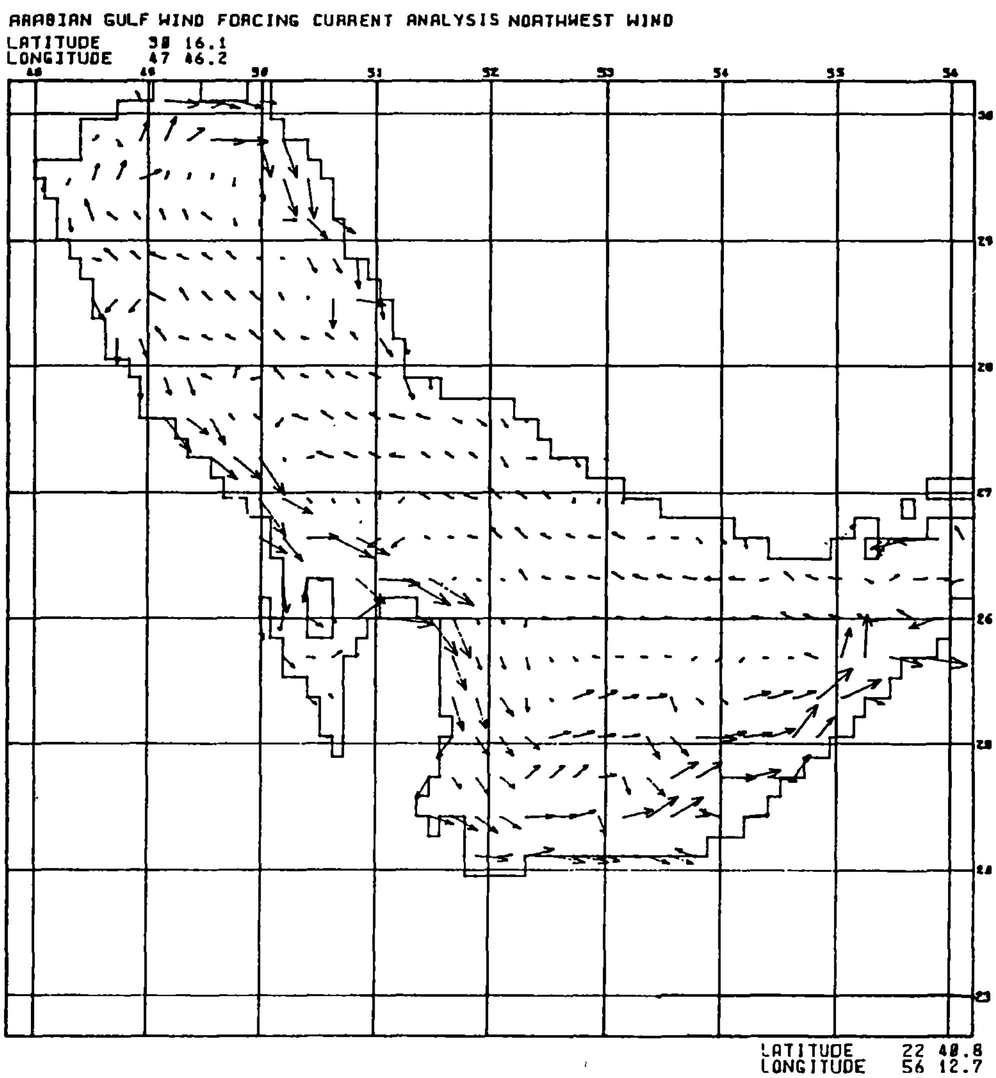

Figure 12. Steady state surface circulation in the Gulf (from GALT et al., 1984).

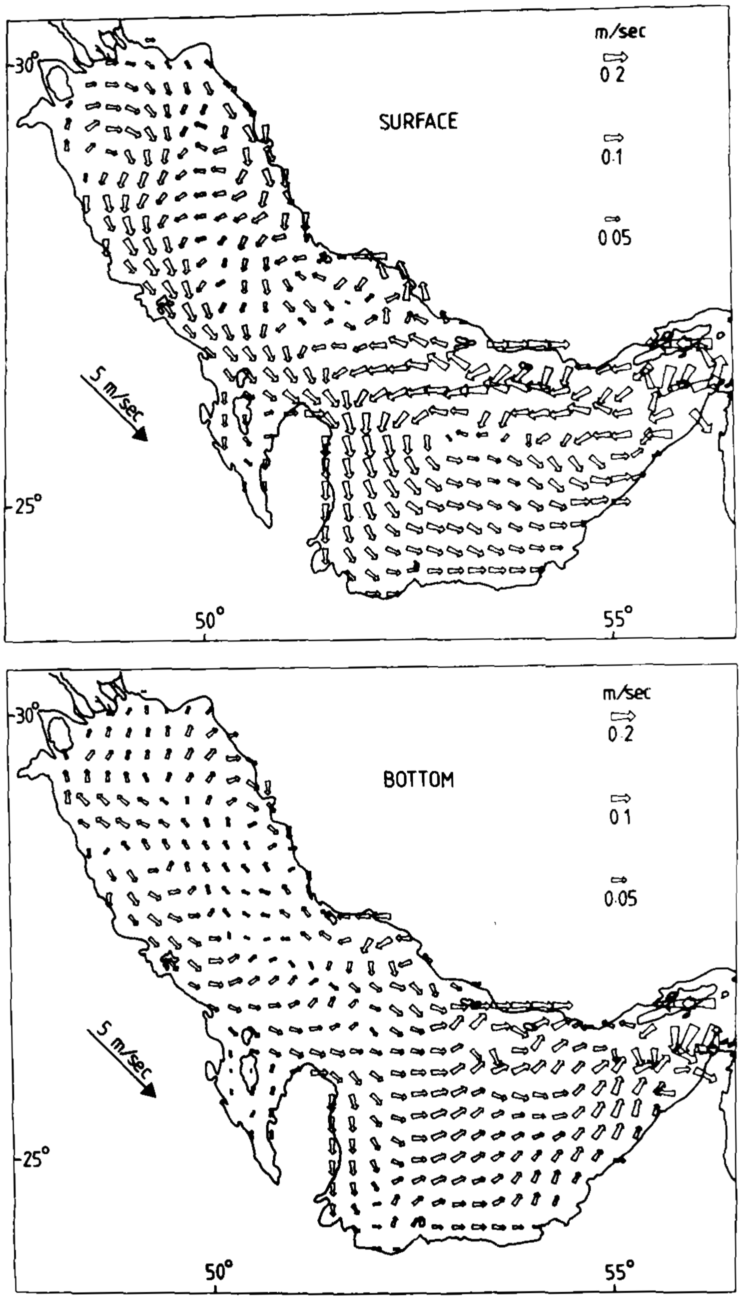

Figure 13. Predicted model velocities for surface and bottom cells with prescribed wind stress (vector lengths proportional to cube root of velocity) (from HUNTER, 1983).

FISHERY RESOURCES IN THE NORTH ARABIAN SEA AND ADJACENT WATERS

SIEBREN C. VENEMA

Fishery Resources Officer
Marine Resources Service, Fisheries Department
Food and Agricultural Organization
Rome, Italy

ABSTRACT

The Arabian Sea has drawn the attention of oceanographers and marine biologists because of its unique oceanographic phenomena. The reported upwellings and related high primary production off southwest Arabia, Somalia and the Malabar coasts of India have led to high expectations in terms of harvestable fish resources. Schaefer, for instance, estimated that the potential yield of the resources off southern Arabia, the Gulf of Aden and Somalia are the same order as that of the anchovy fisheries in Peru; viz., around 10 million tonnes. Subrahmanyan estimated the potential of pelagic fish off the west coast of India at 1,119,000 tonnes based on a comparison with the North Sea, while Shomura put it at 500,000 tonnes. He estimated a potential of demersal fish of 1,430,000 tonnes for India and Pakistan, and an additional 800-850 thousand tonnes for Somalia, the People's Democratic Republic of Yemen and Oman. The total pelagic and demersal production for the Arabian Sea was put at around 4 million tonnes based on Cushing's calculations. The above estimates were all very rough and can now be replaced by better ones for practically the entire area.

The rapid developments in stock assessment by acoustic methods, in particular the echo integrator in the late sixties, made it possible to survey large areas in a relatively short time span with one large vessel. A UNDP/FAO project aimed at assessing the resources off southwest India was subcontracted to the Institute of Marine Research in Bergen, Norway, and the R.V. RASTRELLIGER, built in Norway, became the flagship of the UNDP/FAO fleet of research vessels. FAO, through the Indian Ocean Programme, later entered into a trust fund agreement with NORAD for the construction and operation of the R.V. DR. FRIDTJOF NANSEN, an improved version of the RASTRELLIGER, and equipped with all the then available acoustic instruments for resource surveys. A plan for a two-year survey of the waters from Pakistan to northern Kenya was executed in 1975 and 1976, with an extension under a bilateral agreement for the first six months of 1977 off Pakistan. The existence of a large biomass of fish was confirmed; but the bulk of this biomass consisted of mesopelagic fish, a resource suitable for the production of fishmeal, the economic feasibility of the harvesting is not yet known.

In 1979, the R.V. DR. FRIDTJOF NANSEN returned to the Gulfs of Oman and Aden to study further the resources of mesopelagic fish, and in 1981 a similar survey was followed by short surveys in the waters off Djibouti and the Gulf of Suez. In 1983, the DR. FRIDTJOF NANSEN, funded by NORAD and a global UNDP/FAO project, returned once again to the Arabian Sea for a longer-term programme. The harvestability of mesopelagic fish was tested with success in the Gulf of Oman in January/February. The other resources of Oman were assessed in March. After a short survey in the Maldives in August, the first of its kind in that country, the vessel proceeded to Pakistan and Iran to start another series of surveys in that area, followed by others off Oman and in the Gulf of Aden.

OCEANOGRAPHIC CONDITIONS, PELAGIC PRODUCTIVITY AND LIVING RESOURCES IN THE GULF OF ADEN

J.STIRN*, R.EDWARDS*, J. PIECHURA*, M. SAVICH‡, M. GHADDAF, M. FADEL, F. MUTLAQ, O. SABIH, A. SAYED, S. SHAHER AND Z. ZUBAIRI

Marine Science and Marine Resources Research Centre
Aden, P.D.R. Yemen
*UNESCO Field Project 703/PDY/40
‡Azchemiro Kertch, U.S.S.R. (bilateral assignment to P.D.R. Yemen)

INTRODUCTION

The Gulf of Aden extends east-north-eastward from the narrow Strait of Bab-al-Mandab to a line between Ras Bag-hashwa (east of Mukalla, P.D.R. Yemen) and Ras Asir (northeastern corner of the Somali Peninsula) where it opens to the North Arabian Sea. Its approximate surface area is 220 x 10^3km^2, and the average depth is about 1,800 m.

Not as much for its size (which is quite considerable) but for a number of formidable oceanographic phenomena, the Gulf of Aden presents a rather unique, large ecosystem which deserves scientific attention. In addition, its extraordinary biotic richness traditionally has provided considerable amounts of sea food for the inhabitants of the surrounding arid lands. Much more sea food can be expected in the future, provided a rational fisheries development is based upon the scientific knowledge of this ecosystem and its exploitable potential.

The Gulf of Aden was "a highway for international trade almost at the start of recorded history" (BEHRMAN, 1981). It also lies on the route of most research ships working in the Indian Ocean. However, surprisingly little information on its oceanography or bioproductivity has been gathered. Even the International Indian Ocean Expedition (IIOE 1962-65), subsequent compilations of collected data (IIOE atlases: IOBC/PANIKHAR, 1968-70; KREY and BABENARD, 1976; WYRTKI, 1971) and numerous resulting scientific papers did not fill the the gaps in knowledge significantly. Although these documents gave the main oceanographic and bioproductivity characteristics of the Gulf of Aden, they were extrapolated from very few data and thus could not adequately describe the real conditions; e.g. primary productivity is interpreted on the basis of 8 carbon assimilation and 35 chlorophyll *a* measurements covering all seasons and depths. Using Behrman's expression that prior the IIOE the Indian Ocean was a "forlorn ocean", so the Gulf of Aden *during* the IIOE was a rather "forlorn gulf". Fortunately, in the adjacent regions of the Arabian Sea, namely Omani shelf-waters and the area of the Somali Current, where more systematic investigations were focused during and after the IIOE, some detailed studies (e.g. BRUCE, 1973; CURRIE et al., 1973; SCHOTT, 1983; SMITH, 1982, 1984; SWALLOW and BRUCE, 1966; WARREN et al., 1966) were carried out on phenomena whose principles apply also to the oceanographic- bioproductivity conditions of the Gulf (e.g. upwelling, de-oxygenation, etc).

Based on information available since the IIOE, the only oceanographic research in this region has been carried on by Soviet research vessels, and by FAO/Norway fisheries research cruises of R/V Dr. F. Nansen (1975-76 and 1984). The great majority of these cruises, however, emphasized fish stock surveys and other applied purposes, thus oceanographic research, using only just the most essential hydrographic measurements, was a second priority. In spite of that, however, a number of important scientific papers on the hydrography, nutrients, phytoplankton and zooplankton significantly increased the specific knowledge on the Gulf of Aden (GAPISHKO, 1971; KHIMITSA, 1968a,b; SAVICH, 1968, 1969, 1979; SERIY, 1968; SUKHANOVA, 1969). Of course the most important contribution of these investigations was the extensive knowledge on exploitable sea-food resources of this region which will be summarized in this paper.

BASIC OCEANOGRAPHIC FEATURES OF THE GULF OF ADEN

Previous pertinent knowledge has been summarized by the late K. Grasshoff, Y. Halim, S. Morcos and G. Siedler (in HALIM et al., 1980); we loosely quote as follows:

The temperature-salinity-density structure of the area is mainly controlled by the monsoon-related wind and atmospheric pressure field and its seasonal variations, and by the water exchange through Bab-al-Mandab due to the thermohaline circulation caused by intensive evaporation (approximately 2 m) in the Red Sea. As a result, there exists a two-layer structure in Bab-al-Mandab. The high salinity bottom water (up to 40×10^{-3}) flows over the sill and through the Strait down slope into the Gulf of Aden, spreading horizontally at a depth of about 500 m.

The hydrographic structure of the Gulf of Aden is characterized by a surface layer of warm, fairly saline water between 20 m (winter) and 120 m (summer) in thickness, below which is a decrease in temperature and salinity (maximum 36×10^{-3}) and a reduced vertical temperature gradient, caused by the spreading Red Sea water. At greater depths down to 2,500 m is the upper Indian Ocean deep water (5°C, 35×10^{-3}).

Average surface currents are directed into the Gulf of Aden and the Red Sea during winter and out of the Gulf during summer. Gyre-type motions in the Gulf of Aden are indicated by available geostrophic current data.

Upwelling is significant off the Yemeni coast during the SW Monsoon (summer) where Ekman transport tends to carry coastal surface waters away from the shelf, thus causing convective motions on the shelf. CURRIE et al. (1973) discussed upwelling in Omani shelf waters, but did not provide any evidence of the upwelling in the Gulf of Aden itself. Upwelling also may be significant off Socotra where major changes in the circulation pattern result from the seasonal reversal of the monsoon. Though it is more intensive along the coast, a significant feature of the south Arabian upwelling is its width and the much greater depth of the water supplied than is usual in other coastal upwelling regions.

Short-term atmospheric events may cause considerable deviations from the mean conditions, particularly close to the coast. These variations in the hydrographic structure have a strong effect on the distribution of nutrients, oxygen and chlorophyll (WYRTKI, 1971; KREY and BABENARD, 1976) and, therefore on the primary production and the fish population.

The hydrochemical regime of the Gulf of Aden is marked by a number of essential features which also have a significant influence on the concentration and distribution of nutrients. Together with the distribution of oxygen, they certainly influence the abundance of fish, especially the plankton feeders. Very little is known about the variations of oxygen and nutrients in the euphotic layer, both in time and space.

The general water circulation in the upper layers, particularly the upwelling processes that occur along the Yemeni coast and south of Socotra, is strongly dependent on the monsoon, its temporal variations, and on the topography of the coast line. Compilation of data from the International Indian Ocean Expedition (IIOE) suggests that the depth of the mixed layer in the Gulf region may be as shallow as 20 m or less during the period of the SW monsoon in April through September, whereas it is rather deep (>100 m) south of Socotra during this period. From October to February the situation changes completely, creating a deeper mixed layer along the Yemeni coast (>60 m) and a shallower one (>40 m) during March through June. It must be stressed, however, that the number of observations and occupied stations in the region are too few to allow any firm conclusions or detailed analysis.

In the upper thermocline, nutrients have high gradients (0.3 - 2.0 ug-at l^{-1} for phosphate-phosphorus, <0.5 - 20 ug-at l^{-1} for nitrate-nitrogen and <3 - >20 ug-at l^{-1} silicate- silicon). The content of oxygen varies between saturation at the ambient temperature (~4.5 - 5 ml l^{-1}) in the surface layer and values of less than 0.75 ml l^{-1} below the thermocline. Any degradation of the gradient in the upper thermocline would facilitate an upward-directed transport of the accumulated nutrients in the sub-thermocline layers. But again the scarceness of data does not allow any detailed analysis of these important processes. It is believed that large short- and long-term variations occur, particularly near the Yemeni coast, both in space and time. These are assumed to be closely related to the prevailing wind and its fluctuations.

Low oxygen content in subsurface waters suggests a long residence time, but the reported high productivity in the euphotic layers, which may be significantly deeper than the upper thermocline (reported values are 60 - 80 m for the euphotic zone in the Gulf of Aden), may cause a high oxygen consumption through the subsequent remineralization processes. The high silicate content suggests, however, a relatively long residence time. This, in turn, has a consequence not only for the oxygen consumption and nutrient accumulation processes but also for the eventual accumulation of harmful substances, such as chlorinated hydrocarbons, petroleum hydrocarbons and trace metals.

The upwelling of subsurface waters is the principal mechanism by which the upper layers of the west Arabian Sea are enriched with nutrient salts. Though the data coverage leaves large gaps in time and space, this part of the Arabian Sea appears to be one of the most productive parts of the Indian Ocean. The mean production of the entire west portion of the Indian Ocean, up to several hundred kilometers offshore, is several times that for the world's oceans. Primary production is as large, if not larger than that encountered in upwelling regions along the coast of Peru and off West Africa. One of the most important findings of the IIOE is the extremely high rate of primary production and the large standing crops of phytoplankton and zooplankton in the Arabian Sea, especially along the west side (WOOSTER et al., 1967).

The inner waters of the Gulf of Aden and those off Ras-Fartak and Socotra show chlorophyll *a* values of about 0.5 mg m^{-3}. The potential productivity in the upper 50 m exceeds 1.00 mg C m^{-3} hr^{-1}. In autumn, surface photosynthesis is moderately high throughout the area, but is decreasing. It is, however, higher off Ras-Fartak and Socotra. During the NE monsoon (December-February), chlorophyll *a* is low, and so is the potential productivity (McGILL and LAWSON, 1966).

The highest zooplankton volumes for the Indian Ocean are found in the Arabian Sea off the Somali coast and Cape Fartak and further east. Here average values as high as 54.7 ml m^{-2} have been obtained; the rest of the Indian Ocean yields values less than 15 ml m^{-2} (RAO, 1973). During the NE monsoon, a biomass of 20 - 40 ml m^{-2} occurs in the region of the Gulf of Aden, but occurrences of higher volumes up to 80 ml m^{-2} have been identified more offshore, between Socotra and Ras-Fartak. The period of the rising SW monsoon (May-June) yields a comparable volume of 20 - 40 ml m^{-2} between Ras-Fartak and the Gulf of Aden. During the full SW monsoon period (July-September) high biomass values of 40 ml or above are found around Socotra. Unfortunately data from the rest of the Yemeni waters are too scanty for comparison. The transition period between the end of the SW monsoon and the beginning of the NE monsoon shows a general rise in the biomass, with a dense patch of up to 80 ml m^{-2} off and around Cape Fartak. From this brief account of the seasonal distribution of the zooplankton biomass, it appears that Democratic Yemen waters, particularly in the area between Socotra and Cape Fartak, show a consistently high level of secondary production in spite of the seasonal fluctuations in primary productivity. It also appears that the area of higher density spreads out from the area of upwelling, in the direction of the prevailing currents.

Unfortunately, the authors of the above review (HALIM et al.) were probably aware of more recent results, obtained by Soviet investigations of the Gulf of Aden (references of available sources see above). On the one hand these more recent and extensive data have substantiated many of the review's statements, particularly for primary and secondary productivity. On the other hand they provide clear evidence on some oceanographic phenomena of a crucial importance, such as the four-layer structure of the Gulf's water masses, development of upwellings in the Gulf itself, etc.

Nevertheless, one has to agree fully with the following introductory statement of the above quoted review: "... development and management of the fisheries resources is (sic) dependent on a sound and comprehensive knowledge of the physical, chemical and biological conditions which control those resources. Unfortunately, these conditions are, as yet, inadequately known ... basic hydrographic information is still fragmentary and knowledge about water structure and circulation is scarce" (HALIM et al., 1980).

Recognizing these gaps in knowledge, both as related to applied fisheries and to the general objectives of basic marine sciences, the recently established Marine Science and Marine Resources Research Center (MSRC)[1] started in 1984 with systematic oceanographic and pelagic bioproductivity investigations of the Gulf of Aden. Since then a considerable amount of data has been gathered which is used in this paper to discuss oceanographic and bioproductivity conditions in this region.

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1 MSRC as a unit of PDRY Ministry of Fish Wealth was established in 1983, partly through UNESCO Fund-in-Trust Project 703/PDY/40, financed by the Islamic Development Bank, Jeddah.

Simultaneously, rather extensive fish stocks surveys were carried out in this region by MSRC and FAO/Norway cruises of R/V Dr F. Nansen, allowing us to consider exploitable resources of this region for the first time interrelated with the complex and oscillating ecosystem of the Gulf of Aden.

Preparation of this paper would not have been possible without the generous support and cooperation from the Institute of Marine Research, Bergen, Norway. In addition to allowing us to use their fish stock and hydrographic data, ship facilities were provided for our research team during the summer of 1984 when our ship could operate due to heavy monsoon seas. Assistance of scientists bilaterally assigned to PDRY from USSR is gratefully acknowledged.

MATERIAL AND METHODS

OCEANOGRAPHY AND PELAGIC BIOPRODUCTIVITY

Since February 1984, 10 monthly and 4 seasonal cruises were carried out on R/V Ibin Magid (except in August 1984, by R/V Dr. F. Nansen); stations are shown in Figure 1. Standard hydrographic sampling and measurements were performed at all stations down to a depth of 1,000 m when applicable. Hydrographic samples were taken by Nansen bottles, nutrient/phytoplankton samples by atoxic Van Dohrn samplers. Samples for nutrients were taken at the surface, 50, 100, 200, 500, 800 and 1,000 m; phytoplankton/chlorophyll at surface, 50 and 100 m. Zooplankton samples were taken from vertical hauls (twice at 30 - 0 m, one sample being preserved deep-frozen for subsequent biomass determination, and at 100 - 0 m) with a WP-2 net (mouth 0.25 m^2, mesh 200 µm).

Except for pH and oxygen, which were measured on board, all samples were preserved and deep frozen (chlorophyll samples were filtered immediately after the sampling and dry filters frozen) because laboratory facilities were not available on the ship. Consequently, subsequent analyses, although carried out soon after the cruises, showed quite doubtful results for nitrite and ammonia, while the rest of nutrient analyses delivered consistent results.

Salinity was determined by the Grasshoff modification of Mohr-Knudsen titration. Nutrients were analyzed spectrophotometrically: phosphate after Murphy and Riley, ammonia after Solorzano as modified by Koroleff, nitrite after Bendschneider and Robinson, nitrate by the Grasshoff modification of the Morris and Riley reduction method, silicate after Koroleff modification of the Chow and Robinson procedure. Chlorophyll determinations were carried out basically as recommended by SCOR/UNESCO; however, the Humphry trichromatic formula was applied. Zooplankton biomass was obtained as displacement volume, dry weight (70°C) and ash-free weight (500°C). Processing of preserved zooplankton samples for quantitative taxonomic studies followed procedures as recommended by FAO/STIRN. Quantitative taxonomic analyses of preserved phytoplankton were carried out by the improved Uttermoehl settling-inverted microscopy technique recommended by SCOR/UNESCO.

FISHERIES BIOLOGY

Demersal Fish

These were surveyed by the following trawling methods:

- R/V Ibin Magid trawl of 28 m headline with a vertical opening of 4 m. The mesh in the codend was normally 65 mm but sometimes was changed to 120 mm when trawling for cuttlefish. Trawls were taken at 3.2 to 3.6 knots for 1 to 3 hours duration; depths were between 15 and 120 m. In total, during 1983-85 on 9 cruises over 300 exploratory trawls were made, covering the whole stretch of Yemeni coastal waters, yet focusing upon traditional trawling grounds in the eastern part.

R/V Dr. F. Nansen trawl of 41 m headline with a vertical opening of 6 m. The codend contained a fine mesh liner of 1-m mesh. Trawling speed was 2.8 knots and trawls were of 0.5 hour duration. Demersal fish stocks were assessed by the trawl swept area method, based on the distance between the wings of the 28 m headline trawl of 20 m and the 41 m headline trawl of 25 m. The area covered by the former was around 12.8 ha per hour and the latter 15.5 ha per hour. Catches were separated into species which were then counted and weighed, or estimates of catch weight made from baskets. Random samples were taken for length and weight measurements.

Pelagic Fish

These were surveyed by 3 scientific echo sounders (120, 50 and 38 kHz), 2 echo integrators and 1 net sonde (50 kHz). The pelagic trawl had a 48 m headline and an opening of around 10 m. A fine mesh codend liner was used. The trawl catch was used to identify species in fish schools.

Pelagic fish studies on the coast were made from purse seine landings where catch, effort, speciation and age composition were recorded. These data were used for virtual population analysis (VPA).

RESULTS

The aim of this paper is not to present the results of our recent investigations in the Gulf of Aden in any detail, which will be done elsewhere, but rather to show selected data in order to give an overall view on the most crucial oceanographic conditions and processes as related in particular to pelagic productivity and living resources of this region.

HYDROGRAPHIC CONDITIONS

Water Masses

The upper 1,000 m layer of the Gulf of Aden appears to have a 3-layer structure (Table 1) as follows:

Surface Water (SW) occupies the uppermost layer, ranging in depth from 20 m down to 125 m. Its lower boundary usually coincides with the thermocline. It forms locally by intensive heating and evaporation, thus having high temperature and salinity, and is relatively well saturated with oxygen except during the summer in upwelling zones. The volume and physical properties of SW have pronounced spatial and temporal changes (Figs. 2-8). Usually the lower boundary lies deeper in the eastern and western parts of the Gulf and is shallower in the center as well as on the shelf. Temperature and salinity decrease slightly from west to east and from shores towards the center. However, temporal changes are much more pronounced than spatial ones. During summer, due to wind-driven water circulation (southwest monsoon) and upwelling, the surface layer is the thinnest; it occupies only the upper 20 m of waters along the Arabian coast and up to 70-80 m in the central part.

Upwelling of the subsurface water raises the thermocline, which also leads to low oxygen content, about 1 ml l^{-1}, in the lower layers of the surface water. The highest temperature (26-32°C) and salinity ($35.8 - 36.8 \times 10^{-3}$) of the SW are observed, however, a bit earlier: in May-June, when the heating and evaporation are most intensive. The thickest SW, 100 - 150 m, is observed during winter time, when surface cooling and density mixing depress the thermocline. Usually at this time temperature and salinity reach their minimum values. The lower boundary of SW can be placed along the isopycnal surfaces of 24 - 25.

Table 1 Physical chemical properties of the water masses of the Gulf of Aden as measured during 1984-1985.

Water Masses	Depth Range (m)	T°C	($S \times 10^{-3}$)	σT	0_2 ($ml\ l^{-1}$)
SURFACE	0	24.0-30.0	36.0-36.8	22.0-24.0	4.0-5.0
WATER	20-125	16.0-25.0	35.4-36.3	24.2-25.4	1.0-4.5
SUBSURFACE	70-400	13.0-28.0	35.3-35.8	26.0-26.5	0.2-1.2
WATER	200-500	13.0-16.0	35.5-36.5	26.6-27.2	0.4-1.0
RED SEA	400-900	11.0-17.0	35.8-37.6	27.2-27.6	0.4-1.2
WATER	1000 and over	8.0-12.0	35.0-36.2	27.4-27.6	0.5-1.0

Subsurface Water (SSW) is distinguished by an intermediate salinity minimum, oxygen minimum and high nutrient content. It is located between the SW and a depth of 400-450 m. This water originates most probably at the subtropical front in the Indian Ocean (SERIY, 1968) and comes into the Gulf of Aden from the Arabian Sea. The general direction of its movement is towards the Bab-el-Mandab Strait; it rises up a bit on its way west, and its temperature, salinity and oxygen content increase slightly.

Seasonal changes of the SSW, though much smaller than for SW, can be still distinguished. The highest values of temperature (15-18°C) and salinity (36.0 - 36.5 x 10^{-3}) are observed in May-June, and the lowest (13 - 15°C and 35.3 - 35.8 x 10^{-3} respectively) during southwest monsoon and afterwards. Oxygen content is very low, usually below 1 ml l^{-1} and very often below 0.5 ml l^{-1}. The core layer of SSW has a density of about 26.0 - 26.5.

Red Sea Water (RSW) As this mixed water comes through the Strait of Bab-el-Mandab, it sinks into the Gulf of Aden and moves generally towards the east, at depths between 400 m in west down to 1000 m towards the mouth of the Gulf. The depth of the core layer of RSW increases to the east, and its temperature, salinity and oxygen content decrease in the same direction. Red Sea Water is distinguished by higher temperature, salinity and oxygen content and lower nutrient values. There are some seasonal changes of RSW indexes, partly due to seasonal variability in water exchange through the Strait of Bab-el-Mandab. The inflow into the Gulf of Aden increases during winter and spring, with the maximum in May, and decreases in July-August (SERIY, 1968). So higher temperatures and salinities are observed by the end of winter and in the spring, and lower values during summer and autumn. The most probable density in the core layer of RSW is 27.2 - 27.6. Our data clearly show this water mass (Figs. 6 and 7) in the western part of the Gulf, and much less (or not at all) in the east.

Water Temperature

Other than solar heating and heat exchange with the atmosphere, the thermal structure of Gulf of Aden waters is formed by water circulation and inflow and outflow from the Red Sea and Arabian Sea. Water circulation and upwelling particularly cause the redistribution of the heat in the upper layers to such an extent that along the Arabian coast the minimal surface temperature is recorded in summer (Fig. 4), when the maximum should be expected. Generally the temperature at the surface during late spring, summer and early autumn increases from the Arabian coast towards the Somali coast. Sometimes, during May-June and September-October particularly, maximum values are recorded in the center of the Gulf where an anticyclonic gyre is formed. During the remaining part of the year, October-November through April, the picture is reversed, temperature generally decreasing towards African coast, or minimal values occurring in the center in the cyclonic gyre.

The vertical distribution of temperature shows a shallow, seasonally variable surface mixed layer: about 20-100 m. The depth of thermocline and the vertical gradient within it are changeable as well. In spring and summer a secondary thermocline is observed near the surface. Below the main thermocline temperature decreases slowly towards the bottom. In the western part of the Gulf, close to the Bab-el-Mandab Strait, a small increase of temperature can be observed at the depth where RSW is spreading.

Seasonal changes of water temperature, due to incoming solar energy variability and to monsoon wind-induced circulation, are usually restricted to the surface layers. Surface temperature usually reaches its highest values, over 30°C by the end of May - beginning of June, when there is high input of solar energy, weak and variable winds and a shallow secondary thermocline (Fig. 3).

In July-August, when surface heating is even greater, the temperature is a bit lower, on average about 0.5°C, than in May-June, due to strong SW monsoon winds, intensive mixing and cooling by upwelled waters. In areas where coastal upwelling is strong and reaches the surface, temperature can fall as low as 17°C (Fig. 5). The upper mixed layer remains thin; the thermocline is pushed up by upwelling and sometimes destroyed. Vertical gradients in the thermocline increase and in some regions can reach maximum values (Fig. 2c). In autumn the surface temperatures first increase slightly to about 30°C, and then from October they decrease to a minimum in February (about 25-25°C). At the same time the surface mixed layer becomes thicker due to surface cooling and convection, reaching down to about 100 m or more (Fig. 2c). The thermocline deepens to a depth of about 80-150 m, and in some cases the vertical gradient within it increases. Below the thermocline much smaller seasonal changes of temperature can occur, particularly in the RSW due to seasonal changes of volume and temperature of incoming waters. In the areas with a strong upwelling, summer decreases in temperature also can be observed in the deep waters down to a depth of 500-700 m and more (Fig. 3).

Salinity

The vertical and temporal changes of salinity in the Gulf of Aden are small compared to horizontal variability. The main factors which increase salinity are evaporation and inflow of highly saline RSW. Water circulation further complicates the picture, but at all depths salinity generally increases towards the Bab-el-Mandab Strait and towards the African coast. On the surface the differences in salinity between the far northeast region and the vicinity of Bab-el-Mandab can reach 1.5×10^{-3}, but usually they are $0.5 - 0.8 \times 10^{-3}$. Sometimes during the winter monsoon particularly, the maximum values are observed in the central part of the Gulf where the centers of anticyclonic gyres are located.

Vertical profiles show two layers with distinctly higher salinity and two layers with lower values. The surface layer has higher salinity (36.0 to over 37.0×10^{-3}) due to intensive evaporation throughout the year and nearly non-existent precipitation and river inflow. Below this, to about 400 m, salinities less than 36×10^{-3} are observed, as described above. Below this depth salinity increases significantly, in some regions to over 37×10^{-3} within the layer of the RSW, and then decreases in the bottom water.

Seasonal changes in salinity usually do not exceed 0.5×10^{-3}, depending on intensity of evaporation, water exchange with adjacent seas and water circulation (KHIMITSA, 1968b). In the surface layer high salinities are observed during winter, spring and in the early autumn, and the low ones in the summer and in early winter (Figs. 5 and 6). In the deeper layers, except in the RSW, seasonal variations of salinity are very small, usually less than 0.2×10^{-3}. At about 500-800 m (RSW) the annual variations, particularly in the western part of the Gulf, can exceed 1.5×10^{-3}, with higher values in the winter and in the early spring and lower in the summer (Fig. 5). To the northeast annual changes in this layer are much smaller (Fig. 6).

Water Circulation and Upwelling

The main driving force of surface currents in the Gulf of Aden, the wind, has a pronounced seasonal pattern which is manifested in the strong and quite steady southwest monsoon wind, a weaker northeast monsoon and two intermediate periods with the weak and variable winds. The southwest monsoon is usually restricted to July-August and northeast monsoon to November through February-March, so twice a year the surface water circulation is restructured by the prevailing winds. Winter northeast winds cause surface waters to move generally towards Bab-el-Mandab, particularly along the Arabian coast (SERIY, 1968). Along the African coast there is a weak current in the opposite direction, and a cyclonic eddy is observed in the central part of the Gulf. Its center changes position in time and according to changes in the wind field. The alongshore Ekman transport of the water has a shoreward component, and accumulation of surface water on the shelf is usually observed.

During the summer monsoon with southwest winds, the surface water moves generally in the opposite direction, e.g. towards the Arabian Sea. The alongshore wind stress, with the coast to the left (northern Hemisphere), causes a net offshore transport in the surface Ekman layer. To preserve the water balance the deeper subsurface water replaces surface waters which have been pushed away. Our observations, made in August 1984 on R/V Dr. F. Nansen, show a picture of upwelling along the Arabian coast of the eastern Gulf of Aden. It is known (SERIY, 1968) that in the western part of the Gulf the southwest monsoon wind is much less stable and that it ceases earlier. The presence of much colder and low oxygenated waters in the vicinity of Aden (Figs. 3 and 4) in the subsurface layer below 20 m indicates upwelling, even if not observed at the surface. Surface temperature distribution (Fig. 5) shows three partly separated sources of coastal upwelling, with the lowest temperature below 17°C in the vicinity of Mukalla.

Such "patchiness" in coastal upwelling has been noted in other upwelling regions (BOJE and TOMCZAK, 1978). The temporal and spatial scale of the circulation is usually of smaller magnitude than can be resolved by the observations (CURRIE et al., 1973). A striking feature of this upwelling is its extent and influence over a very wide area of the Gulf, as seen in the temperature and oxygen distributions at a depth of 20 m (Figs. 5 and 10). In the main coastal upwelling regions, the width of the active upwelling is roughly equal to the Rossby radius, which is tens of kilometers (HUYER, 1983). Such an offshore extent of upwelling has been described further eastward, along the southern Arabian peninsula. BOTTERO (1969) confirmed that upwelling extends at least 400 km offshore and parallels the Arabian coast for a distance over 1,000 km. This phenomenon also was mentioned by SWALLOW (1984) and CURRIE et al. (1973). Because of this, the upwelled water must be supplied from much greater depths than is usual in other coastal upwelling regions (CURRIE et al., 1973). Our data, density particularly, indicate the upwelling from as deep as 500 m.

Usually in coastal upwelling regions there are strong undercurrents flowing opposite to the surface current direction. However, in our case the distribution of physical properties does not indicate an undercurrent.

Some indications of weak and restricted upwellings are also observed during the rest of the year. This can be caused by short-term fluctuations (periods of several days) in wind stress, even when the long-term wind stress is not favourable.

NUTRIENTS

Inorganic Phosphate

As is generally typical for the North Arabian Sea (SEN GUPTA and NAQVI, 1984) the intermediate and deeper waters of the entire Gulf of Aden contain almost inexhaustible reserves of phosphate. Throughout the year the whole column from 200-1,000 m generally contains phosphate concentrations above 2.0 $\mu M\ dm^{-3}$ $P\text{-}PO4^{-3}$, reaching maxima of about 2.5 $\mu M\ dm^{-3}$ in the western and 3.0 $\mu M\ dm^{-3}$ in the eastern parts of the Gulf. Even at 100 m concentrations are rarely below 1.0 $\mu M\ dm^{-3}$. The lower part of euphotic layer (seasonally between 30 and 50 m) consistently shows high levels of 0.4 - 0.6 $\mu M\ dm^{-3}$, and very rarely minima of 0.1 - 0.2 $\mu M\ dm^{-3}$ are recorded in inshore waters. Almost the same levels are found in uppermost layer and at the very surface; a few records fall below 0.2 $\mu M\ dm^{-3}$, and in one localized inshore area concentrations fall below 0.05 $\mu M\ dm^{-3}$.

During the summer the upwelled intermediate waters enrich the very surface waters with excessive phosphate concentrations of 1.15 - 2.50 $\mu M\ dm^{-3}$, probably the highest surface concentrations recorded in the tropical oceanic environment. Due to lateral and offshore circulation, these phosphate-rich surface waters seem to spread over the Gulf, hence supplying sufficient quantities for primary productivity. Therefore it seems clear that the phosphorus is never a limiting factor.

Nitrate - Nitrogen

Similarly to phosphate the nitrate concentrations are quite uniform and very high in the whole column, from 200 to 1000 m; values are consistently above 25 $\mu M\ dm^{-3}$ N-NO_3, with deep maxima of 33 $\mu M\ dm^{-3}$ in the western and 39 $\mu M\ dm^{-3}$ in the eastern Gulf. Waters in the lower part of the euphotic layer and down to 100 m, however, show more pronounced oscillations, with an average range of 3 -15 $\mu M\ dm^{-3}$ (except during summer upwellings), indicating a dynamic relationship with the primary production. In the upper euphotic layer nitrate temporal dynamics is much more expressed. Although locally and rarely nitrate may approach depletion, with concentrations below 0.5 $\mu M\ dm^{-3}$, the oscillations in the surface layer, which average 0.6 - 3.0 $\mu M\ dm^{-3}$, do not indicate limitation of a balanced primary production. In blooming conditions such surface nitrate levels of course would be depleted; however, this appears to happen only during the summer when upwelling can supply enormous amounts of nitrate to the surface layers, resulting in surface concentrations of 3 - 6 $\mu M\ dm^{-3}$ in the Gulf generally and well above 20 μMdm^{-3} in upwelling centres (maximum recorded surface concentration is 33.8 $\mu M\ dm^{-3}$!).

Nitrite and Ammonia

Both nitrate and ammonia have been measured regularly from all collected nutrient samples in the Gulf of Aden. However, since these samples have been preserved deep frozen, we consider the results to be only semi-quantitative.

Nitrite concentrations average 0.05 to 0.5 $\mu M\ dm^{-3}$ N-NO_2. Extremely low concentrations below 0.03 $\mu M\ dm^{-3}$ or below the detection limit, are found often (about 30% of measurements), usually (but not always) in the upper euphotic layer. Highest concentrations, in the range of 0.8 - 1.8 $\mu M\ dm^{-3}$, are found during the summer in the whole column of surface layer down to 200 m, restricted to upwelling centres. High concentrations, in the range 0.6 -1.2 $\mu M\ dm^{-3}$, also have been recorded during the rest of the year, with irregular spatial distributions generally associated with the layer just below the euphotic layer (50-100 m). This may correspond to the upper nitrite maximum as described for the Arabian Sea previously (SEN GUPTA and NAQVI, 1984).

Recorded ammonia concentrations average 0.4 - 1.0 $\mu M\ dm^{-3}$ N-NH_4 (65% of all measurements), with the most usual concentrations of about 0.5 $\mu M\ dm^{-3}$. Minima (less than 0.2 $\mu M\ dm^{-3}$ or below the detection limit) are rarely observed. Considering the rather modest concentrations, there is no indication for any regularity of a spatial-temporal distribution of ammonia. However, high ammonia concentrations (extremely high values being 12.0 - 21.5 $\mu M\ dm^{-3}$) are found as a rule in the upper layers (surface - 200 m) on the edge of the shelf during the winter, and in the upwelling centres during the summer. Both cases may indicate active bacterial desemination, or massive animal excretion of ammonia whose concentrations exceed its assimilation by phytoplankton and/or its nitrification.

Silicate

Silicate concentrations throughout the euphotic layer never fall below about 0.5 $\mu M\ dm^{-3}$ Si - Si $(OH)_4$, and usually they are well above 2.0 $\mu M\ dm^{-3}$. Therefore the silicate is not considered a potentially limiting nutrient, even though diatoms predominate phytoplankton communities. However, the spatial and temporal distribution of Si, as shown in Figure 12, provides an additional example of upwelled deep intermediate waters (500 m) into very surface layers. It shows also the "diluting" effect of the inflowing Red Sea waters in the western part of the Gulf, where in the deep layers (800-1,000 m) maximum silicate concentrations are about 47 $\mu M\ dm^{-3}$, while in corresponding depths of the eastern waters they are above 70 $\mu M\ dm^{-3}$.

In conclusion it appears that the Gulf of Aden, considering its nutrient balance, presents a unique oceanic environment where nutrients, namely phosphorus and nitrogen, are never limiting factors of primary production, except perhaps during inshore blooms. Observed oscillations in phytoplankton standing crops seem to originate from physical (circulation, spring overheating of euphotic layers) and biological (grazing) components of the ecosystem.

OXYGEN

The spatial distribution of dissolved oxygen during the major part of the annual cycle (except in upwelling areas during the summer) appears to be quite stable and uniform, yet generally at extraordinarily low levels (e.g., SEN GUPTA and NAQVI, 1984; SWALLOW, 1984; WYRTKI, 1971, 1973). The western part of the Gulf is significantly better oxygenated than the central and eastern waters, due to input of RSW. Even so, the well-saturated layer (O_2 concentrations 4.5 - 5.2 ml l^{-1}) extends no deeper than 50 m. The next deeper layer (to the depths of 200-300 m) shows a gradual decrease of concentrations to 1.0-1.5 ml l^{-1}; at greater depths they fall below 1.0 ml l^{-1}, but never less than 0.5 ml l^{-1}. The deepest measured layers (800-1,000 m) usually show slight increases (to about 1.0 ml l^{-1}) due to inflow of RSW. Central and eastern parts of the Gulf have, as mentioned, lower contents of available oxygen. The well-saturated surface layer (4.5 - 5.2 ml l^{-1}) is usually shallower, about 20 m, although the concentrations greater than 4.0 ml l^{-1} may go down to 100 m during the winter, particularly in the areas of assumed central gyres. The layer in which O_2 gradually drops to the level of 1.0 - 1.5 ml l^{-1} extends to the depths 100-200 m, while concentrations less than 1 ml l^{-1} are usual in deeper waters, often approaching anoxic concentrations (about 0.5 ml l^{-1} and less; minima reaching 0.08 ml l^{-1}).

Sudden changes in oxygenation occur during the summer, with a general decrease of oxygen content in the whole Gulf except in assumed central gyres and along frontal zones. These are probably places where upwelled surface waters sink, saturated with oxygen from blooming phytoplankton. Actually, this seems to be the only major source of the oxygen for the central and eastern waters of the Gulf of Aden, other than Red Sea input in the west. However, drastic changes in oxygenation (or better to say deoxygenation) affects the upwelling areas. The semi-anoxic environment appears in less-pronounced western upwellings (as recorded west of Aden and offshore Berbera) up to 50 m below the surface (Figs. 8 and 10)[2], but in the east such an environment in extensive upwelling centres extends down from 20 m (Figs. 9 and 10). The only oxygenated layer occurs in the upper 10 m due to photosynthesis. This phenomenon can be considered an ecologic catastrophy upon the ecosystem and its communities (including living resources), and it will be discussed in the next chapters.

PELAGIC PRODUCTIVITY

Phytoplankton

Excluding short-term spring minima and upwelling-induced summer maxima, the phytoplankton abundance and biomass (estimated as chlorophyll *a*) are very high throughout the year, averaging 150,000 cells dm^{-3} and 0.5 μM dm^{-3} chlorophyll *a* in the euphotic layer. Chlorophyll values higher than previously reported for the Gulf of Aden (KREY and BABENARD, 1976 etc.) are the highest (in average) for the Indian Ocean and seem disproportionally high compared to phytoplankton abundance. This is partially due to the large size of diatoms which can account for a major part of the chlorophyll content. On the other hand, small-size components (e.g., naked nannoplanktonic flagellates) contribute measured values of chlorophyll, but they are not included in abundance counts since they disintegrate in preserved samples.

A marked decrease in standing stock occurs in spring, with abundances falling below 50,000 cells dm^{-3} in the eastern parts of the Gulf. Even during the summer relatively low abundances (in the range between 50,000 - 100,000 cells dm^{-3}) occur locally inshore and largely offshore. As mentioned previously, these small-amplitude oscillations cannot be explained by nutrient limitations since they hardly exist.

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4 Since true surface upwelling (surface temperatures 17-18°C) also were recorded previously (SERIY, 1968) in western waters, temporary semi-anoxic conditions most likely appear much closer to the surface.

The maximum standing stocks in the Gulf of Aden occur during the summer and post-monsoon autumn, with average abundances above 200,000 cells dm^{-3} and chlorophyll *a* values in the range of 1.0 to 2.0 $\mu M\ dm^{-3}$. However, in upwelling areas, extensive blooms can occur, with abundances of 1-6 millions cells dm^{-3} and chlorophyll *a* values up to 5.1 $\mu M\ dm^{-3}$. To our knowledge such extremes have not been recorded previously in the open Indian Ocean, although they most probably do occur in other upwelling zones in the northern Arabian Sea. Nevertheless, such a massive production of organic matter (a considerable part of which is probably "wasted" by bacterial decomposition) may contribute a great deal not only to the richness of higher trophic levels, but also to the marked oxygen depletions discussed above.

The taxonomic composition of phytoplankton during the blooming phase shows very low diversity, as is typical for "immature" upwelling ecosystems (MARGALEF, 1978). Few species of diatoms(*Chaetoceros curvisetus*, *Ch. compressus*, *Thallasionema nitzschioides*, *Rhizosolenia fragilissima*, *Nitzschia delicatissima*, *Skeletonema costatum* and *Asterionella japonica*) account for 70-80% of the total abundance. In contrast, throughout most of the year the phytoplankton communities have quite diverse compositions; seventy species in a sample is not unusual.

Zooplankton

Judging from the present data base, standing stocks of zooplankton exhibit much less pronounced seasonal oscillations, (Figs. 13 and 14) and more homogeneous spatial distribution, even during summer upwelling (Fig. 16). However, in the eastern upwelling center the maximum standing stocks occur during the period of upwelling and consequent phytoplankton blooms. This is not the case in central and western waters, where zooplankton populations actually fall in the initial period of NE monsoon.

Two smaller peaks in populations occur at the end of winter and during the SW monsoon in summer. A significant minimum is evident in spring, which hydrographically is characterized by overheating (up to 32°C in the mixed layer).

The average range of zooplankton standing stocks throughout the year, expressed as displacement volume per haul (to a depth of 30 m), is 5-10 cm^3, with minima of 2 to 3 cm^3 and maxima of 14-24 cm^3. Although still preliminary, dry weights and ash-free dry weights for zooplankton biomass range from 52 to 301 mg m^{-3} and 23-111 mg m^{-3}, respectively. These values support previous statements which consider zooplankton stocks of the NW Arabian Sea as the richest in the Indian Ocean (GAPISHKO, 1971; LENZ, 1973; RAO, 1973). Moreover, our values exceed previously reported ones and indicate the Gulf of Aden (in terms of zooplankton standing stocks) to be one of the richest areas of the world ocean.

Although it is rather premature to report on the composition of studied zooplankton communities, some currently available information seems worthy of mention. Average abundances, both during the winter and the summer periods range from 1,000 to 6,000 specimens m^{-3}, the maxima being 8,000-16,500 specimens m^{-3}. The most common taxonomic groups are as follows:

Taxon	% Relative Abundance
Cnidaria	0.1 - 3.5
Pteropoda	0.1 - 2.6
Appendicularia	0.5 - 18.3
Chaetognatha	0.5 - 8.1
Thaliacea	0.2 - 4.3
Ostracoda	0.1 - 2.5
Copepoda	65.5 - 96.8

EXPLOITABLE LIVING RESOURCES

The first reported survey of the fish stocks in the Gulf of Aden was given in FAO (1973), based on the work of A. Druzhinin. Surveys were conducted in 1975/76 on the R/V Dr. F. Nansen, as reported in FAO/NORWAY/UNDP (1978), INSTITUTE OF MARINE RESOURCES BERGEN (1975-77), KESTEVEN et al. (1981). A stock assessment of cuttlefish in PDR Yemen waters was made by SANDERS (1981), while stocks of oil sardines on this coast were assessed by SANDERS and BOUHLEL (1984). The results presented here were obtained from the R/V Ibin Magid, and also by participation in the 1984 surveys of R/V Dr. F. Nansen.

DEMERSAL FISH STOCKS

Seasonal Changes

West of Aden, the dominant families in the October through March period are Leiognathidae (pony fish), Carangidae (jacks), Pomadasyidae (sweet lips), Lethrinidae (emperors) and Nemipteridae (threadfin breams). In the central sector, just east of Mukalla, dominant "fish" are *Sepia* sp. (cuttlefish), Sparidae (sea breams), Dasyatidae (rays), Pomadasyidae and Synodontidae (lizard fish). In the east, on the main trawl ground west of Ras Fartak, dominant stocks are *Sepia* sp., Ariidae (catfish), Sparidae, Dasyatidae and Sphyraenidae (barracudas).

Generally, the differences between the Aden area and localities east of Mukalla are more marked than differences between localities on the eastern part of the coast. Trawl catches in the east are influenced by the large resident stocks of cuttlefish, which do not extend significantly into Omani waters (SANDERS, 1981). During upwelling in August 1984, the catches in the west sector were dominated by Centrolophidae (Indian ruffs), Sparidae, Synodontidae, *Sepia* sp. and *Loligo* sp. (squids). The first taxon usually occurred in deeper waters, while the Sparidae consisted of species normally found below 100 m. In the central sector at Mukalla the dominant taxon was Callionymidae (dragonets), with small amounts of *Sepia*, Triglidae (gurnards) and Sparidae. The first was not encountered in October through March and was presumed to have migrated from greater depths. The Triglidae also were deeper water fish. In the east sector Callionymidae again dominated, followed by Triglidae, Nemipteridae, Clupeidae (sardines) and Synodontidae.

These data indicate that during upwelling the normal fish stocks migrate to shallower depths, to be replaced by deeper water groups. Some typical taxa were found close inshore, while others were taken in pelagic trawls near the surface. The occurrence of these truly demersal types, particularly species lacking swim bladders, at the surface (e.g. gurnards and cuttlefish) suggest that vertical as well as inshore migrations had occurred. To the east of Ras Fartak a few large catches of Nemipteridae and cuttlefish were made at 17 m depth, giving the impression that these had migrated from trawl grounds on the west side.

Generally speaking, trawl catches were highest around Ras Fartak in the non-upwelling period, lowest around Mukalla, and intermediate at Aden. During upwelling a large variation in CPUE (catch per unit effort) along the coast was between 0 and 3,058 kg/hour, giving the impression of pockets of low and high density stocks.

Depth Changes

Trawl catch composition and CPUE changed significantly with depth in a seasonal pattern. October through March at Ras Fartak, a large component of *Sepia* was found at 15 and 30 m, due to migration of mature adults into shallow water for spawning. Here the Ariidae (catfish) were an important component of catches, with Sparidae being abundant at 30 m. At 50 m depth, Carcharhinidae (small sharks) dominated, with Carangidae and Pomadasyidae as the other main commerical fish. Below this, at 70 m, portunid crabs dominated, with Nemipteridae and Carangidae as the commercial fish. At this time of the year the CPUE was highest around 30 m due to the cuttlefish.

By April/May the demersal stocks had migrated to greater depths. The main concentration was now between 100 and 120 m, where Sparidae and cuttlefish were dominant. It is assumed that the crabs had moved deeper although a few were found. At 30 m the CPUE was reduced to 40 kg hr^{-1} compared with 381 kg hr^{-1} in October. The dominant families at 30 m were Sparidae, *Sepia* and Carangidae, with a much reduced speciation compared to October. At 114 m depth the CPUE was the highest recorded, 660 kg hr^{-1}, where the main taxa were *Sepia*, Sparidae and Serranidae (groupers).

During upwelling at the same locality at Ras Fartak, catches deeper than 20 m depth were dominated by Callionymidae (97%), while in shallower water Nemipteridae dominated (90%). Since the latter normally occurred between 70 and 100 m at this locality, a strong inshore migration was indicated.

Locality Changes

Local changes have been indicated for the west, central and east sectors in Figure 17. Variations along the coast in the non-upwelling period were too complex to discuss here, but some groups were only found at Ras Fartak (e.g. Drepanidae, moonfish), while others reached their maximum abundances here and were scarce elsewhere (e.g. Trichiuridae, Indian threadfins). In other groups the speciation was different along the coast (e.g. Carangidae).

Standing Stocks

Estimates of the quantities of demersal fish on the shelf were made using the swept area method of SHINDO (1972) with a catchability coefficient of 0.5, or an assumed escapage of fish both through the trawl meshes and avoidance of the net of 50%. During upwelling demersal trawl data may have been misleading due to intrusion of deeper water species on to the coastal shelf. Further, the restricted survey of R/V Dr. F. Nansen in August may have missed pockets of inshore demersal fish. The average standing stock estimates for depths to 120 m are given in Table 2, while the seasonal variations (with depth) in standing stocks at Ras Fartak are given in Table 3, where movement by season can be seen. Higher stocks occurred on the eastern trawl grounds, which were the main location of cuttlefish stocks. The shelf area of this coast is around 26 x 10^3 km^2 to a depth of 200 m. R/V Dr. F. Nansen surveys in 1975/76 gave a demersal stock estimate of 164 - 238 x 10^3 tons using integrated echosounding. The results of the R/V Ibin Magid surveys have indicated a minimum stock of 116 x 10^3 tons, partly using data from R/V Dr. F. Nansen to calculate stocks between 120 and 200 m.

A decrease in stocks, probably due to trawling for cuttlefish, has occurred since 1975. Formerly cuttlefish were trawled over a wider area, with grounds existing to the west of Aden, but are now trawled between Mukalla and Ras Fartak. The stock estimates given here for the east sector in Table 2 coincide with those in FAO (1981) for neighbouring Omani waters.

Other Demersal Resources

In shallow coastal areas with muddy bottoms, the shrimp *Penaeus semisulcatus* was found. East of Ras Fartak catch rates reached 17 kg hr^{-1} during spawning in October. The deep sea lobster, *Puerulus sewelli*, is trawled at 200 to 300 m depth off Mukalla and Ras Fartak.

Table 2. Average stock estimates of demersal fish for the coastal shelf to a depth of 120 m (tons km^{-2}).

Area/Season	West	Central	East
Non Upwelling	3.6	3.2	6.0
Upwelling	1.9	1.4	2.5

Table 3. Seasonal variations in standing stocks of demersal fish at Ras Fartak with depth (tons km^{-2}).

Depth in m.	October	April	August
10-20	1.3	0.6	8.2
30-50	6.0	0.8	2.1
80-100	1.2	0.9	1.0
100-120	0.9	10.3	0.6

PELAGIC FISH STOCKS

Epipelagic Stocks

These have mainly been assessed with integrated echo-sounding from R/V Dr. F. Nansen (STROMME, 1984) with an assessment of the stocks of oil sardine, *Sardinella longiceps*, given by SANDERS and BOUHLEL (1984). Figure 18 shows the distributions of the stocks in March and August, 1984.

In March concentrations of pelagic fish were found to the west and east of Aden, west and east of Mukalla and around Ras Fartak. The oil sardine was identified around Aden and at Mukalla, extending east to Ras Fartak. The scads, *Trachurus indicus* and *Decapterus russelli*, were found at Aden and Ras Fartak. A further concentration of sardines to the east of Ras Fartak, suggested in ANON (1981), was located in August. During the August upwelling, the pelagic stocks appeared to have migrated offshore along the eastern part of the coast, although the ship did not go close to shore. A large concentration was found to the south of Mukalla, with another one south of Ras Fartak. The former contained *Sardinella longiceps* and *Scomber japonicus* (chub mackerel), while the latter was mainly chub mackerel. Around Ras Fartak were scads, *Trachurus indicus*.

The fish in the offshore area south of Mukalla had ripe gonads and were in spawning condition. A stock estimate given by STROMME (1984) was 290 x 10^3 tons for the entire coast, but the area west of Aden was not surveyed. The previous estimate was 376-495 x 10^3 tons for 1975/76.

A stock of pelagic fish not covered by these surveys was the Indian mackerel, *Rastrelliger kanagurta*, which occur around Aden and supports a large seasonal fishery. Further, the considerable stocks of Spanish mackerel, *Scomberomorus commerson*, which are fished between Aden and Mukalla, probably were underestimated since they avoid ships. Resident in the western waters are stocks of bonito tuna, *Euthynnus affinis*, while in the east are mainly skipjack, *Katsuwanus pelamis*. In about April through June the larger yellowfin tuna, *Thunnus albacares*, appears off Mukalla and supports a fishery. SANDERS and BOUHLEL (1984) calculated the stocks of *Sardinella longiceps* between Mukalla and Ras Fartak as 32 x 10^3 tons after upwelling in October.

From these various estimates of pelagic fish stocks it is likely that the estimate of STROMME (1984) may be more applicable to the total stocks at the present time, which may be of the order of 300 x 10^3 tons. However, sufficient data to divide this stock into constituent species are not available.

Mesopelagic Stocks

These stocks, composed mainly of myctophid fish (with *Benthosema pterotum* as dominant species), have been studied from the R/V Dr. F. Nansen and reported by SANDERS and BOUHLEL (1982) and GJOSAETER (1984). These fish are of small size but have a fast growth, reaching 4 cm in 6 months when spawning occurs. They occur where the depth exceeds 150 m, which in the present context would be the shelf edge. Vertical migrations occur between 150 to 200 m depth in the day rising to within 10 m of the surface at night.

Stock densities in the Gulf of Aden were estimated at above 100 tons km^{-2} (GJOSAETER, 1984). Based on growth and recruitment studies, SANDERS and BOUHLEL (1982), estimated the annual yield from the Gulf of Aden at about 15×10^6 tons.

GENERAL CHARACTERISTICS OF EXPLOITABLE RESOURCES

Demersal fish populations vary in composition and standing stocks with season, depth and locality on the coast. In the west around Aden the dominant taxa in the non-upwelling season are Leiognathidae, Carangidae, Pomadasyidae, Lethrinidae and Nemipteridae. During upwelling Centrolophidae, Sparidae, Synodontidae, *Sepia* sp. and *Loligo* spp. are prominent. In the east around Ras Fartak the dominant non-upwelling taxa are *Sepia*, Ariidae, Sparidae, Dasyatidae and Sphyraenidae, while during upwelling Callionymidae, Triglidae, Nemipteridae, Clupeidae and Synodontidae predominate.

Standing stocks at 30 m depth have been measured to be 3.6 and 6.0 tons km^{-2} in the west and east respectively during non-upwelling, falling to 1.9 and 2.5 tons km^{-2} during upwelling. Maximum standing stocks occur at 30 m depth in October, at 110 m in May and 20 m or less in August.

During upwelling demersal populations migrate both inshore and vertically in the water, and appear to separate into pockets along the coast. Deeper water species move on to the shelf and into shallow water at this time. These migrations probably occur in response to oxygen minima.

Pelagic fish stocks are concentrated along the coast in the non-upwelling season at Aden, Mukalla and around Ras Fartak. During upwelling part of the stocks in the east migrate offshore, while those at Aden remain inshore. Pelagic trawls have showed Clupeidae (i.e. *Sardinella longiceps* , the oil sardine) to be dominant in the non-upwelling period.

Large pelagic fish (e.g. tunas, Spanish mackerels) have not yet been assessed, but concentrations appear mainly in the western two thirds of the Gulf, with highest stocks in the Aden region. Here substantial stocks of Indian mackerel, *Rastrelliger kanagurta*, are found.

Total demersal stocks may exceed 116×10^3 tons, while pelagic stocks may be around 300×10^3 tons.

SUMMARY

This region of the Indian Ocean presents a rather unique, large-size ecosystem of which the most remarkable characteristics are as follows:

- Oceanographic conditions that reverse in response to the changes between NE and SW monsoons also respond to variable and poorly understood influences of the Somali Current and inputs from the Red Sea and the North Arabian Sea intermediate - deep water masses.
- Upwelling occurs regularly during the SW monsoon season in many areas of this region. However, the most affected areas are in the east where the upwelling is much greater than in "classic" upwelling regions of the world oceans.
- Because of the above conditions, the supply of nutrients to euphotic layers is sufficient throughout the annual cycle to support a balanced primary production. During the summer and the subsequent postmonsoon season, however, the production dramatically increases, reaching densities up to 6 million cells dm^{-3} and 5 μg dm^{-3} of chlorophyll *a*. Secondary production of zooplankton is also very high, with average dry weight biomass of about 100 mg m^{-3}. This suggests that these waters are among the richest not only within the Indian Ocean but also in the world oceans.

- It is logical to expect that such high productivity supports very large stocks of exploitable resources. This expectation is true for the largely underexploited clupeid, carangid and scombrid fishes and an immense potential (est. 15 million tons) of mesopelagic myctophids. However, stocks of demersal fish (which are at present the target of commercial fisheries) appear to be limited and also heavily exploited.
- Unfortunately the overall high pelagic productivity in an "immature" upwelling ecosystem indirectly contributes to the disequilibrium of its oxygen balance, as a large part of produced organic matter is "wasted" by bacterial decomposition. In the case of the Gulf of Aden this is combined with the general deficit of the oxygen in the North Arabian Sea. Consequently, in the largest part of the Gulf the semi-anoxic conditions (0.5 ml l^{-1} or less of dissolved oxygen) appear throughout the upper water column and below 150 - 200 m depth. During the summer this anoxia is generally pushed into the surface layers, and can be extreme in upwelling centres where semi-anoxic conditions prevail below a depth of 20 m or even less.
- Anoxia, which can be considered as an ecological catastrophy, affects the ecosystem in many poorly understood ways. It certainly contributes to an extraordinarily low standing stock of the benthic communities on the shelf bottoms, and consequently to the missing or very weak relationship of the benthos with the demersal fish populations. This in turn also means the trophic dependence of demersal fish on pelagic food potentials that leads to greater degree of overall instability. The appearance of anoxia in shelf waters also causes large migrations and structural recombinations in demersal fish communities, and most likely the migrations of epipelagic fish also. Although anoxia-induced fish mass-mortalities have not been observed in the Gulf of Aden, "hidden" mortalities are assumed to occur.

REGIONAL COOPERATIVE INVESTIGATIONS OF THE NW ARABIAN SEA

The unique oceanographic and marine biological phenomena in the Gulf of Aden and adjacent parts of the Arabian Sea are still poorly documented, and therefore deserve greater attention. Moreover, the crucial economic and nutrition needs of bordering nations the Gulf of Aden provide more practical reasons for increased research. The development of a rational and economically healthy exploitation of the living resources in this region, however, cannot work properly without an adequate scientific basis. To achieve such a scientific basis requires a rather complex knowledge on the ecosystem and living resources (Fig. 19). Using BEHRMAN's (1981) words, "finding fish is not enough"; it leads to the tasks as shown in the lower part of Figure 19.

Since the Gulf of Aden and adjacent parts of the Arabian Sea inevitably must be considered as a functional unit, at least from the standpoint of bioproductivity and oceanographic processes, investigations must cover the whole region. However, the area is extensive and the seas, particularly during the summer monsoon, are heavy. Therefore the sea-going research requires a sizeable well-equipped research vessel, operating continuously year around with an interdisciplinary scientific team aboard, covering the area as shown in Figure 20. Probably no single country bordering this area (Djibouti, Somalia, P.D.R. Yemen, A.R. Yemen and Oman) alone can support the costs of such a study, and even so, this would lead to an uneconomic duplication of efforts. Therefore it is proposed herewith to set up a regionally operating organization, shared by all bordering countries and supported by an UN interagency group (UNDP/FAO/UNESCO-IOC/UNEP). A tentative outline of scientific programme of interdisciplinary oceanographic and fisheries investigations for this area is proposed in the Appendix.

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APPENDIX:
REGIONAL COOPERATIVE INVESTIGATIONS OF THE NW ARABIAN SEA
Outlined Proposal of a Tentative Scientific Programme

1. Objectives
   - A. Fundamental study of the oceanography and bioproductivity.
   - B. Better understanding of monsoon weather generally and sea-air relationships in particular.
   - C. Assessment of pelagic fish stocks and their spatial-temporal variations in structure and distributions as related to both oceanic and ecosystem dynamics.
   - D. Providing information needed for planning, implementation, strategy and technique of an intensified fisheries and for the protection of pelagic resources.
2. Area and Disposition of Sea-going Investigations
   The proposed area is shown in Figure 20, with a logical southward extension, reaching at least 8°N. It shows also: needed stations for continuous meteorologic recording (e.g., automatic weather stations); network of stations for long-range hydrographic and current records; tentative tracks of seasonal (winter, premonsoon, SW monsoon and postmonsoon seasons); and monthly cruises for oceanographic and bioproductivity investigations. The operations needed for the assessment of stocks and fishery biology studies of pelagic fish (integrated echo-sounding, sampling and exploratory fishing) are envisaged to go along with oceanographic cruises, covering the areas of outlined transects and (between them) in coastal waters.
3. Topics of Scientific Programme
   - A. Meteorological records, in particular wind; air temperature and evaporation; solar radiation.
   - B. Hydrography and water masses.
   - C. General circulation.
   - D. Physical processes in upwelling phenomena, including: wind stress, sea-level variations, surface sea water temperature monitoring; (including utilisation of satellite IR images), dynamics of upwelling and adjacent sinking systems.
   - E. Physics chemical factors of primary productivity: nutrients (inorganic and organic sources), CO_2 system.
   - F. Primary productivity as phytoplankton standing stocks and their structure and functional productivity.
   - G. Zooplankton standing stocks and their structure.
   - H. Total orgtanic matter as related to its decomposition.
   - I. Development of anoxic conditions and recycling of nutrients.
   - J. Oxygenation and anoxia.
   - K. Pelagic fish investigations:
   - L. Assessment of stocks, their distribution and migrations.
   - M. Ichthyoplanktological investigations as indicating both stocks and their dynamics.
   - N. Research on biology, behaviour and population dynamics of dominant and/or commercially important species.
   - O. Specific studies on: mesopelagic fish (and invertebrate) populations, large epipelagic fish, pelagic cephalopods.
   - P. Exploratory fishing operations for potentially commercial species.

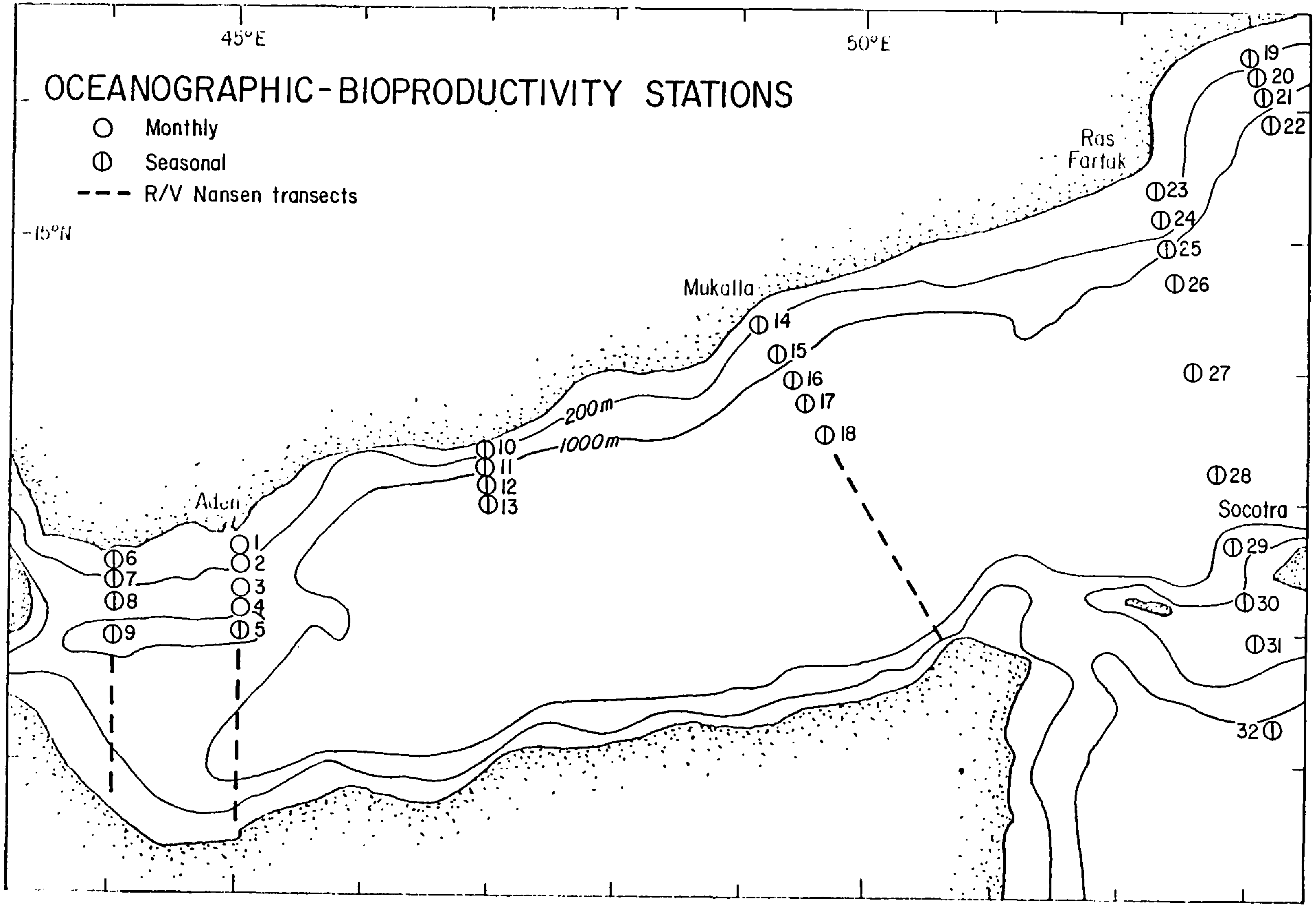


Figure 1. Map of the Gulf of Aden showing oceanographic bioproductivity stations during 1984-85.

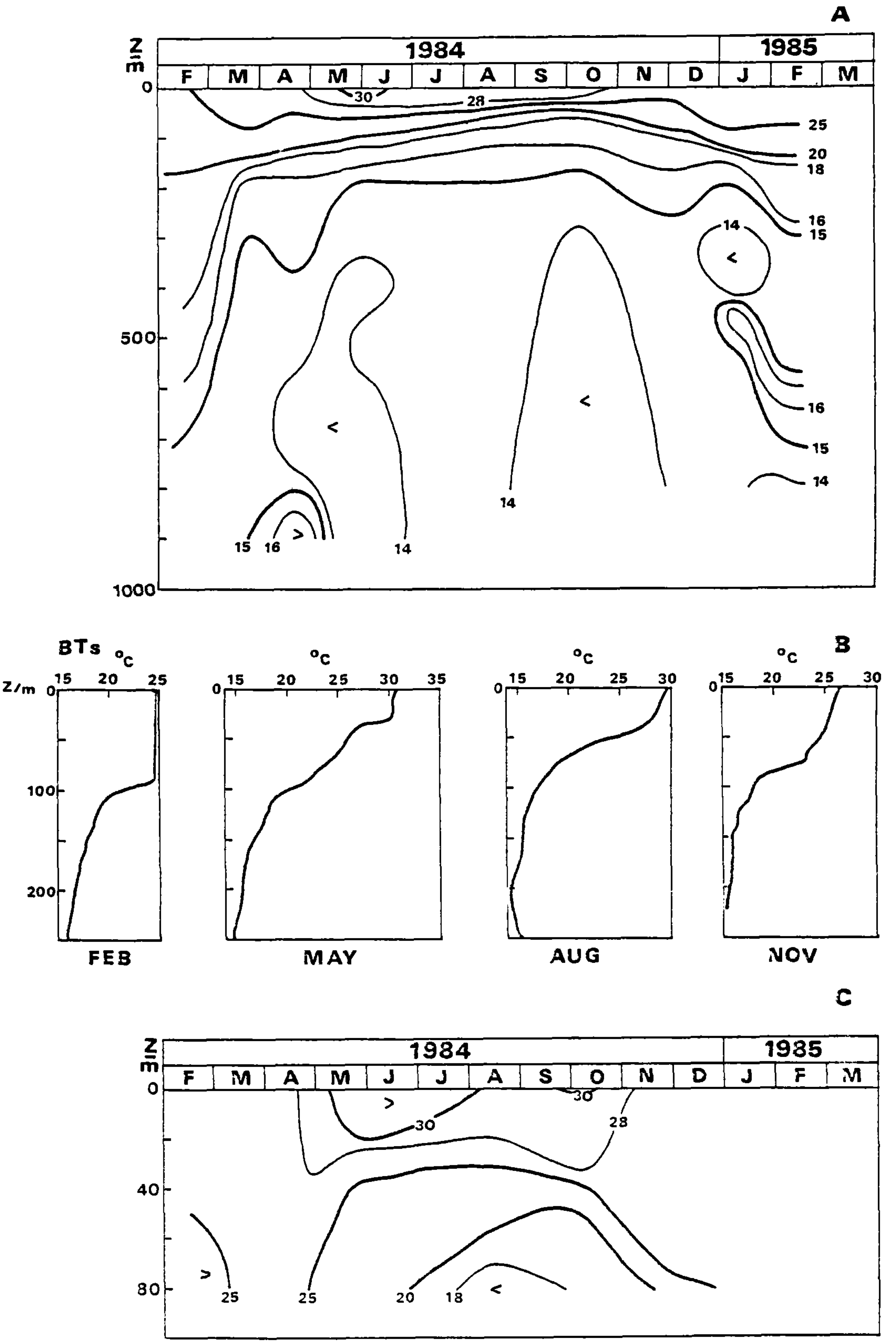
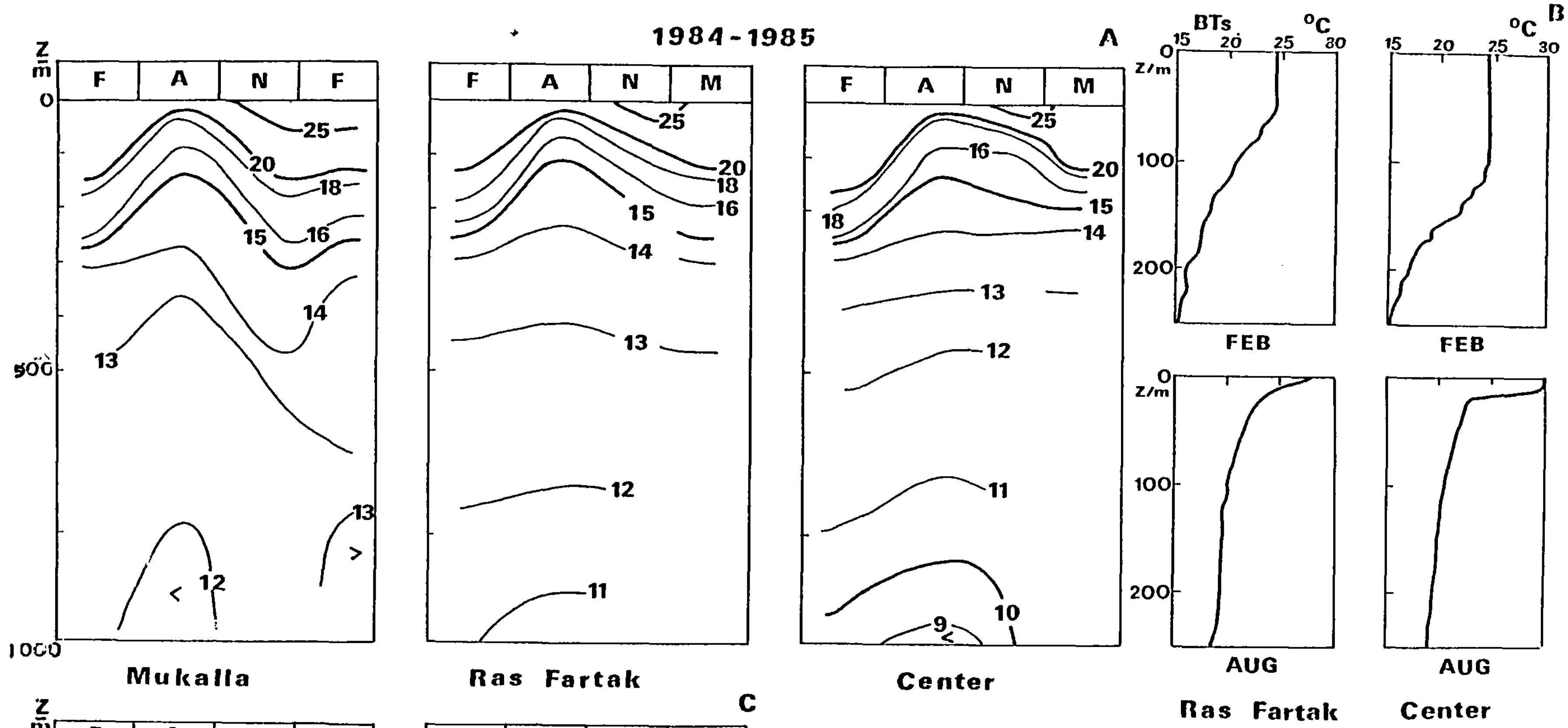


Figure 2. Temporal-spatial variations of sea-water temperature during 1984-85 in the northwestern waters of the Gulf of Aden: A) offshore station No. 4 of the monthly transect 45°E. B) corresponding typical bathythermograms. C) inshore station no. 1 of the transect.



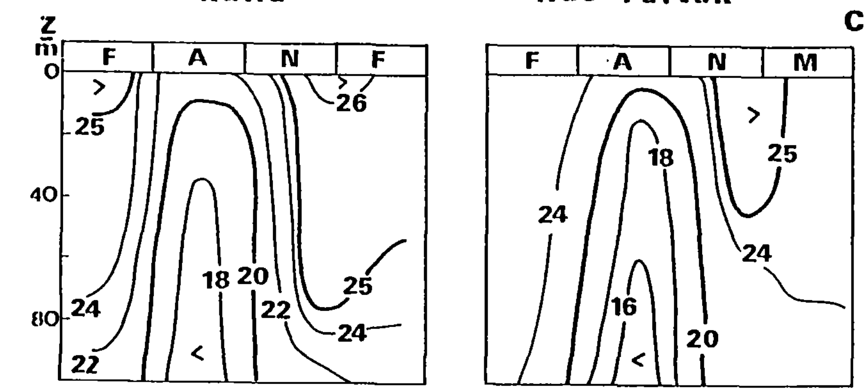


Figure 3. Temporal-spatial variations of sea-water temperature during 1984-85 in the north-eastern waters of the Gulf of Aden at the selected stations: A & C) offshore (No. 15) and inshore (No. 14) at Mukalla, offshore (No. 25) and inshore (No. 24) at Ras Fartak, and central station (No. 27) in the mouth of the Gulf (transect Ras Fartak-Socotra). B) typical bathythermograms.

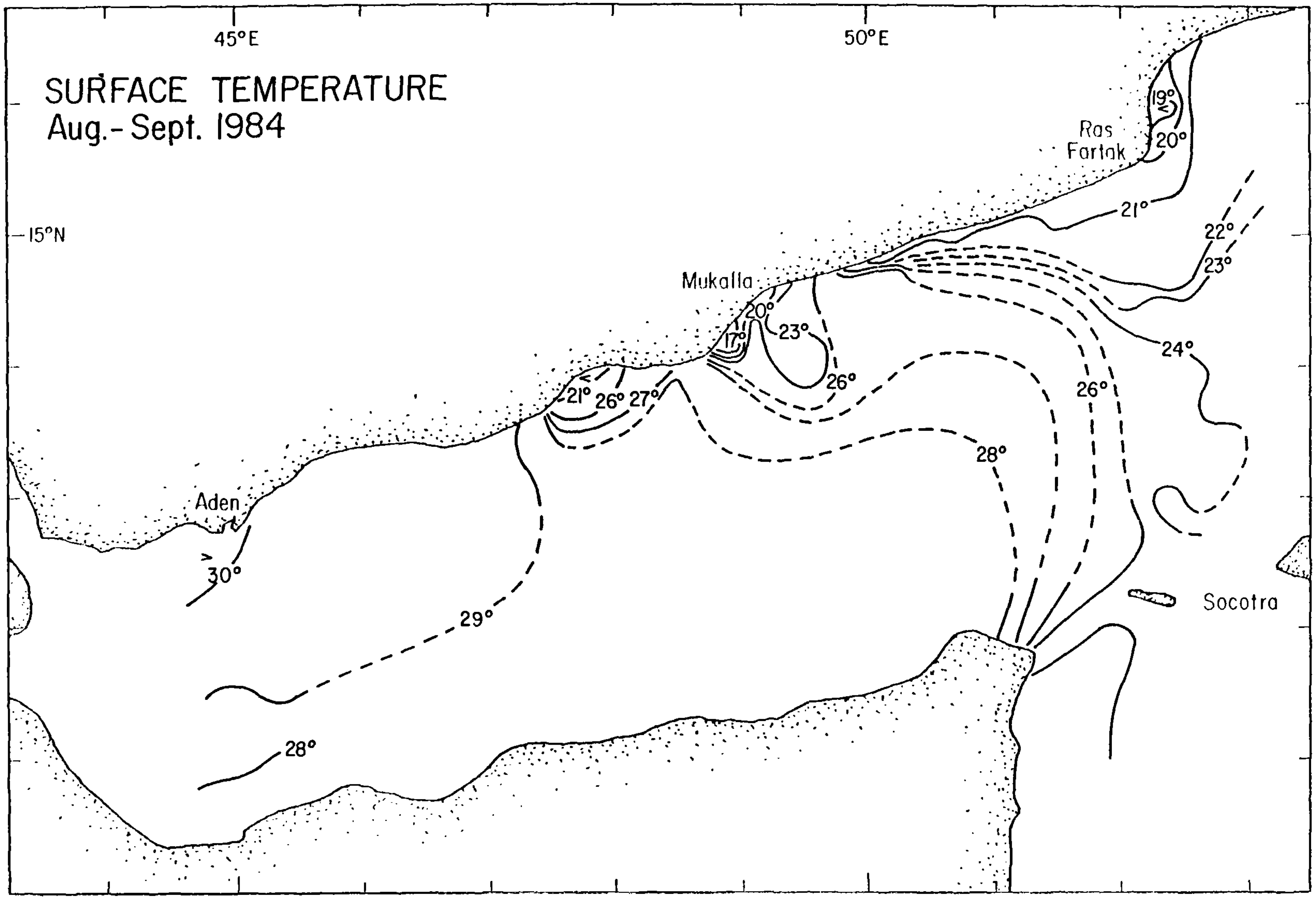


Figure 4. Distribution of subsurface (4 M) sea-water temperature in the Gulf of Aden as recorded in August 1984 (intake sea-water temperature sensor) on R/V Dr. F. Nansen, showing surface expressions of upwelling centres.

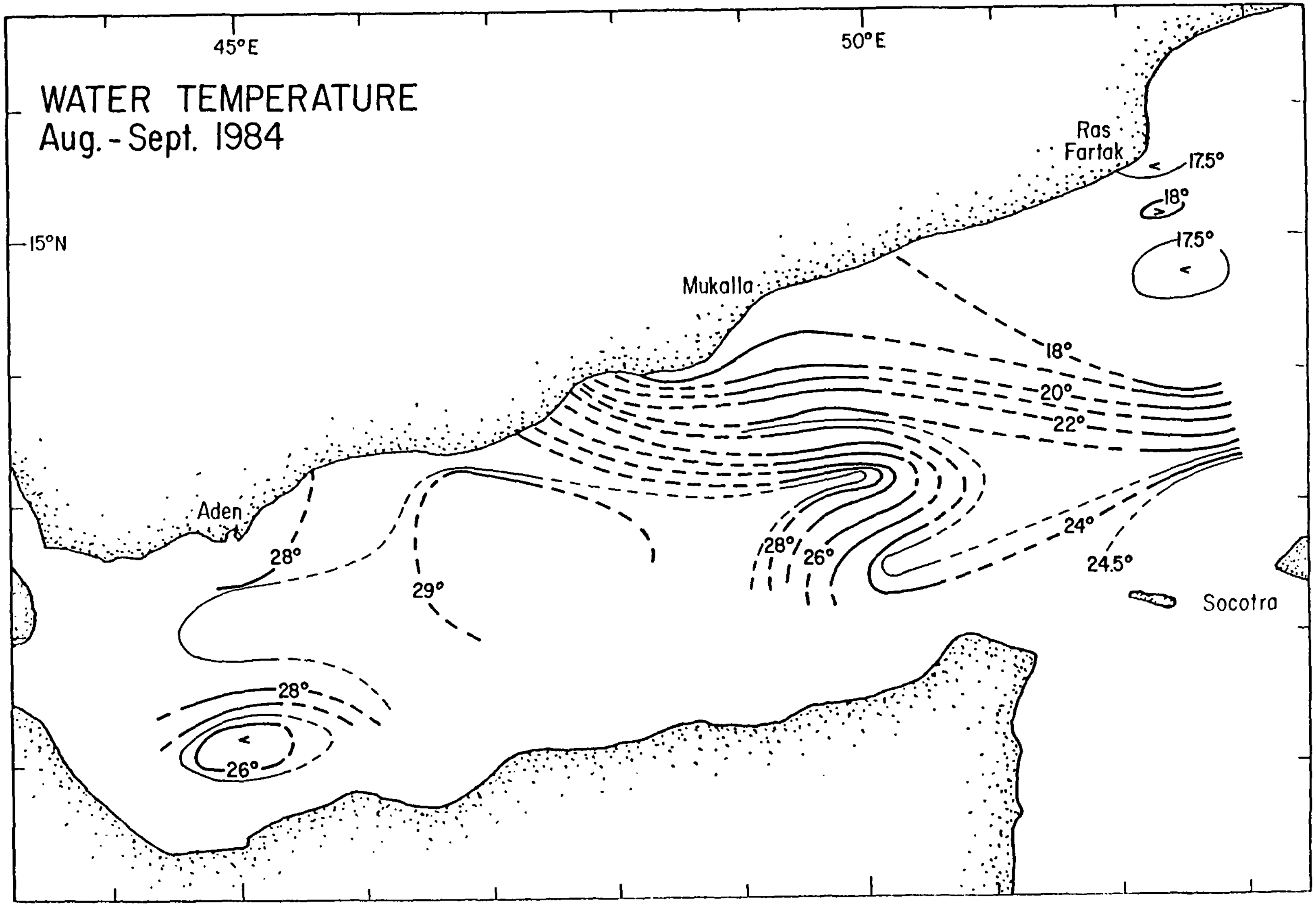


Figure 5. Distribution of sea-water temperature in the Gulf of Aden at 20 m in August 1984, showing extensive northeastern upwelling zone and two residual centres of previous upwelling in the western Gulf.

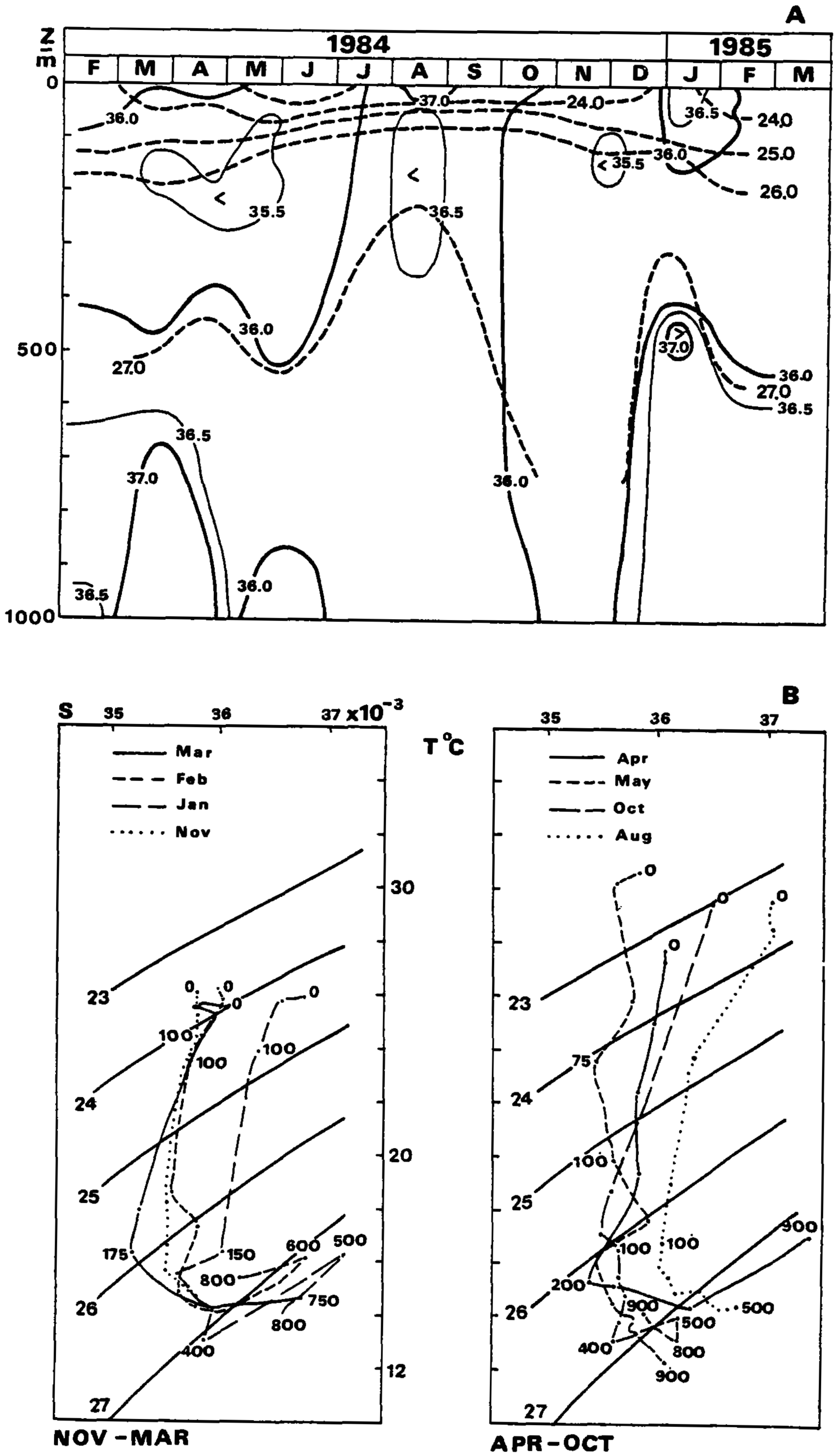


Figure 6. Temporal-spatial variations of salinity and density during 1984-85 in the northwestern Gulf of Aden (Station 4 of the monthly transect 45°E) (A); corresponding T-S diagrams for winter and summer periods (B).

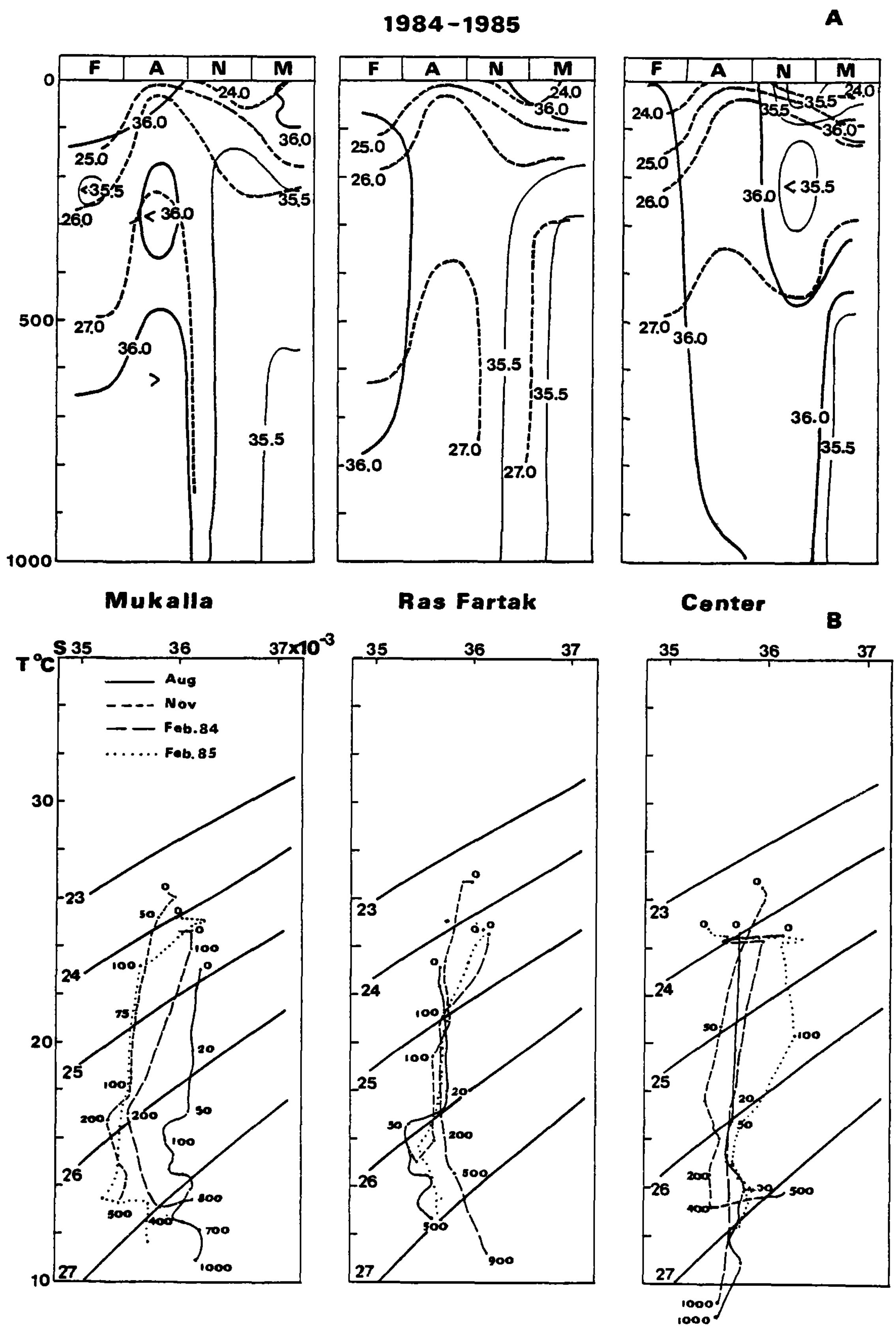


Figure 7. Temporal-spatial variations of salinity and density during 1984-85 in the eastern Gulf of Aden at selected offshore stations Mukalla (No. 15), Ras Fartak (No. 25) and the center of the Gulf's mouth (No. 27: transect Ras Fartak-Socotra), and corresponding T-S diagrams.

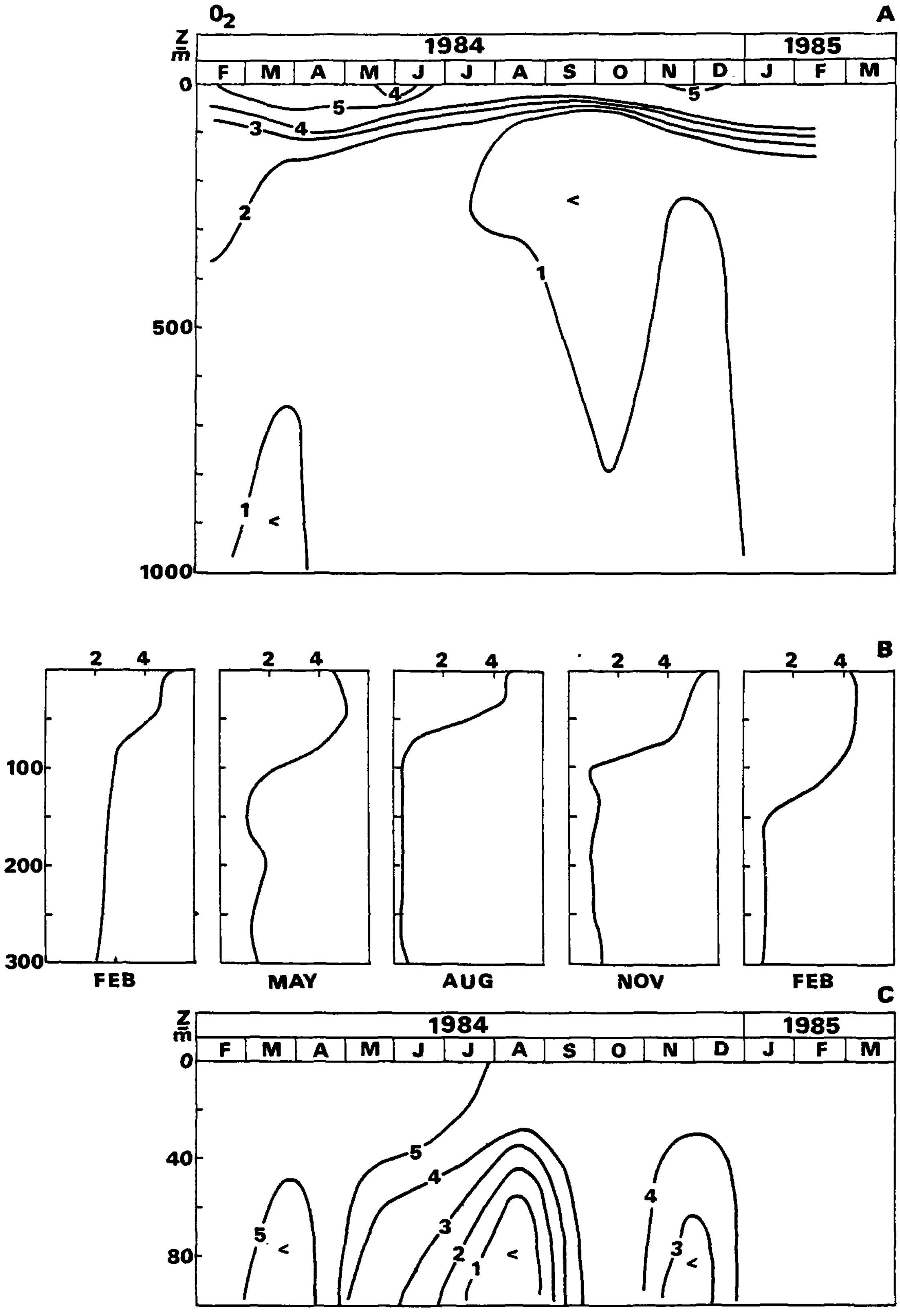
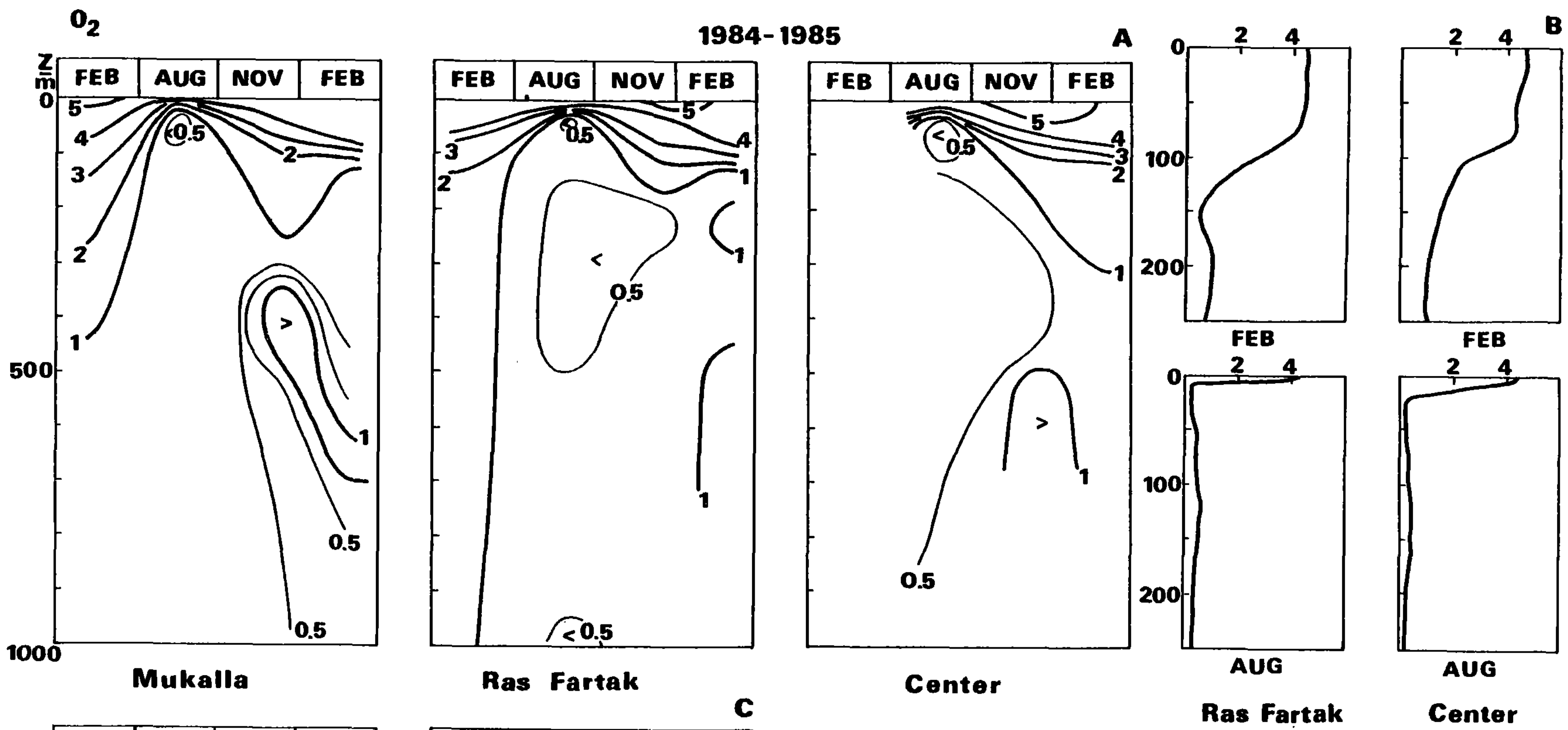


Figure 8. Temporal-spatial variations of dissolved oxygen concentrations (in ml dm^{-3}) during 1984-85 in the northwestern Gulf of Aden: A) offshore station No. 4 of the monthly transect 45°E; B) corresponding typical vertical distributions; C) inshore station No. 1 of the transect as above.



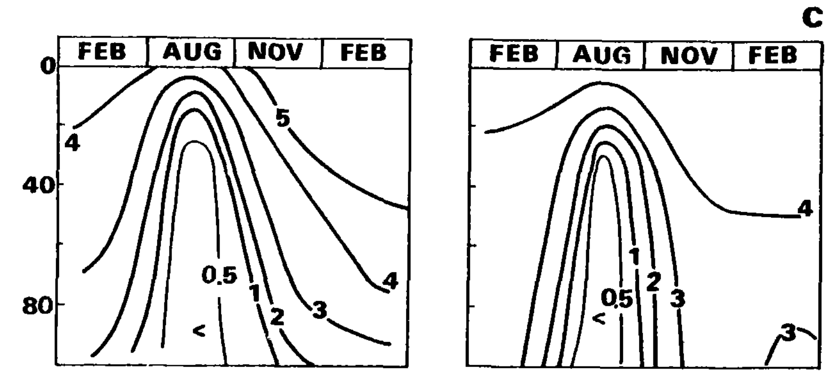


Figure 9. Temporal-spatial variations of dissolved oxygen concentrations (in ml dm^{-3}) in the northeastern Gulf of Aden during 1984-85; A & C) offshore (No. 15) and inshore (No. 14) at Mukalla, offshore (No. 25) and inshore (No. 24) at Ras Fartak, and central station (No. 27) in the mouth of the Gulf (transect Fartak-Socotra); B) typical vertical distributions of O_2

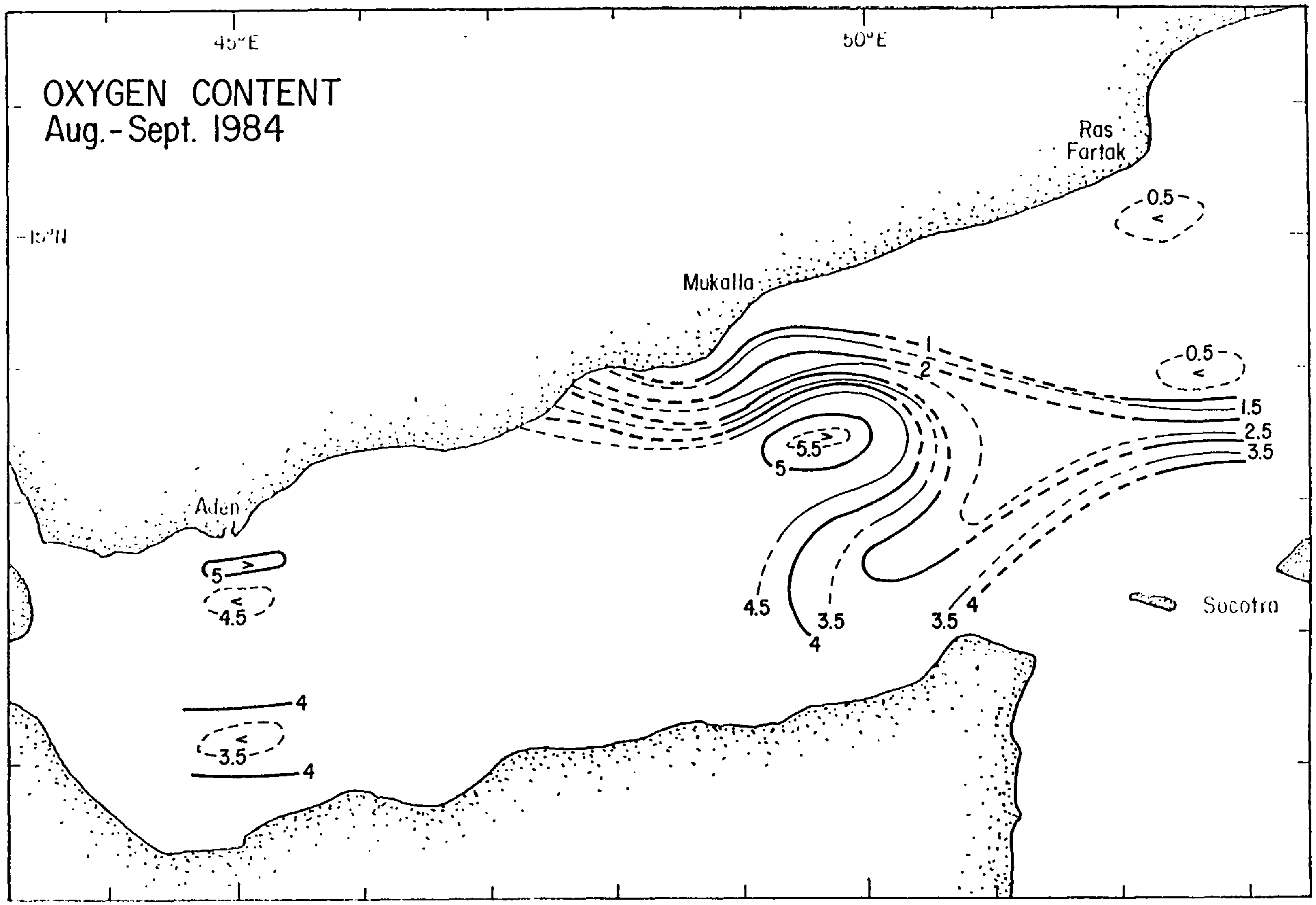


Figure 10. Distribution of dissolved oxygen in the Gulf of Aden at 20 m in August 1984, showing eastern semi-anoxic zones, two O_2 lows from previous upwelling in the west, and a central O_2 high, probably associated with a compensation gyre.

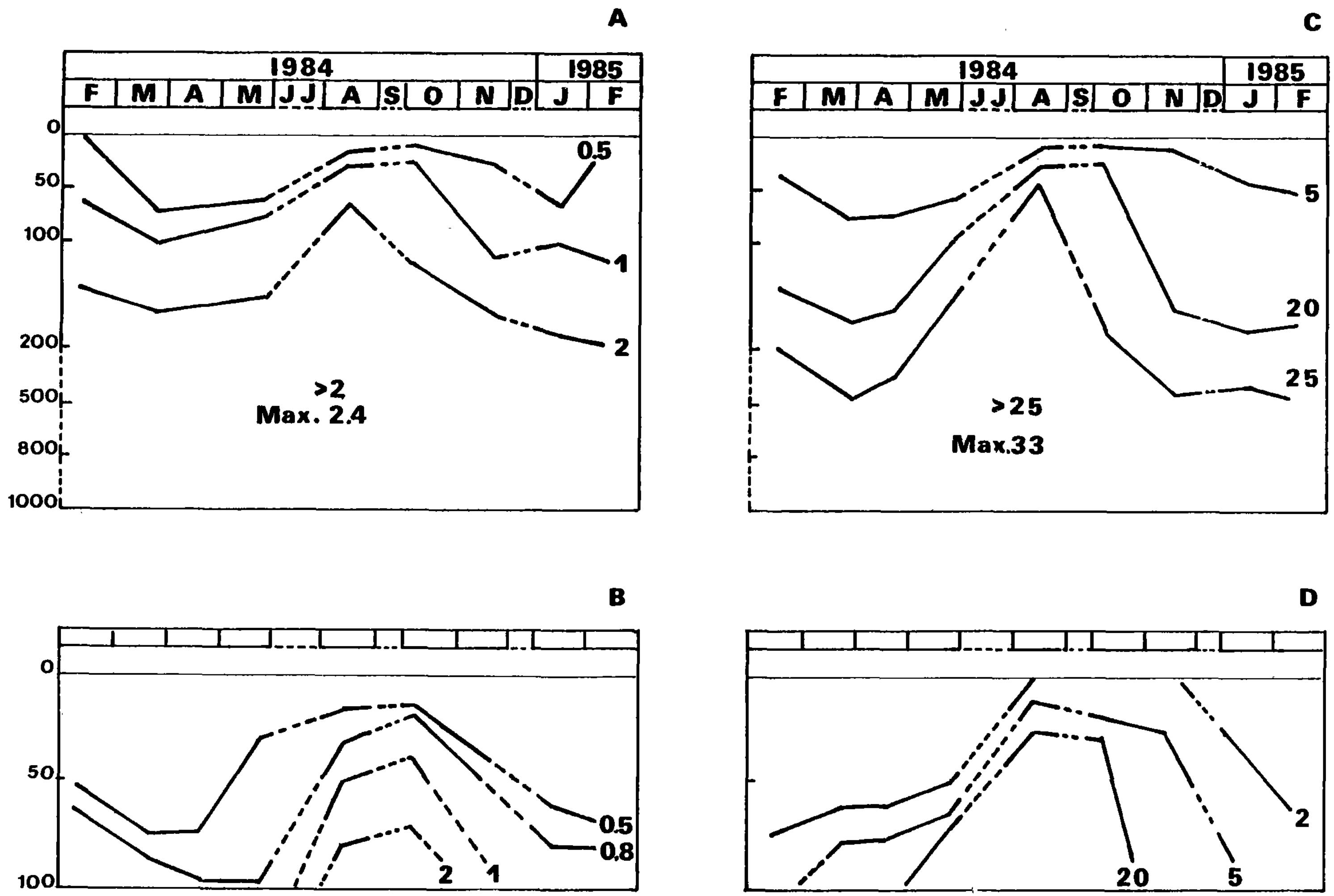


Figure 11. Temporal-spatial variations of phosphate: A & B ($P - PO_4^{-3}$) and nitrate; C & D ($N - NO_3^-$) concentrations (in µM dm^{-3}) during 1984-85 in the northwestern Gulf of Aden; A & C offshore station (No. 4) and B & D inshore station (No. 1) of the monthly transect 45°E.

A

STAT.	Z/m	Si-Si(OH)4 in μM·dm⁻³			
		Months 1984-1985			
		FEB	AUG	NOV	FEB
OFFSHORE MUKALLA	0	2.55	11.70	4.77	5.93
	50	6.80	23.66	7.95	7.42
	100	15.30	22.36	24.91	10.60
	200	—	27.51	—	16.96
	500	47.05	41.18	33.39	24.86
	800	45.80	43.26	37.63	40.39
OFFSHORE RAS FARTAK	0	1.90	27.30	0.79	0.42
	50	2.90	42.20	1.69	0.53
	100	8.65	—	5.94	11.24
	200	18.00	46.73	—	24.80
	500	—	33.80	18.02	20.67
	1000	49.90	77.24	—	—
GULF MOUTH CENTER	0	4.90	41.86	2.81	9.12
	50	6.40	37.44	10.60	12.08
	100	7.65	—	19.61	10.33
	200	—	51.48	23.32	—
	500	16.40	46.18	32.33	38.80
	1000	52.40	72.73	—	49.27

B

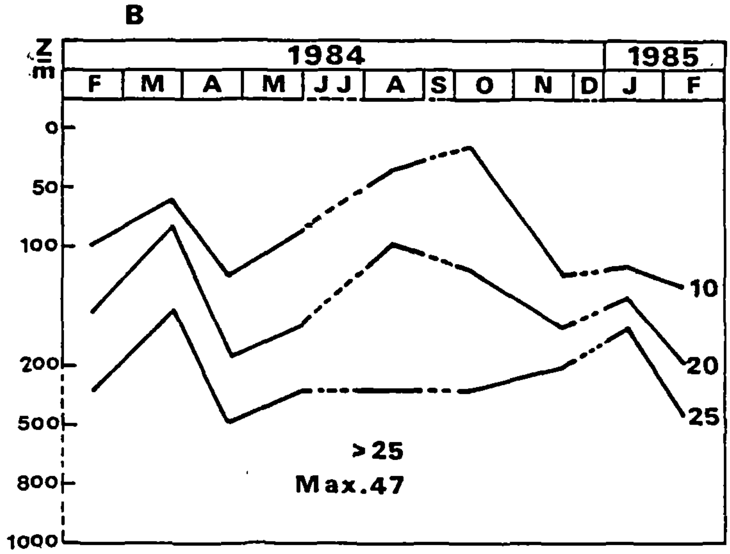


Figure 12. Temporal-spatial variations of silicate Si - Si (OH)4 concentrations (μM dm⁻³) during 1984-85 in: A) in northeastern waters (for locations see the caption of Fig. 11); B) northeastern Gulf of Aden (offshore station No. 4 of the monthly transect 45°E).

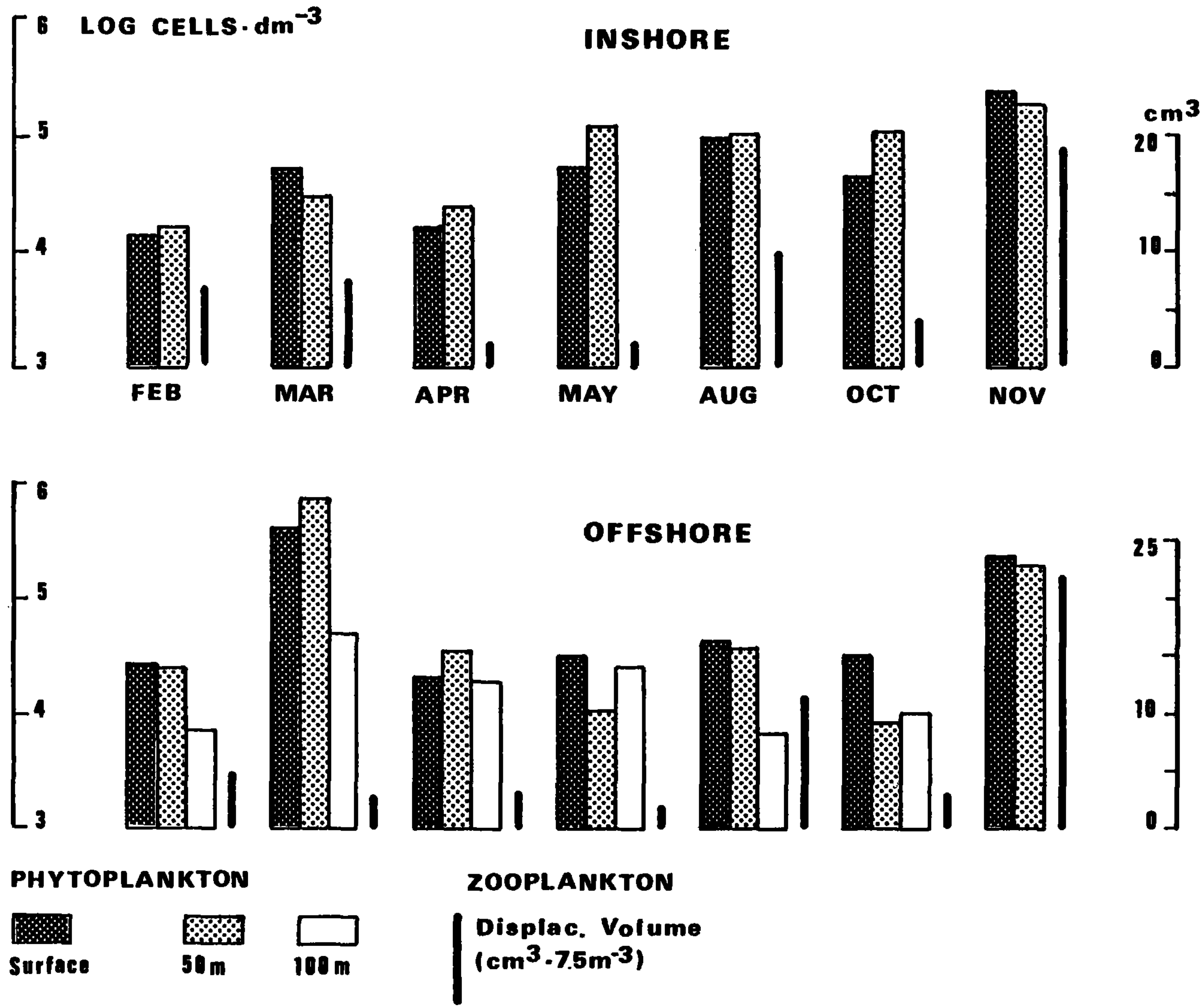


Figure 13. Temporal variations of phytoplankton abundance (in log of number of cells dm^{-3}) at the surface, 50 m and 100 m layers, and of zooplankton standing stocks of 0-30 m layer (in ml of displacement volume per haul) during 1984 in the northwestern Gulf of Aden (offshore station No. 4 and inshore station No. 1 of the monthly transect 45°E).

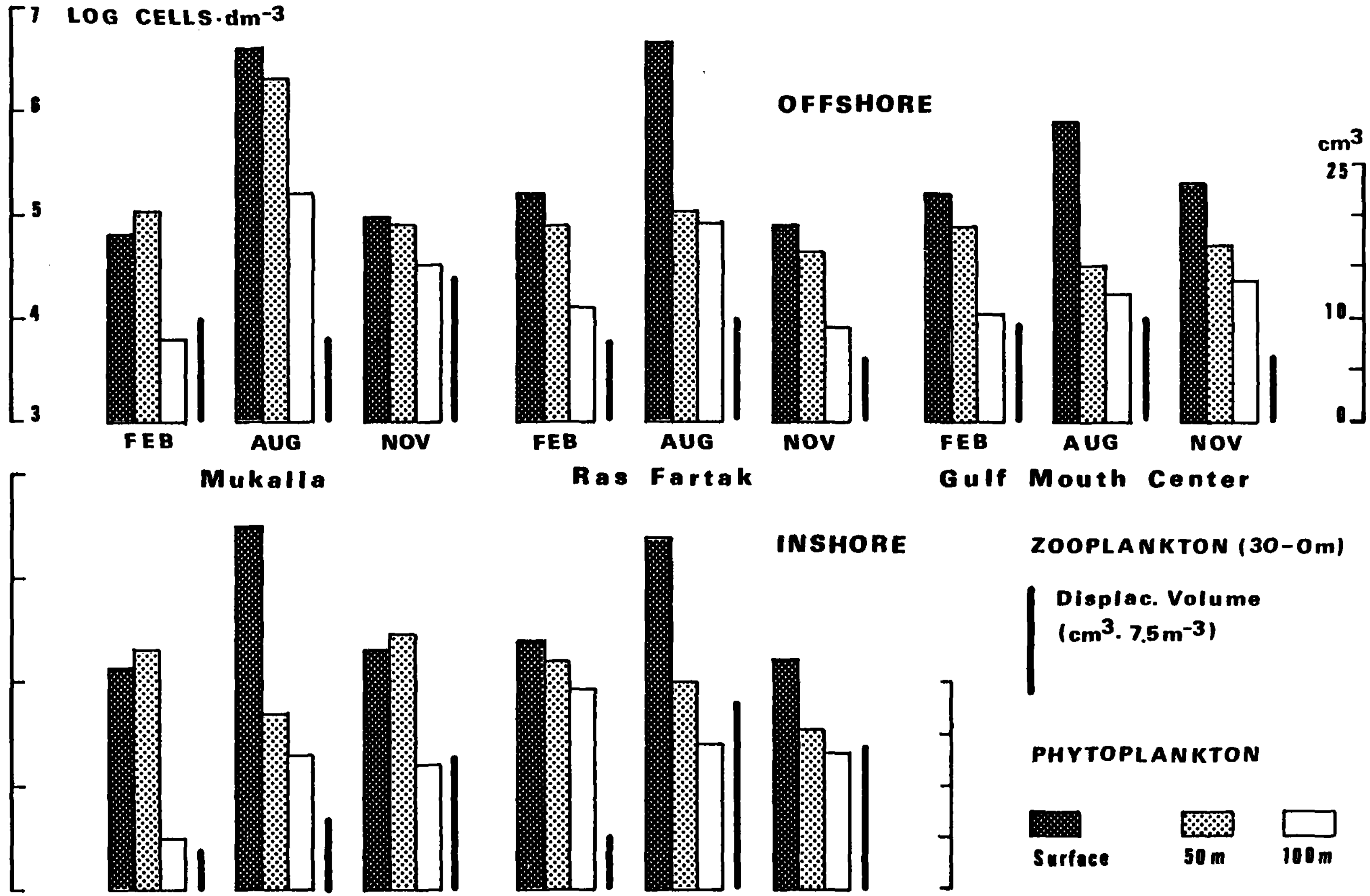


Figure 14. Temporal variations of phytoplankton and zooplankton standing stocks (for explanation see Figure 13) in the northeastern Gulf of Aden. (Mukalla: offshore Stat. No. 15, inshore No. 14; Ras Fartak: offshore Stat. No. 25, inshore No. 24; center of the Gulf's mouth offshore Stat. No. 27 of the transect Ras Fartak Mukalla).

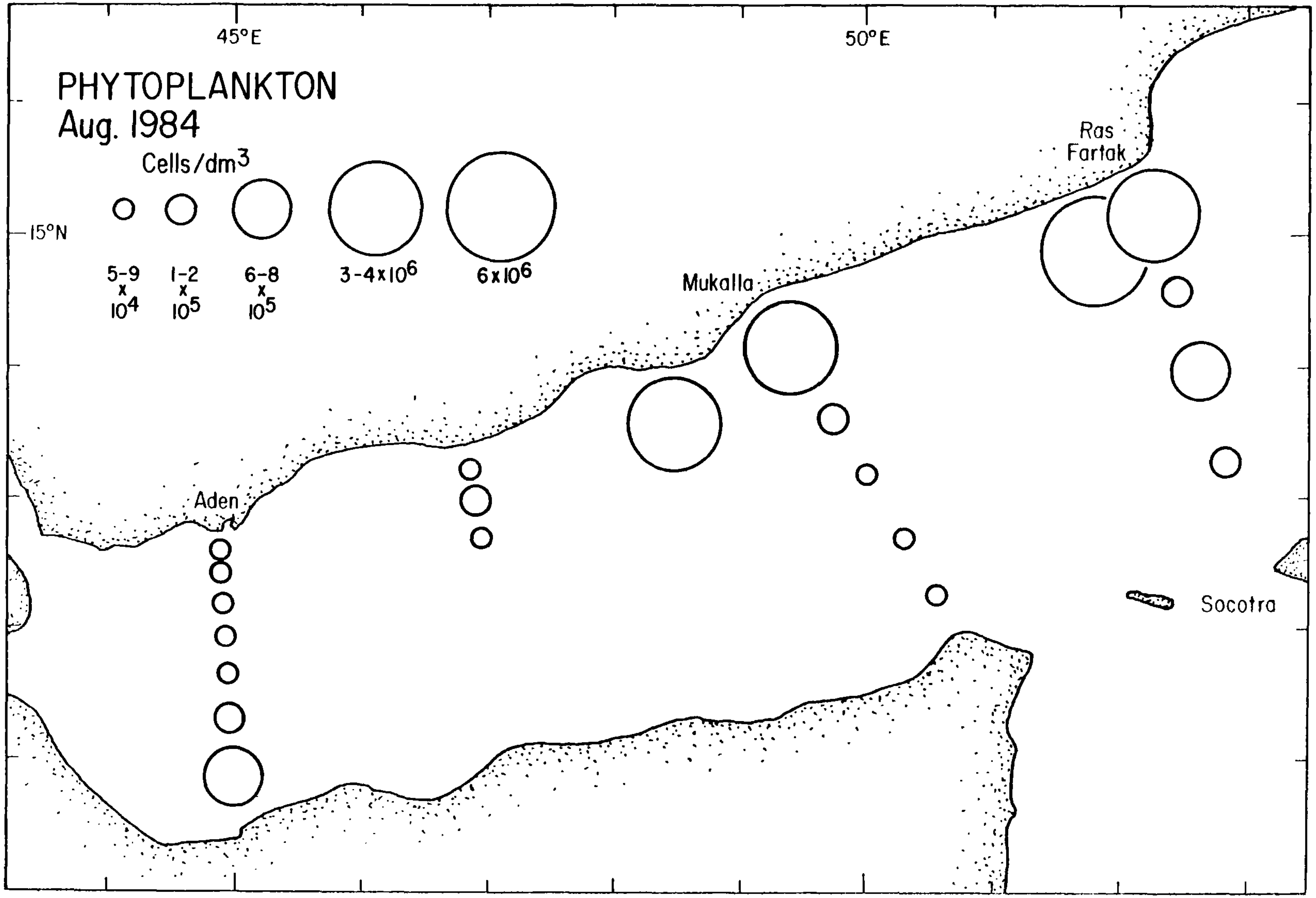


Figure 15. The distribution of phytoplankton abundance in the surface layer of the Gulf of Aden in August 1984 (in numbers of cells.dm^{-3}), showing blooms of 2-6 x 10^6 cells dm^{-3} in areas of upwelling centres.

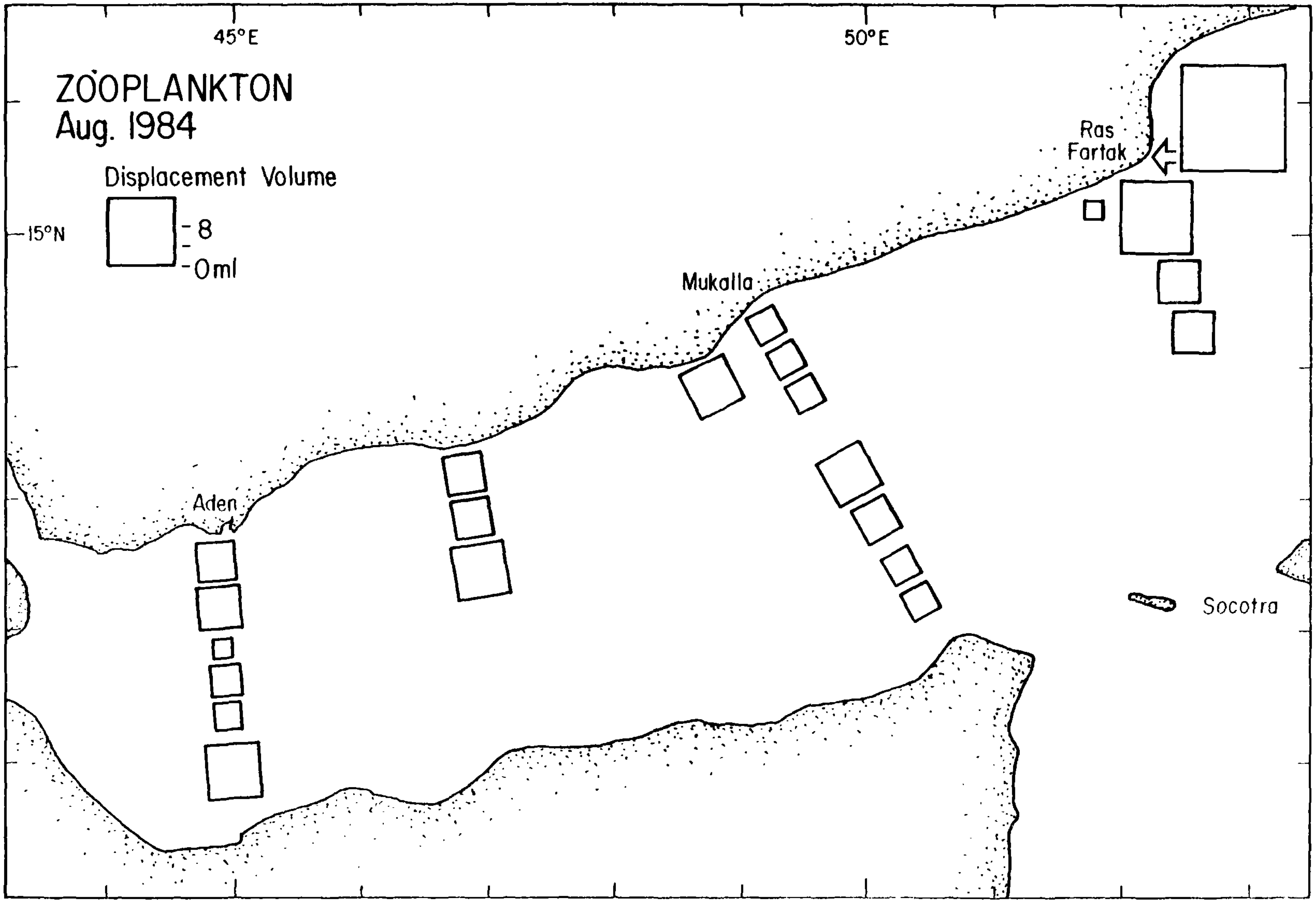


Figure 16. Distribution of zooplankton standing stocks in the surface layer (0-30 m) of the Gulf of Aden, August 1984 (in ml of displacement volume per haul), showing highest values associated with simultaneous phytoplankton bloom in Ras Fartak area.

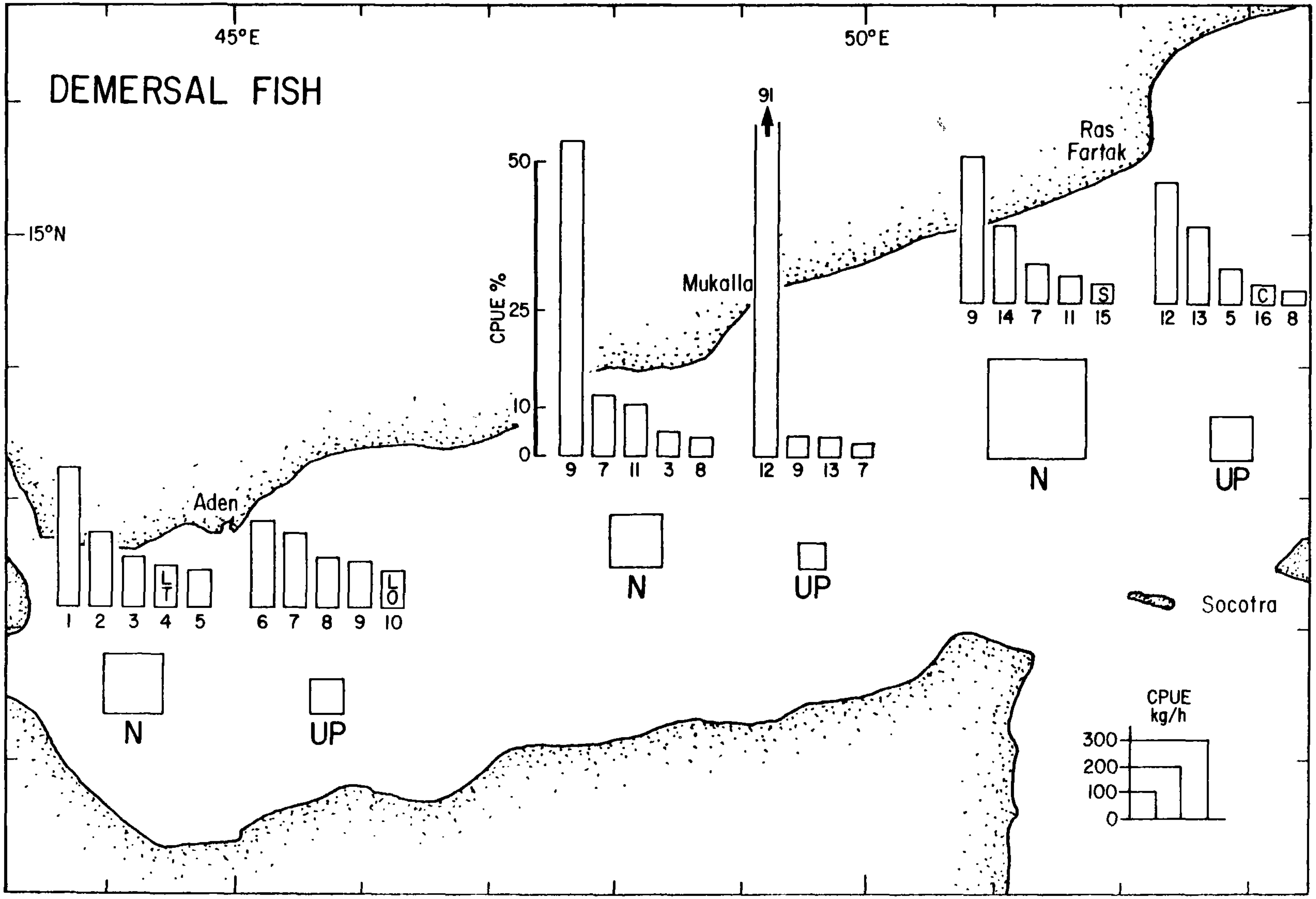


Figure 17. Trawl catches in catch per unit effort (CPUE) expressed as kg hr^{-1}) and their taxonomic composition (% of CPUE) in the northern shelf waters (average trawling depth 30 m) during non-upwelling (N) and upwelling (UP) seasons. Dominant taxa as follows:

1	-	Leiognathidae	2	-	Carangidae	3	-	Pomadasyidae
4	-	Lethrinidae	5	-	Nemipteridae	6	-	Centrolophidae
7	-	Sparidae	8	-	Synodontidae	9	-	*Sepia*
10	-	*Loligo*	11	-	Dasyatidae	12	-	Callionymidae
13	-	Triglidae	14	-	Ariidae	15	-	Sphyraenidae
16	-	Clupeidae						

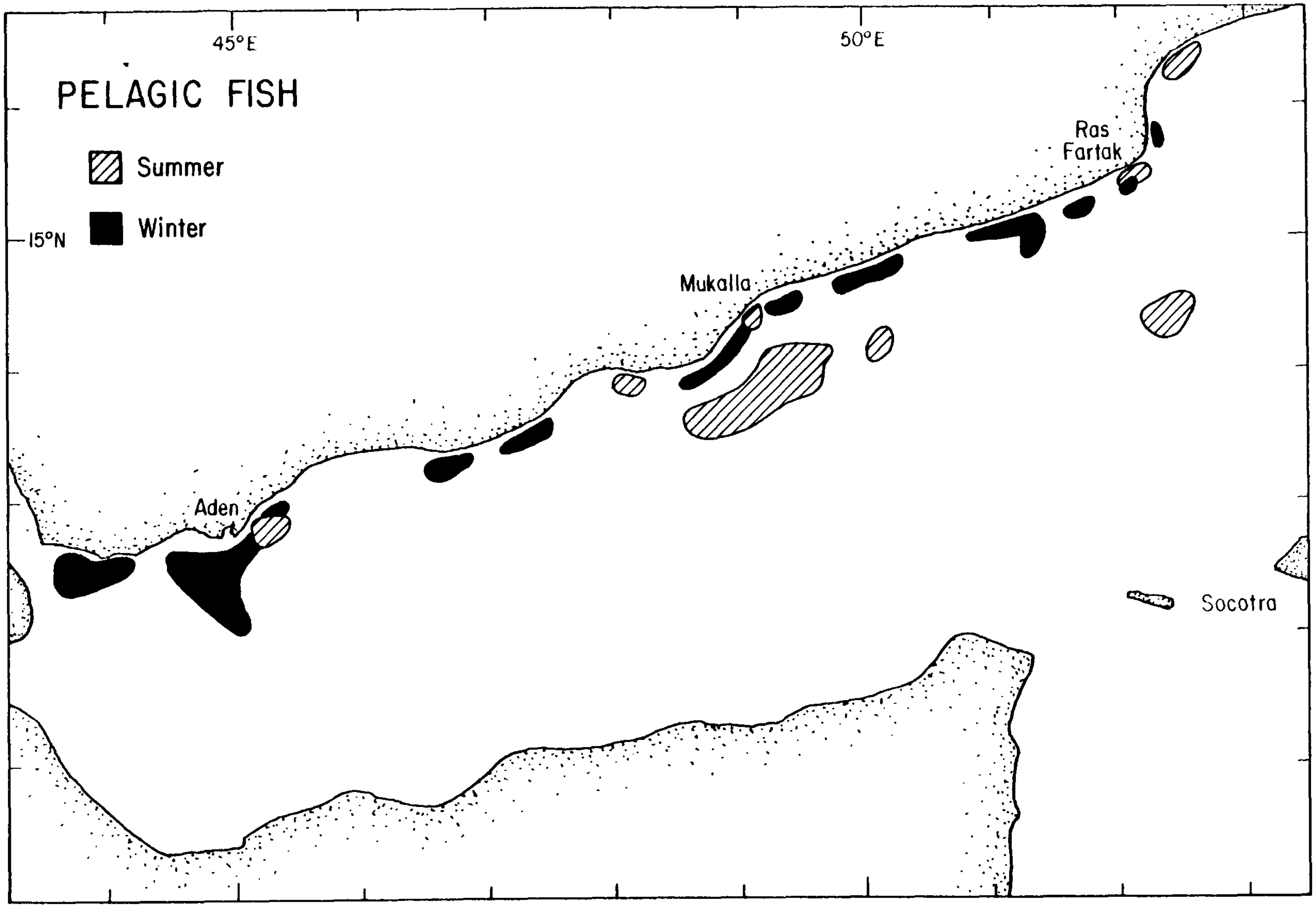


Figure 18. Distribution of small epipelagic fish in the northern Gulf of Aden obtained from integrated echo-sounding on the R/V Dr. F. Nansen, (after Institute of Marine Research, Bergen, 1984; STROMME, 1984).

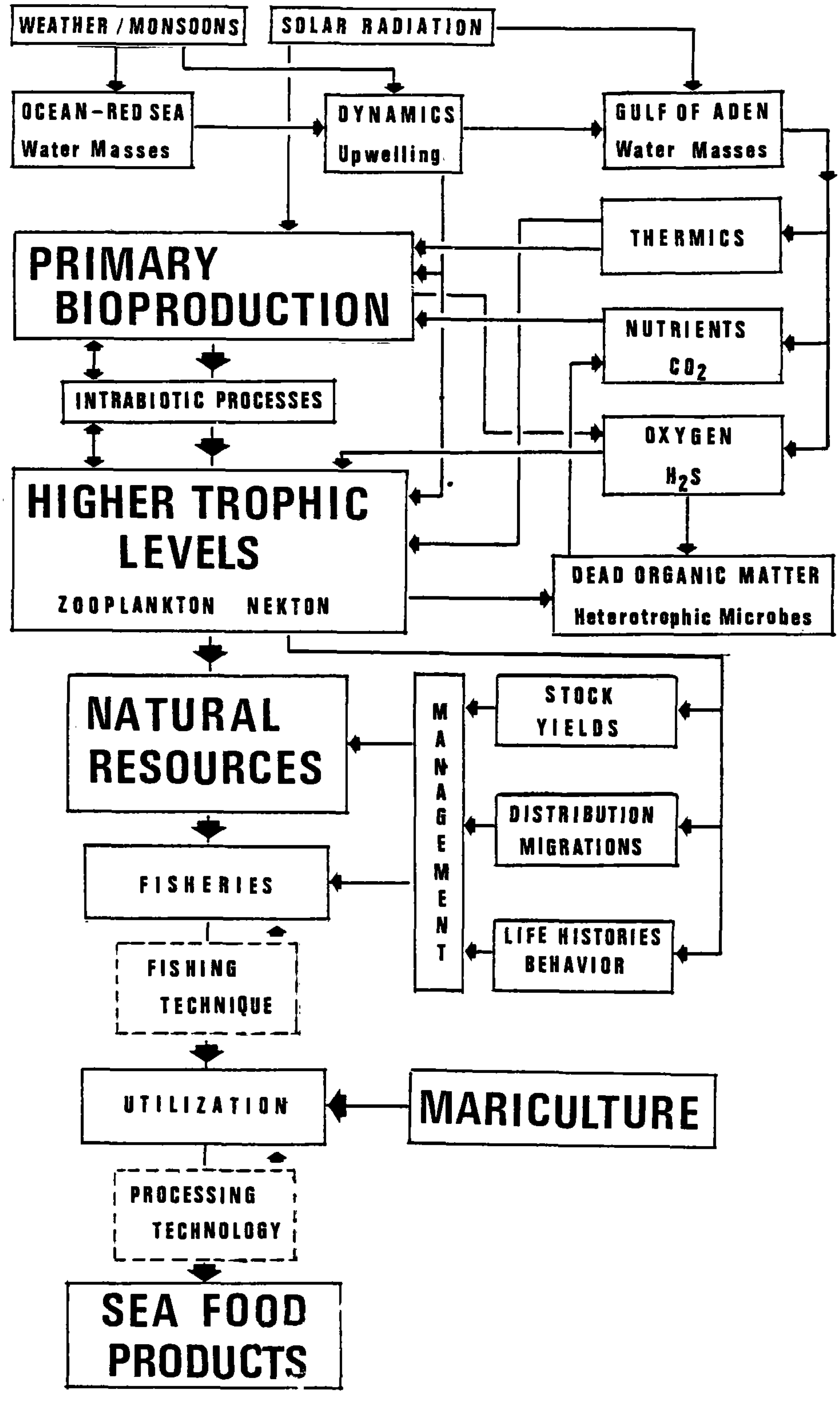


Figure 19. Schematic presentation of the main relationships within the ecosystem of the Gulf of Aden as related to exploitable living resources and fisheries.

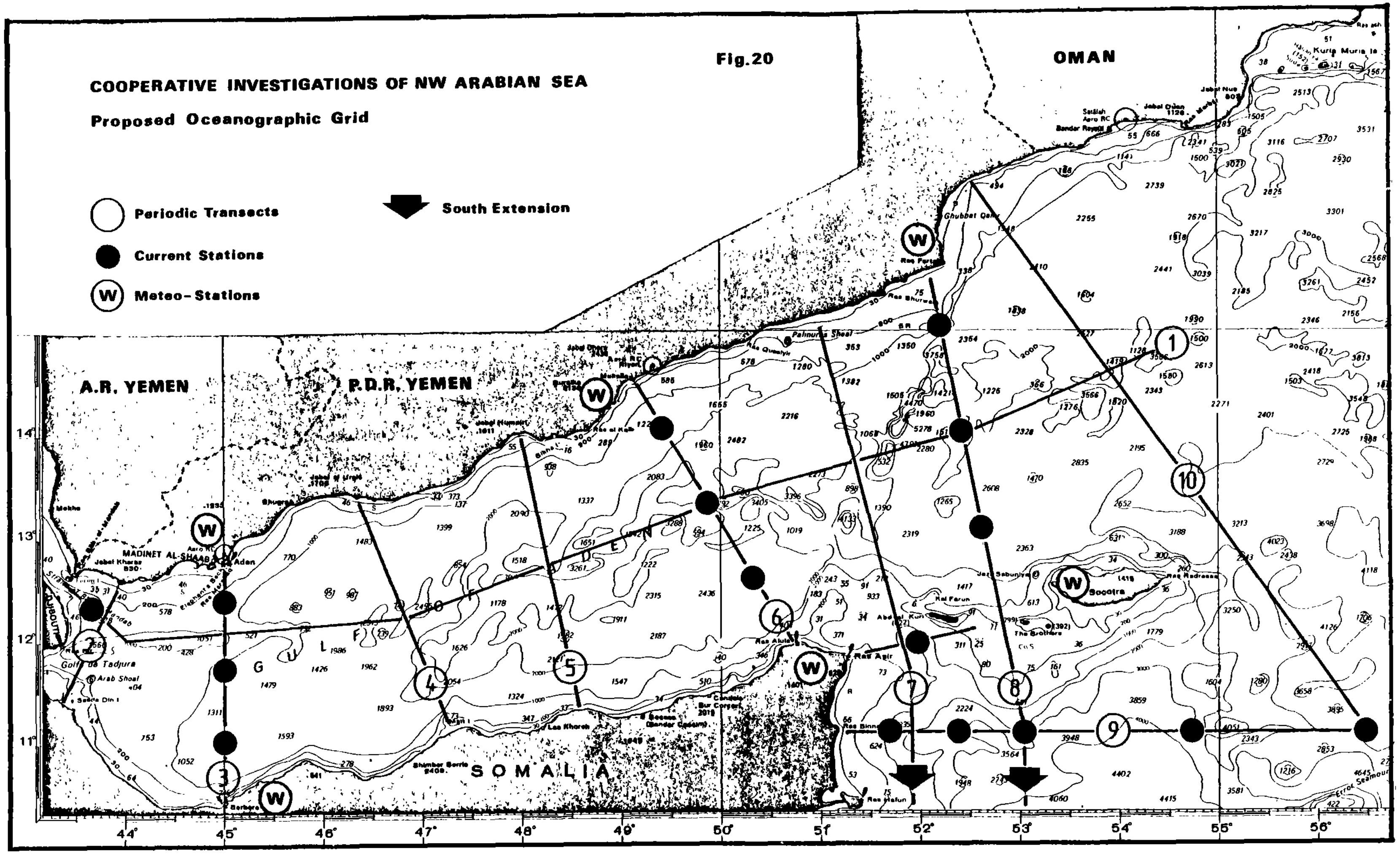


Figure 20. Map of the Gulf of Aden and adjacent areas of the Arabian Sea, showing a proposed network of co-operative oceanographic-bioproductivity investigations; explanations are given in the text.

RED SEA BIOLOGICAL OCEANOGRAPHY AND RELATED PROBLEMS

HJALMAR THIEL AND HORST WEIKERT

Institut für Hydrobiologie und Fischereiwissenschaft der Universität Hamburg, 2000 Hamburg 50, Zeiseweg 9, FRG

INTRODUCTION

Isolated from the Indian Ocean water by the shallow Hanish Sill (137 m depth) north of Bab el Mandeb, the Red Sea has unique hydrographic conditions and resulting plant and animal life. Our study of the Red Sea has been in the context of risk assessment from mining the Atlantis II Deep in the central region of this long and narrow sea (KARBE et al., 1981; THIEL et al., 1986). We present a short ecological characterization of the Red Sea, and we then identify problems that could be studied by small research groups or through international cooperative programs. Much of the ecological information concerning the pelagic ecosystem has been extracted from a review by WEIKERT (in press); results on the benthos are summarized by THIEL (in press).

THE CENTRAL RED SEA

HYDROGRAPHY

Although the hydrographic conditions of the Indian Ocean and the adjacent Gulf of Aden can be termed "normal", about 10 m below sill depth Red Sea water is 7°C warmer and 4‰ more saline than normal ocean water at a similar depth. At 2000 m depth these differences increase to about 19°C and 6‰. Surface water transport in summer is directed south throughout the Red Sea by the prevailing northerly winds. During winter, however, the southern half is governed by southern winds, and the current direction is to the north. The net northern transport is mainly due to evaporation. Our main study area was situated in the transition zone, 19-21°N.

The 100-m thick surface layer exhibits high temperatures and salinities; oxygen concentrations are around saturation, but nutrient levels are low (Figs. 1 and 2). Temperature and nutrients decrease towards the north, contrasting to the salinity and oxygen gradients. In deeper layers, temperature and salinity are markedly elevated as compared to the Indian Ocean, averaging around 21.6°C and >40.7‰. Inorganic nutrients and oxygen are inversely related to each other: Concentrations of the former reach maximum values at depths of 350-450 m in the central Red Sea (Fig. 2). The concurrent minimum oxygen concentrations are due to culminating degradation of the organic matter derived from sinking and vertical transport. This zone occurs at about 300-400 m in the south and 400-500 m in the north of the basin, reflecting the decline of biological productivity in the surface water. Beneath the intermediate waters, oxygen increases to >2 ml/l with depth, whereas the nutrient levels remain more or less constant. Except for the extreme north and south of the basin, the thermocline and the halocline prevent the nutrients in the deeper waters from being recycled into the euphotic zone. This is a characteristic feature for oligotrophic tropical ocean regions.

BIOLOGY

The central Red Sea can be described as oligotrophic, with a mean daily primary production per year of 170 mg C m^{-2} d^{-1} (WEIKERT, 1987, in press) and 258 mg C m^{-2} d^{-1} near the reefs off Jeddah (DOWIDAR, 1983). Adequate coverage in neither time nor space, however, has been achieved for the total basin. HALIM (1984), unaware of the above data nor of earlier Russian studies (KHMELEVA, 1970), gives an average for the total Red Sea of less than 100 mg m^{-2} day^{-1}, which clearly is too low. Because of low primary production, the standing stocks of zooplankton and benthos are also low. WEIKERT (1982) gives figures of 2.7g wet weight m^{-2} for the zooplankton standing stock for the upper 100 m. To depths down to about 1000 m the standing stock resembles those found in other oligotrophic seas, such as the Mediterranean Sea and the central Pacific gyre region (WEIKERT, 1982) (Fig. 3). However, the deeper waters throughout most of the Red Sea are almost depleted of zooplankton, and, as a result, deep-sea benthos is also very sparse (Table 1) (see also THIEL, 1979, 1981). The electron transport system activity (respiration potential) in sediment cores from the Red Sea also shows low values (THEEG, 1985; THIEL et al., 1987). However, although respiration of total benthic communities is low, relative to benthic densities they are high. We therefore have concluded that energy consumption for maintenance is relatively high, whereas less is available for production. We have related this shift in the usage of energy in relation to the high environmental temperature, which is encountered by all organisms in the Red Sea. Comparable results from the Atlantic and Arctic Oceans (PFANNKUCHE et al., 1983; THIEL and WEIKERT, 1984; PFANNKUCHE and THIEL, 1987) support this hypothesis.

Considering our results in an overall context suggests that the oligotrophic conditions of the central Red Sea are explained by low mixing rates in surface waters due to the unique meteorological and hydrographical settings. Mineralization of nutrients is enhanced by the high temperatures, but nutrient recycling is very slow. Primary production is limited by nutrient availability at low mixing rates, yet it occurs throughout the year. The accelerated degradation of organic matter in the water column leaves little food for bathypelagic and bathybenthic species, which occur in unusually low numbers and presumably have low productions. This short characterization from our results in the central Red Sea evidently is generally applicable to the entire Red Sea, with some local exceptions.

OTHER RED SEA REGIONS

The strongest deviations from the results described for the central Red Sea are found in the southern regions, undoubtedly due to inflow of water from the Indian Ocean. Nutrient import seems to be strongest during the summer months into the lower part of the euphotic zone (KHIMITSA and BIBIK, 1979; POISSON et al., 1984). As a result, nutrients can increase by up to 25% above levels in the central Red Sea. This stimulates primary production by as much as 300% (compared to the data of WEIKERT, 1981 and KHMELEVA, 1970) and, subsequently, zooplankton standing stocks as much as 100% (BECKMANN, 1984). Water masses from the Gulf of Aden also transport living and dead organic matter which fertilizes the southern Red Sea. As far north as 16°N BECKMANN (1984) found imported plankton from the Indian Ocean that was bound to die, as most species cannot withstand the more severe conditions in the Red Sea. The higher production and the import of organic matter are thought to be responsible for lower oxygen concentrations throughout the water column of the southern Red Sea.

North of the central region nutrients, primary production and standing stocks of zooplankton decrease, whereas oxygen concentration increases, both in the surface and subsurface waters, due to lower temperatures and diminished degradation of organic matter. The decrease in nutrients appears to be most pronounced in the phosphates. While 0.1-0.3 $\mu mol\ l^{-1}$ were determined in the euphotic zone in the south, only 0.01-0.05 $\mu mol\ l^{-1}$ were found in the north. In the Gulf of Aqaba, recycling of nutrients is achieved by vertical convection in the winter period, whereas during summer a strong thermocline stabilizes the water column (LEVANON-SPANIER et al., 1979). Consequently, the Gulf exhibits a moderate primary production in winter that decreases with the formation of the thermocline (LEVANON-SPANIER et al., 1979). In contrast, the Gulf of Suez seems to be well mixed from persistent winds so that a medium primary production might be expected throughout the year.

According to SUKHANOVA (1969) and KHMELEVA (1970), primary production in the northern Red Sea may be half that in the central region. For the zooplankton, a similar decrease to the north is obtained from the data of DELALO (1966) and GORDEYEVA (1970). The ecological consequences of predicted seasonal upwelling events in the farthest north, however, have not yet been investigated.

Benthic studies in the deep waters have not demonstrated latitudinal differences in the Red Sea, although it is known that deep benthos standing stocks should mirror surface water productivity. No explanation is available for these observations. The low values for benthic standing stocks, together with a variability in faunal densities and in sedimentary conditions, may obscure existent differences. However, MURINA (1971) described a north-south increase of benthic standing stocks from samples mainly collected along the narrow shelf, i.e., in depths close to the epipelagic zone. Summarizing the data, she found benthic biomasses in the south to be four times those in the north.

PROPOSED RESEARCH

Our limited knowledge allows us to draw only a general picture of the oceanic system in the Red Sea. However, we must bear in mind that the Red Sea basin contains a unique environment and resultant biocoenoses as determined by the basin's size (compared to other oceans), young geological age, isolation from the Indian Ocean, hydrography, and biology.

The characteristics and adaptations within the Red Sea are worthy of study on their own right but also for comparisons with other oceans. Cooperative oceanographic investigations using the Red Sea as a tropical ocean ecosystem model can help us describe and understand the world ocean. For example, relative to other warm seas, this basin obviously has a less complex pelagic and benthic system, since diversity in the zooplankton, micronekton and benthos seems to be strongly reduced.

Within such a broad context we propose that long-term studies be initiated from several research institutions along the Red Sea coasts (and Indian Ocean), applying intercalibrated methods. We suggest the following topics:

- assessment of primary production throughout the year, including specific observations on the ecologically important Cyanobacteria (*Oscillatoria*) blooms.
- assessment of zooplankton production (different size classes) and micronekton, based on total standing stock and biomass and on population studies. The demographic and genetic consequences of the import of Gulf of Aden species' to the Red Sea needs to be considered. In this context, species reproductive capacities in different geographic regions, their depth distribution and vertical migrations are of particular interest.
- assessment of benthic production, taking into account algae, corals and other animals from the shelf and in deeper regions.
- assessment of organic matter transport between coral reefs (shelf) and oceanic regions.
- assessment of brine-influenced benthic communities in the deep Red Sea.

Many of the pertinent questions, basic or applied, can be combined fruitfully in common research programmes.

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Table 1. A comparison of meiofauna densities (individuals 10 cm^{-2}; values rounded, omitting Foraminifera) from different areas. Summarized from various sources (THIEL, 1983).

Region	approximate water depth in meters 500	1000	2000
Red Sea	400	50	100
Western Mediterranean	--	1200	500
East Atlantic: NW Africa	1500	1200	300
East Atlantic: Gulf of Biscay, Portugal	800	700	500
Norwegian Sea: Iceland Faroe Ridge	2600	2000	1000
North Atlantic: Iceland Faroe Ridge	1000	40	300

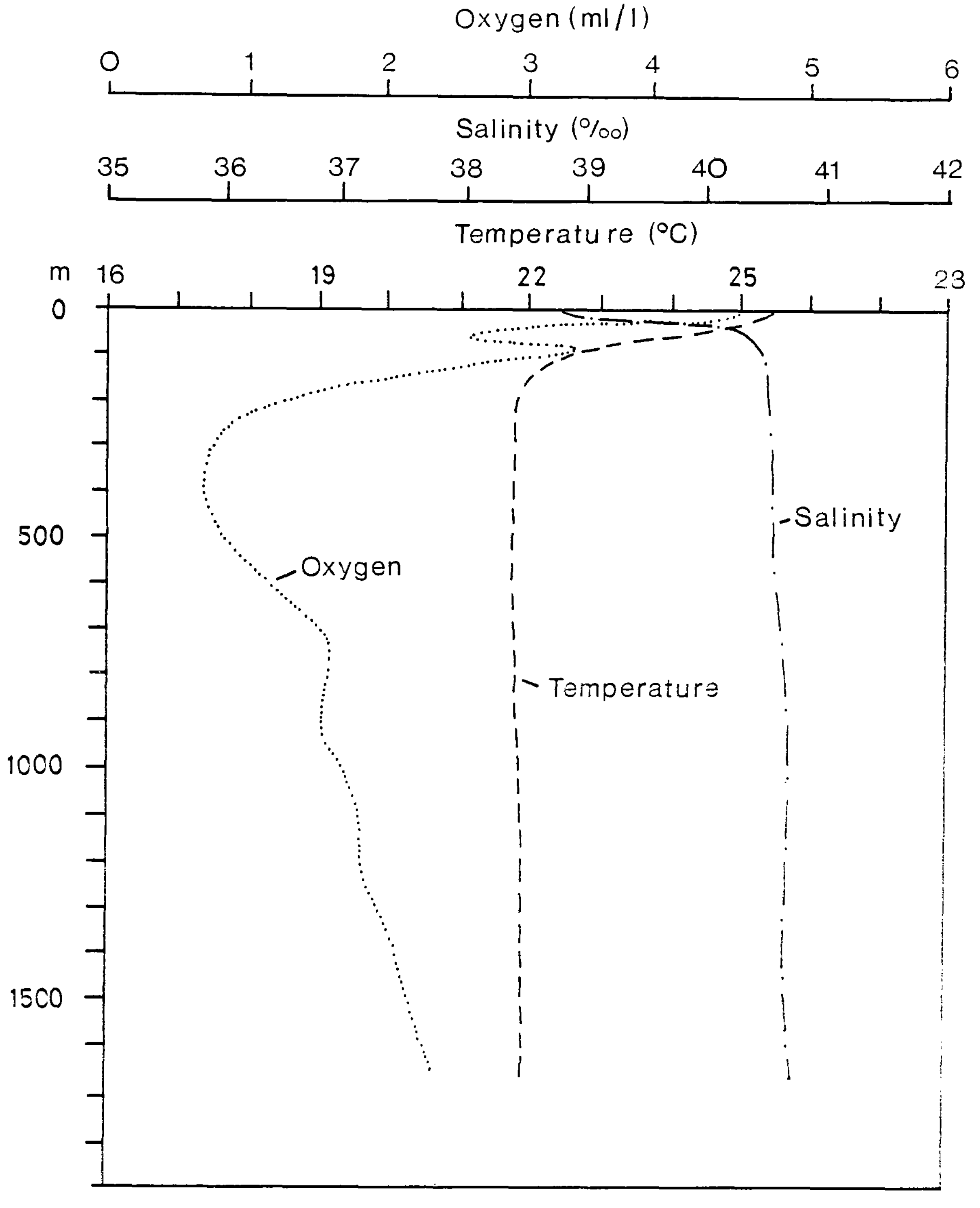

Figure 1. Depth profiles of temperature, salinity and oxygen in the central Red Sea (from KARBE *et al.*, 1981).

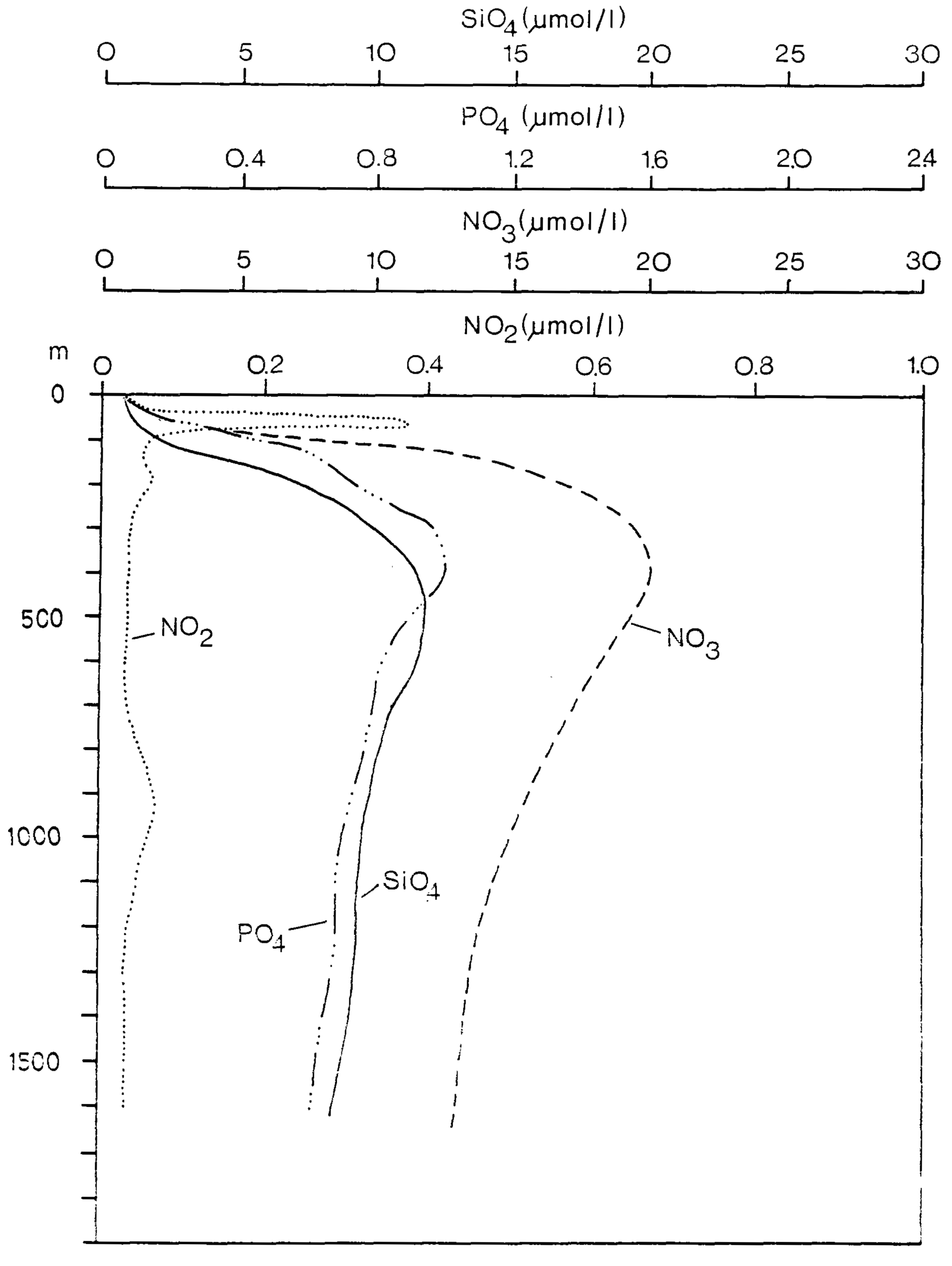

Figure 2. Depth profiles of nitrate, nitrite, phosphate and silicate in the central Red Sea (from KARBE et al., 1981).

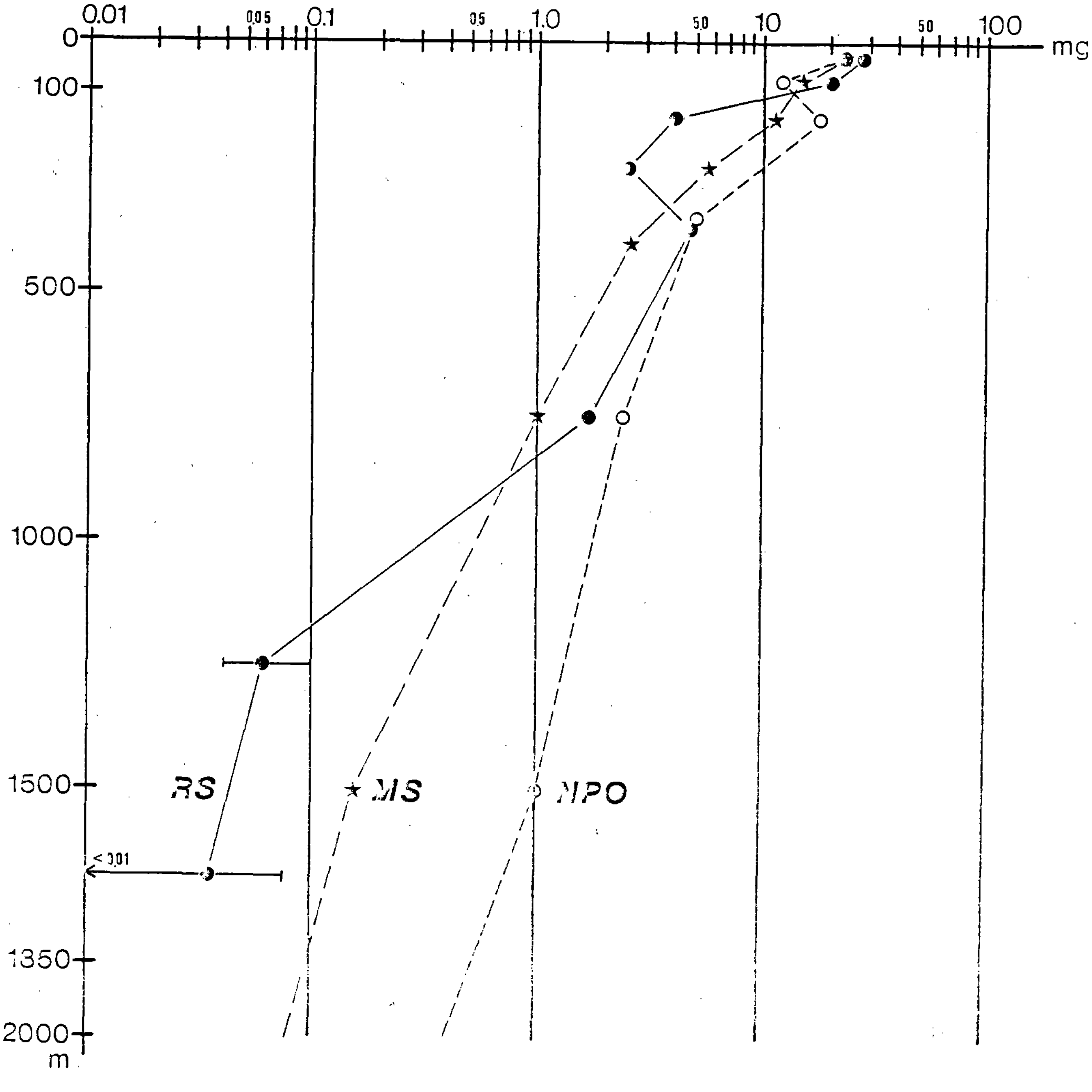

Figure 3. Vertical distributions of plankton biomass (wet weight in mg) in the North Pacific Ocean (NPO), in the Mediterranean (MS) and in the Red Sea (RS) (from WEIKERT, 1982).

THE CRUST BENEATH THE RED SEA — GULF OF ADEN RIFT SYSTEM:

A REVIEW[1]

HOW KIN WONG AND EGON T. DEGENS

Geologisch-Paläontologisches Institut und Museum der Universität Hamburg
Bundestraße 55, D-2000 Hamburg 13, FRG

ABSTRACT

The Red Sea — Gulf of Aden — East African Rift System is one of the key areas to study rift-drift processes. Within this single system, geodynamic processes ranging from uplift, block-faulting, volcanism, and seismic activity, to hydrothermal circulation and seafloor spreading are represented. In this paper we review the geological, geophysical, sedimentological and geochemical characteristics of the Red Sea-Gulf of Aden, with special emphasis on the structure and evolution of the crust, as well as on their implications to continental rifting and plate accretion.

INTRODUCTION

Northern and northeastern Africa has always attracted the attention of geoscientists, particularly from the hydrogeological, tectonic and geomorphological points of view. With the recent popularity of plate tectonics, this area together with its neighbours to the east and south has gained even greater significance. While the East African Rift is generally believed to be characterized by continental rifting, the Gulf of Aden and the Red Sea already have undergone active seafloor spreading. These three tectonic features meet at the Afar Depression, a triple junction with active volcanism and earthquake activity attesting to ongoing tectonism.

Rifting is a process generally believed to precede seafloor spreading. Studies of continental rifts in Europe, North America and Africa (ILLIES and FUCHS, 1974; ILLIES and MUELLER, 1970; MAGUIRE and KHAN, 1980; NOLET and MUELLER, 1982; PRODEHL, 1979; RAMBERG and NEUMANN, 1978) suggest that rifting may start with the diapiric intrusion of sub-lithospheric material from a depth of several hundred kilometers into the lithosphere. A zone of tension is produced along which partition of the continental crust eventually occurs. OXBURGH and TURCOTTE (1974), on the other hand, argued that the tensional failure at the earth's surface may simply be a response to membrane stresses set up as the plate moves over a non-spherical earth. Other unresolved questions include the spatial distribution of hypocenters in a rift system, the seismic velocity structure of the crust and upper mantle within the graben and over its flanks, and the detailed structure of the crust/mantle transition.

The process of crustal formation at oceanic ridges is still not well-understood. It is generally agreed that under the ridge crest, a mixture of mantle material and basaltic melt is being injected into the crust. The basaltic melt, being lighter, migrates upwards and accumulates near the top of the injection diapir as magma chambers (BALLARD and VAN ANDEL, 1977; BRYAN and MOORE, 1977; LEWIS, 1983). These magma chambers are very limited in lateral extent (at most a few km), and it is unlikely that they represent a steady state phenomenon, certainly not in the Atlantic. The extrusion of pillow basalts and lava flows from such magma chambers onto the seafloor is episodic, both in space and in time. Under the ridge axis, plate-driving forces produce an extensional strain which facilitates hydrothermal circulation, since crack propagation as a result of convective cooling is enhanced. Eventually, hydrothermal alteration of the crust and infilling of cracks by precipitation take place. Whether the Moho represents a petrologic boundary or is determined by the depth to which hydrothermal penetration extends is still not resolved.

[1] Modified from a paper that appeared in Mitt. Geol.-Paläont. Inst., Univ. Hamburg, Georg Knetsch Heft 56, p. 53-94.

THE RED SEA

PHYSIOGRAPHY

The Red Sea is an elongated depression oriented northwest-southeast. It has a length of 1800 km (2100 km including the Gulf of Suez), and a maximum width of almost 400 km. The maximum recorded depth is 2920 m and the total water volume reaches 215,000 km^3 (Fig. 1). At its southern end (16°-17°N) the coral-island Farasan and Dahlak Archipelagoes rise to slightly over 60 m above sea level. Farther south, the Red Sea narrows to the Strait of Bab-el-Mandeb, to the north of which a series of volcanic islands occupy the medial section. In the Red Sea proper a very distinct main trough extends northwards from the Zebayir Islands to the southern tip of the Sinai Peninsula. An axial graben is cut into this main trough (Fig. 2; BÄCKER et al., 1975; LAUGHTON, 1970; ROSS et al., 1969), varying from 35 to 80 km in width, and up to 2100 m deeper than the adjoining seafloor of the main trough. In the northern part of the Red Sea, the axial trough is either very narrow (19 km at 23°N) or may even be absent, being replaced by a U-shaped depression that is not fault-bounded. Displacement of the talwegs and abrupt termination of the longitudinal tensional faults suggest that the axial trough is offset by a number of northeast-striking lineaments (BÄCKER et al., 1975). These lineaments mark the location of transform faults and divide the medial part of the Red Sea between 19.5° and 20.8°N into trough and inter-trough zones. The trough zones possess characteristics typical of an active rift, while the inter-trough zones are devoid of large magnetic anomalies (ELTAYEB et al., 1979; SEARLE and ROSS, 1975) and have a surficial sedimentary layer of up to 1 km in thickness. Whether these inter-trough zones represent areas unaffected by the present seafloor spreading phase is still a matter of debate.

The Red Sea shelf is shallow and flat. North of Jiddah (about 22°N), Saudi Arabia, the shelf is narrow and interrupted by sharp topographic breaks. The adjoining coastal plain is likewise narrow, with terraces well above the present shoreline. South of Jiddah, both the shelf and the coastal plain broaden until they merge imperceptibly with each other (Fig. 2). The general aspect is one of emergence to the north, and of submergence to the south.

On the Arabian Peninsula, a magnificent erosional escarpment sub-parallel to the coast extends from the latitude of Jiddah to Yemen, forming the crest of the tilted Arabian block (Fig. 1). In Yemen, altitudes of over 3700 m are reached; a mean height of more than 2000 m is found in Saudi Arabia. No major normal faults parallel to this escarpment have been mapped between the coastal plain and the crest. In Ethiopia a similar scarp bounds the coastal plain on the west, reaching altitudes of over 3000 m. At the latitude of Massawa (15.7°N), the scarp turns south, leaving the Red Sea trend to join the East African Rift. The Afar Depression borders the southwestern Red Sea. Here extensive volcanism and block faulting have given rise to a complicated morphology.

Escarpments and volcanic rocks are absent on land bordering the Red Sea north of Jiddah (Fig. 1). Instead, the lower coastal slopes merge seaward as interfingering fans. In the extreme north, the Red Sea bifurcates into the Gulf of Suez and the Gulf of Aqaba, the former being controlled primarily by normal faulting and the latter by left-lateral strike-slip faulting with a small extensional component. Narrow coastal plains ascend directly into fault-bounded basement blocks within a few kilometers of the shoreline.

GEOLOGY

The stratigraphic record shows that the Jurassic seas spread northward to the latitude of Massawa (16°N) and southward to 29°N across the Sinai Peninsula. By the end of the Cretaceous the Tethyan Sea had reached southwards to approximately 21°N (WHITEMAN, 1968; CARELLA and SCARPA, 1962). However, in the south, the Cretaceous marine facies is regressive, with the shoreline extending northwards only to 13°N. There is no evidence that the Red Sea depression existed during this time. Tectonic activity during the Mesozoic in and around the Red Sea was generally weak or lacking, the only major exception being the beginning of the emplacement of the thick trap series in Yemen, where basalts are found interbedded with Cretaceous sedimentary rocks.

The early Tertiary sedimentary record in the Red Sea area is rather incomplete. It is generally accepted that during the Eocene, transgressive marine deposits developed in the northern Sinai, in Egypt, and in northern Sudan (CARELLA and SCARPA, 1962; SAID, 1962; WHITEMAN, 1968). On the other hand, the Horn of Africa had completely emerged by the end of the Eocene. Deposition of the Red Series in the Afar Depression began in the late Oligocene (BARBERI et al., 1972), and Oligocene and Miocene marine sediments are found on the southern coast of the Arabian Peninsula.

During the Oligocene, separate, extensive uplifts occurred on either side of the Red Sea. This was accompanied by the outpouring of alkaline basalts that now form a thick trap series on the high plateaus behind the Arabian and Ethiopian scarps. The Miocene sedimentary sequence is characterized by a great thickness of clastics (2—3 km) and evaporites (3—4 km) in the Red Sea marginal trough and the surrounding continents (BROWN, 1970; LOWELL and GENIK, 1972; ROSS et al., 1973). During this time, volcanic activity continued in the Afar Depression and in western Saudi Arabia. As the Red Sea depression subsided further, uplift continued in the Arabian-Yemen block on the east and the Ethiopian segment on the west. It should be noted that during the deposition of the massive salt deposits, the Red Sea probably remained connected to the open ocean. Tectonic subsidence accompanied by brine migration to the morphological deeps and final precipitation of salt minerals are responsible for the observed evaporite sequence which locally reaches 5 km in thickness.

At the beginning of the Pliocene a dramatic change in the sedimentary record occurred, whereby the Miocene evaporitic sequence gave way abruptly to marine oozes and marginal clastics. Within the Red Sea axial trough, a thin veneer of sediment directly overlies basalt that has extruded into the axial trough during active seafloor spreading (GASS, 1970; SCHILLING, 1969).

MAGNETICS

The Red Sea has very large magnetic anomalies over the axial trough, with amplitudes of the order of 2000 nT (nano-Teslas or gammas) diminishing northwards. In contrast, magnetic anomalies over the main trough are much smaller, usually a few hundred nT (ALLEN, 1970; GIRDLER and STYLES, 1974; PHILLIPS et al., 1969; PHILLIPS, 1970; ROESER, 1975, 1976; SEARLE and ROSS, 1975). Apart from anomalies 1 and 2 in the axial trough, the magnetic pattern cannot be correlated over the entire length of the Red Sea. In part this is due to the paucity of good magnetic data extending from coast to coast, particularly in the northern Red Sea.

Figure 3a shows a composite magnetic profile obtained from stacking 6 adjacent aeromagnetic profiles from the central Red Sea and then reducing it to the pole. Beneath this is a synthetic profile generated using the reversal time scale of TARLING and MITCHELL (1976). The magnetism is assumed to reside in layer 2 for the new crust in the axial trough and in layer 3 for the older crust beneath the main trough. The composite and synthetic profiles are directly comparable only if two seafloor spreading phases and a 70 km westward drift of the spreading axis — when the second episode of spreading began — are assumed (GIRDLER, 1978; NOY, 1978). In this model the more recent spreading phase started about 4.5 m.y. ago and has continued up to the present at a rate of 0.7 cm yr^{-1}. The anomalies of the earlier phase suggest that it started about 30 m.y. B.P. (upper Oligocene) and continued to about 15 m.y. B.P. (mid-Miocene). To illustrate the uncertainties in the interpretation of these anomalies, a second model is shown in which the same dates are used for the recent phase of spreading but younger dates, namely all within the Miocene from 20 m.y. to about 6 m.y. ago, for the earlier phase. Here a 27.5 km westward shift of the spreading center is required at the re-initiation of spreading (NOY, 1978; Fig. 3b). From the magnetic anomalies alone it is impossible to choose between these two possible models.

In the northern Red Sea between 26° and 28°N a magnetic profile studied by GIRDLER and STYLES (1976) suggests a low half-spreading rate of 0.5 cm yr^{-1} over the past few million years. Between 20.5° and 22°N in the axial trough, large magnetic anomalies have been interpreted by SEARLE and ROSS (1975) to represent an inter-trough zone (the Atlantis-Hatiba inter-trough zone) over which magnetic anomalies are absent. This absence has been explained by the magnetic end-effect model whereby a number of short segments of spreading axis are assumed to be offset by small transform faults. These transform faults are associated with zones of very low (possibly zero) intensity of magnetization which could be a result of destruction of coherently magnetized bodies by intense brecciation, by leaky intrusive activity along zones so narrow that the magnetizations are effectively randomly distributed, by intrusion into sediments, or by virtue of the very different mineralogy of fracture zone basalts. The axial anomalies are compatible with the seafloor spreading hypothesis. From 16° to 19°N, ROESER (1975, 1976) identified the axial anomalies, thereby establishing isochrons from 0—5 m.y., although the 5 m.y. isochrons are considered doubtful. Magnetic anomalies in the southernmost part of the Red Sea are not well established.

Earlier studies suggested that the initial phase of spreading might have occurred between 34-41 m.y. ago, from Late Eocene to Early Oligocene (GIRDLER and STYLES, 1974). Although there is evidence that tectonic activity in the Red Sea and in the Gulf of Aden may have begun in Jurassic or Cretaceous times, most investigators agree that separation started in the Miocene following an Oligocene-Miocene phase of uplift and faulting (BEYDOUN, 1970). The pronounced phase of Oligocene and Miocene volcanism in the Red Sea area was probably the forerunner of the major episode of rifting in the Red Sea. The mid-Miocene age for the initiation of spreading also agrees with the results of the Deep Sea Drilling Project in the Gulf of Aden, where 13 m.y. old sediments were drilled immediately above the acoustic basement close to the 2000 m isobath, just beyond the area of correlatable magnetic anomalies (FISCHER, BUNCE et al., 1974). In addition, the volcanism that accompanies initial fragmentation in the Afar Depression is earliest Miocene, 25 to 23 m.y. old (BARBERI et al., 1975). Finally, the water depth over the marginal trough of the Red Sea (1—1.5 km) is more compatible with a mid-Miocene age. The predicted depth to oceanic basement according to the cooling plate model for a 20 m.y. old lithosphere is about 4 km. Correcting for the sediment cover, which includes up to 5 km of Miocene evaporites, the expected water depth would be 1.3—1.8 km. On the other hand, should the crust be about 38 m.y. old (implying that the initiation of spreading was at the Eocene-Oligocene boundary), the theoretical corrected water depth would be 2—2.5 km. This gives a much larger discrepancy compared to observations. Of course, the alternative that the observed depths are accounted for by a subsided, attenuated continental crust cannot, on the basis of isostasy alone, be discounted.

It should be noted that the development of the Red Sea inevitably has been influenced by that of the Gulf of Aden. Recent thoughts on the Gulf suggest that the possibility of seafloor spreading taking place in a single phase starting about 16 m.y. cannot simply be dismissed. The implication is that a similar single-phase development in the Red Sea also is possible.

SEISMICITY

Earthquake activity in the Red Sea and Gulf of Aden is confined largely to the axial trough (Fig. 4). It is undoubtedly related to active seafloor spreading, although surprisingly few earthquakes appear to occur in the northern part of the Red Sea (FAIRHEAD and GIRDLER, 1971; McKENZIE et al., 1970; SYKES, 1968). The focal depths are all shallow, usually less than 100 km. The 3 fault-plane solutions available so far indicate strike-slip motion in a NE-SW direction in the central and southern Red Sea, suggesting that the central rift may be offset by a number of transform faults, whereas the solution at the Gulf of Suez is characterized by normal faulting (McKENZIE et al., 1970; SYKES, 1968). While these solutions give an indication of the tectonic pattern of this area, they are not accurate enough to be applied quantitatively. Epicenters in the Afar Depression are largely along the margin rather than being confined to the axial zone.

GRAVITY

Bouguer gravity anomalies are positive over both the main and axial troughs. In cross-section the axial trough is characterized by a slight minimum (ALLEN, 1970; FLEISCHER, 1969; PHILLIPS et al., 1969; PLAUMANN, 1975; QURESHI, 1971). Such a pattern is typical of mid-ocean ridges and may be attributed to plate accretion. A maximum Bouguer value of 150 mgal has been measured over the axial trough. However, over the inter-trough zone within the axial trough, the Bouguer gravity reaches only 90 mgal. This reduced value can be attributed to the presence of sediments in the inter-trough zone and the resulting lithospheric isostatic response, while elsewhere the axial trough is devoid of sediments (SEARLE and ROSS, 1975).

Modelling of Bouguer anomalies along a 900 km profile across the Red Sea at 20°N suggests that oceanic crust underlies almost the entire width of the Red Sea at this latitude, while the intrusive zone takes up half of the seafloor (BROWN and GIRDLER, 1982). Lithospheric thinning extends to 120 and 180 km landward of the coasts.

In contrast to the Red Sea, Bouguer anomalies in the Gulf of Aqaba reach -100 mgal. Likewise, negative values are typical of the Gulf of Suez. In both areas the negative anomalies are interpreted as a result of transform faulting. Bouguer anomalies in the Afar Depression are largely negative, apparently a result of relief. Positive anomalies are present only over volcanic centers and along the coast to the northeast as well as parallel to the Danakil Horst. Their amplitudes, however, are less than those over the axial trough of the Red Sea.

HEAT FLOW

Nearly 100 reliable heat flow measurements are now available from the Red Sea (ERICKSON and SIMMONS, 1969; GIRDLER et al., 1974; HAENEL, 1972; SCHLEUCH, 1973). They suggest that the entire region is one of the high heat flow (Fig. 5). Over the axial trough an average of 467 ± 116 mW m^{-2} (around 11.1 HFU) has been observed, decreasing to 111 ± 5 mW m^{-2} (2.64 HFU) 50 to 270 km from the spreading axis (GIRDLER and EVANS, 1977). The high variability over the axial zone can be attributed to the effect of hydrothermal circulation in the intrusive zone and to the effects of evaporitic flow as older crust fractures apart (GIRDLER and WHITMARSH, 1974). Heat flow along the coast, averaging 112 mW m^{-2}, may provide a viable geothermal energy source. A temperature of 100°C, for example, is reached at less than 2 km depth. The prevailing thermal gradient is about 40°C km^{-1} or higher, suggesting that conditions are more inducive to the formation of natural gas rather than oil for older source material, but may accelerate oil formation at shallow depths if the source material is young (KLEMME, 1975).

SEISMIC INVESTIGATIONS

A number of seismic refraction profiles have been obtained in the Red Sea (DAVIES and TRAMONTINI, 1970; DRAKE and GIRDLER, 1964; FAIRHEAD, 1973; GINSBERG et al., 1981; GIRDLER, 1969; MAKRIS, 1982; PRODEHL, 1983; TRAMONTINI and DAVIES, 1969). However, their interpretation remains a controversy. Within the axial zone at a depth of about 5 km, velocities ranging from 6.6—7 km s^{-1} are encountered (Fig. 6). They are interpreted to represent the oceanic layer (or layer 3) so that here oceanic crust is implied. Over the Red Sea marginal trough there is a 2—5 km sedimentary cover, with velocities ranging from 3.5—4.5 km s^{-1}. The material comprising this cover is considered to be evaporites and clastic and volcanic material. Still unresolved is the nature of the material underlying this sediment cover. In a detailed survey area between 22° and 23°N to the east of the axial trough, measured velocities of 6.6 km s^{-1} at an average depth of 4.6 km have been interpreted as oceanic crust (DAVIES and TRAMONTINI, 1970; TRAMONTINI and DAVIES, 1969). Others argue that this may represent one of the many greenstone belts within the Precambrian (LOWELL et al., 1975). However, if we accept this latter interpretation, then the total amount of opening in the Red Sea would be relatively small compared to that in the Gulf of Aden. This would imply that major tectonic movements have occurred in the East African Rift System contradictory to geological evidence. In addition, as is mentioned earlier, the continental crust interpretation contradicts results of gravity modelling at least at 20°N (BROWN and GIRDLER, 1982). Around the margins of the Red Sea, velocities from 5.84—6.97 km s^{-1} have been encountered, and they are interpreted as representing the Precambrian basement, hence implying a continental crust.

Continuous seismic reflection profiles in the Red Sea reveal the existence of a very prominent reflector, Reflector S, throughout the Red Sea except in the axial rift (Fig. 7a; LOWELL and GENIK, 1972; LOWELL et al., 1975; PHILLIPS and ROSS, 1970; ROSS and SCHLEE, 1973; ROSS et al., 1969; SEARLE and ROSS, 1975). This reflector is interpreted as the top of the evaporite section commonly correlated with the Miocene-Pliocene boundary. Within the axial trough where active plate accretion has been taking place over the past 4—5 m.y., the evaporitic layer is absent except where flowage has occurred from the sides. Here reflector S is not observed. The dominant structural styles are normal faulting and salt diapirism, and they are often intimately associated with each other. For example, flowage in a piercement salt feature may be triggered by normal faulting from below (Fig. 7b). In the southern part of the Red Sea normal faulting has become so highly developed that a rift structure, a continuation of the Danakil Depression and the Gulf of Zula to the south, has been developed, so that here double rifting occurs (LOWELL et al., 1975). Borehole results suggest that the Miocene salt layer is widespread and reaches 5 km in places.

To summarize, the axial trough of the Red Sea is underlain by oceanic crust, the lateral extent of which however is still uncertain (GASS, 1977), although its extension almost to the two coasts appear likely. Over the axial trough, the marginal trough and the immediate continental regions, lithospheric thinning has taken place.

RED SEA HOT BRINES

Along the axial rift of the Red Sea, hot brines have accumulated in a number of deeps ranging from about 19° to 24°N (fig. 1). These deeps located in isolated axial troughs and separated by inter-trough zones, are almost devoid of sediments; they have larger positive Bouguer anomalies compared to the inter-trough zones, and they exhibit large lineated magnetic anomalies. Thus, they represent sites of active plate accretion (BERTIN et al., 1979). Because of the density (and hence acoustic impedance) contrast between seawater and brine, false echoes are received on echo-sounding records across hot brine areas (Fig. 8; ROSS et al., 1973). The brine itself has temperatures ranging from slightly above ambient to over 65°C, salinities of 100 to several hundred ppt, and an oxygen content of about zero (BAUMANN et al., 1973; BÄCKER and SCHOELL, 1972; EMERY et al., 1969). In contrast, the deep waters above the brine usually have temperatures around 22°C, 40.6 ppt salinity, and 2 ml l^{-1} oxygen. Chloride, sodium, potassium and calcium are enriched relative to ocean water, while sulfate, carbonate and magnesium are depleted (BREWER and SPENCER, 1969). Particularly noteworthy is the enrichment of various heavy metals. Lead, manganese, iron and zinc among others are all enriched well over 1000 times in comparison to seawater (BROOKS et al., 1969). Temperature measurements in the Atlantis II and Chain Deeps, particularly from the upper convection layer through the high temperature zone to the lower convection layer, suggest that water temperature variations may be cyclic (HUNT et al., 1967; ROSS, 1972; SCHOELL and HARTMANN, 1978). This implies that the intensity of discharge or brine temperature fluctuations may also be cyclic and that the recharge action of the brine may be episodic.

Sediments beneath the hot brine deeps can be divided into 7 facies types (BISCHOFF, 1969). Detrital material, which consists of coarse-grained sediments, remains of pelagic organisms plus detrital quartz, feldspar and clay is the typical deposit outside the hot brine basins. Within the brine basins, black fine-grained iron montmorillonite consisting of mostly clay with iron oxides, some sphalerite and a very high water content represents the uppermost sediment facies. Underlying this is a layer of amorphous goethite, fine to medium-grained and orange to yellow in color. The sulfide facies is a black, fine-grained sediment not found outside the basin. It contains the highest concentration of heavy metals that are present chiefly as sulfides, iron monosulfide, marmatite or sphalerite, chalcopyrite and pyrite. Within the iron montmorillonite and goethite amorphous facies are thin, semi-lithified, discontinuous beds of manganosiderite. In addition, massive, white, coarsely crystalline beds of anhydrite and black manganite also occur. Although the average metal content of these deposits is less spectacular than that of the brine, since very high concentrations are confined to the sulfide-rich facies (BÄCKER and RICHTER, 1973; BÄCKER, 1976; BIGNELL and ALI, 1973; DEGENS and ROSS, 1976; HENRICKS et al., 1969), these metalliferous deposits are of economic significance, and efforts have been expended towards commercial exploitation (BÄCKER, 1979).

The Red Sea hot brines certainly do not originate as a result of evaporation. This is evidenced by the differences in chemical composition between the brine waters and waters of the Dead Sea. Also in the Atlantis II Deep area, heavy oxygen and hydrogen isotopes are depleted in comparison to the waters directly overlying the brine. Other hot brine deeps are enriched in these isotopes, however.

Such an isotopic composition has lead to the speculation that the brines are derived from hydrothermal circulation associated with plate accretion (BURKE et al., 1981; SCHOELL and FABER, 1976, 1978; SCHOELL and RISCH, 1976). According to this view, along the immediate ridge flanks which are devoid of sediment cover and where the pillow lavas, pillow breccias, hyaloclastites and minor flows are highly porous (though decreasingly so with depth), sea water permeates into the seafloor (BECKER et al., 1982). Chemical reactions take place between the basaltic material and the circulating water, resulting in the filling of cracks by precipitated minerals and geochemical alteration of the two interacting components (ANDERSEN et al., 1982; EDMOND et al., 1982). These hydrothermal waters emerge eventually along the ridge crest, sometimes in the form of black smokers. If the ridge crest morphology permits them to accumulate, they form hot brine pools such as in the brine deeps of the Red Sea. Accordingly, for the Red Sea hot brines, these hydrothermal waters are paleo-waters of the Red Sea at various ages and reflect the isotopic composition at these ages. Figure 9 shows the change of O^{18} concentrations in foraminifera in the Red Sea relative to present values (DEUSER and DEGENS, 1969; DEUSER et al., 1976). Since temperature variations in Red Sea waters are small, the ΔO^{18} values [defined as δO^{18} (past)— δO^{18} (present)] represent approximately the variation in O^{18} in seawater. From this figure, one may speculate that the Atlantis II, Chain, Albatross and Discovery brines are "light" Red Sea waters from previous pluvial periods. The Valdivia Deep may contain present-day Red Sea deep water or Red Sea paleowater of present-day isotopic composition. The Erba, Port Sudan and Nereus deeps could represent Red Sea paleo-waters which have penetrated the upper crust during times when the isotopic composition of the Red Sea was significantly enriched in heavy isotopes compared to today. The brine waters of the Kebrit and Suakin deeps do not show oxygen and hydrogen isotopic compositions comparable to Red Sea paleo-waters. They may represent accumulations of interstitial waters whose isotopic ratios have been altered by low temperature mineral reactions (SCHOELL and FARBER, 1978).

The mid-water density boundary separating normal Red Sea water from that of the hot brine pools is located approximately at the depth where Reflector S outcrops. It has also been noted that heat flow along the coast may yield temperatures in the order of 100°C at a depth of less than 2 km. These two observations suggest that hydrothermal discharge occurs principally along the flanks of the brine pool rather than in the central region of the axial zone. Hydrostatic pressure at times of sea level rise may induce a pulsating flow of formation waters entrapped in the sediments overlying Reflector S, whereby halite is leached from the underlying evaporite sequence in accordance with mineral equilibria. In view of the fact that freshwaters formed during pluvial times may also enter the submarine aquifers, mixing of waters of different isotopic compositions may occur and could account for the observed range in stable isotope values from saline waters found in the various brine deeps. It is conceivable, however, that some of the heavy metals were leached from basaltic rock in the immediate vicinity of the brine pools or were supplied during active spreading phases, thus implying a secondary source for part of the solutes from deeper regions.

To summarize, the Red Sea brines could be paleowaters (or interstitial waters) which have circulated in the upper lithosphere as hydrothermal waters. Their discharge or recharge mechanism may be triggered by climatic changes and other exogenic factors (DEGENS and ROSS, 1970). This speculative origin is further supported by the helium isotopic composition of brine waters, which at least for the Atlantis II and Discovery deeps exhibit an enrichment in helium with the mantle helium isotopic signature (LUPTON et al., 1977).

THE GULF OF SUEZ AND THE LEVANT SHEAR SYSTEM

The Gulf of Suez is a graben with a long history of subsidence since the late Paleozoic (SAID, 1962). It is underlain by continental crust, and the present tectonics may be one of right lateral shear with normal faulting.

The Gulf of Elat (Aqaba) is a transform boundary with a slight component of opening. It is about 160 km long and 15—10 km wide. Along the shores of the Gulf, coastal plains are either very narrow or are absent. On the western side, large alluvial fans extend into submarine cones incised by canyons on sloping terraces. The eastern shores descend abruptly to deep basins. Three such deep basins have been mapped in the center part of the Gulf (BEN-AVRAHAM et al., 1979a,b). They are elongated, en échelon, and reach a maximum depth of 1830 m. Structure within the Gulf is controlled by normal faulting. This is most intense in the basins and along the boundary faults. Diapirs have been observed in the southern part of the Gulf. A thick sedimentary fill consists mainly of alluvium on the margins and turbidites and pelagic sediments in the basins. Seismic and gravity data suggest that the oldest sediments must lie at least 7 km below the present level of adjacent land (BEN-AVRAHAM et al., 1979a).

The part of the Levant Shear System joining the Gulf of Aqaba to the Dead Sea is very narrow. This area, known as the Araba Depression, is marked by well-documented left-lateral strike-slip motion on either side (FREUND et al., 1970). To the north, the shear system widens to accommodate the Dead Sea and continues northward along the Jordan Valley into the lakes of Tiberias and Hula. North of Hula the depression reorients itself NNE, to pass between the Lebanon and Anti-Lebanon Mountains. At the latitude of Tripoli, the shear system changes to an almost N-S strike, cutting through the basalt of Homs into the Er Rharb Depression. Where the Levant Shear System ends to the north is not clear. QUENNEL (1958) and FREUND et al. (1970) have presented convincing evidence that about 105 km of left-lateral movement has occurred along the Levant Shear; 65 km of this in the Miocene, the remaining 40 km in the Plio-Pleistocene. Of this 105 km, only some 60 km movement are taken up by faults in the Gulf of Elat (EYAL et al., 1981). This strike-slip motion must be taken into consideration in any plate tectonic reconstruction of the area.

THE GULF OF ADEN

The Gulf of Aden links the East African Rift Valley System and the Red Sea to the Carlsberg Ridge in the northwest Indian Ocean, which in turn is part of the worldwide mid-ocean ridge system. The axis of the Carlsberg Ridge is offset 310 km right-laterally along the Owen Fracture Zone at about 10°N, 57°E, and continues into the Gulf of Aden as the Sheba Ridge (Fig. 10). Within the Gulf of Aden, this ridge is offset by a number of transform faults, the most significant of which occurs at 52°E along the Alula-Fartak Trench. Minor offsets have been mapped both magnetically and bathymetrically to the west until the Gulf of Tadjura which marks the western end of the Gulf of Aden. The central rift of Sheba Ridge is characterized by high heat flow (VON HERZEN, 1963) and shallow seismicity (FAIRHEAD and GIRDLER, 1970).

The crustal structure beneath the Gulf of Aden is clearly oceanic. This is demonstrated by seismic refraction results (LAUGHTON and TRAMONTINI, 1970; LAUGHTON et al., 1970: FAIRHEAD, 1973) which show that the measured velocities may be grouped into (Fig. 10):

1) an unconsolidated sedimentary layer (1.83—3.08 km s^{-1}),
2) a layer of volcanic or consolidated sedimentary material (3.94—5.3 km s^{-1}), and
3) the oceanic layer or layer 3 (6.15—6.96 km s^{-1}), below which the upper mantle has a velocity ranging from 7.55—8.45 km s^{-1}.

Anomalously low velocity upper mantle has been found on the western end of the axis of Sheba Ridge (7.06—7.14 km s^{-1}). In general, the crustal thickness is about 12 km, inclusive of the water layer. This, together with the presence of layer 3, strongly suggests that the crust beneath the Gulf of Aden is oceanic. Since profiles within 55 km of the south and north coasts suggest oceanic crusts, the minimum crustal separation within the Gulf must be 260 km.

Gravity data in the western Gulf of Aden also suggest that almost the entire seafloor is underlain by oceanic crust (GIRDLER et al., 1980). The crustal thickness is only about 5 km along the current spreading axis, where a discontinuous dramatic shallowing of the Curie isotherm possibly occurs (TAMSETT and GIRDLER, 1982). The lateral extent of the intrusion zone reaches about 75 km. It overlies anomalous mantle at a depth of just over 10 km (GIRDLER et al., 1980).

Just as in the Red Sea, lineated magnetic anomalies have been mapped in the Gulf of Aden (ALLEN, 1964; COCHRAN, 1981; GIRDLER and STYLES, 1978; GIRDLER et al., 1980; NOY, 1978). However, their interpretation is not unambiguous, and their implications on the tectonic evolution of the Gulf are still debatable (GIRDLER and STYLES, 1982; COCHRAN, 1982a). According to GIRDLER et al. (1980) and GIRDLER and STYLES (1978), the very large amplitude anomalies over the ridge axis is a result of the recent phase of seafloor spreading while the lower amplitudes over the marginal troughs are the result of an earlier phase of spreading. From magnetic data in the western part of the Gulf, GIRDLER and STYLES (1978), GIRDLER et al. (1980) and NOY (1978) deduced the seafloor spreading history in terms of a two-, possibly a three-stage model.

Since no major shear, spreading, or subduction occurs between Nubia and Somalia, they argued that the evolutionary history of the Gulf of Aden must be similar to that of the Red Sea. Hence they assumed a recent phase of spreading from 4.5 m.y. ago to the present at a rate of 1 cm yr^{-1} and an earlier phase from the upper Oligocene (23.5 m.y. B.P.) to mid-Miocene (about 16 m.y. ago) (Fig. 3c). Rates north and south of the Ridge were 1.1 and 1 cm yr^{-1} respectively. An even earlier phase from upper Eocene (43 m.y.) to lower Oligocene (35.5 m.y.) may also be interpreted. With this model, GIRDLER et al. (1980) constructed an isochron map for the western Gulf of Aden (Fig. 11). Here oceanic crust extends across almost the entire width, with an age of about 40 m.y. off the Somalian and the Arabian coasts. The Oligocene-Miocene spreading phase took place along a ridge axis lying in the south (thick line). A period of quiescence followed, lasting about 10 m.y., during which about 1.5 km of sediment were deposited. The Plio-Pleistocene phase of the seafloor spreading occurred along an axis which is displaced 75 km to the north in the eastern part of the western Gulf. This axis has a strike of N 120°E. It is offset by a number of transform faults and bifurcates in the west, one branch going towards the Strait of Bab-el-Mandeb and the other towards the Gulf of Tadjura. The bifurcation point may mark the triple junction between the Arabian, Somalian and North Danakil plates.

To emphasize the difficulty encountered in unraveling the tectonic evolution of the Gulf of Aden, we should point out that COCHRAN, using the same data as GIRDLER et al. (COCHRAN, 1982a) as well as other independent data (COCHRAN, 1981) concluded that the Gulf (and hence, by inference, the Red Sea) has undergone only one continuous episode of seafloor spreading, beginning about 15 m.y. ago (Fig. 3d). The spreading, at least in the central Gulf of Aden, appears to be asymmetric. Noting the existence of a magnetic quiet zone landwards of the recognizable anomalies and that the seaward boundary of this zone is marked by a noticeable increase in basement depth, COCHRAN argued that opening of the Gulf of Aden and the Red Sea started with diffuse extension characterized by dike injection and normal faulting within a wide "rift zone". This extension amounts to about 100 km over at least 10 m.y. and is responsible for the magnetic quiet zone (COCHRAN, 1982b). The diffuse extension gave way eventually to spreading at a well-defined ridge axis, whereby magnetic lineations could be generated.

PLATE TECTONICS OF THE RED SEA AND GULF OF ADEN AREAS

Although the Red Sea has long been considered a classic example of an ocean in its first stage of evolution, its tectonic history and the regional kinematic plate motions still remain an enigma. McKENZIE et al. (1970), FREUND (1970), GRIDLER and DARRACOTT(1972), LE PICHON et al. (1973) and RICHARDSON and HARRISON (1976) have developed kinematic patterns based on the assumption that the coastline or a particular isobath represents the continent-ocean structural interface. This implies that a large part of the Red Sea consists of oceanic crust. Others (ROSS and SCHLEE, 1973; LOWELL and GENIK, 1972; LOWELL et al., 1975; HUTCHINSON and ENGELS, 1972) assumed that oceanic crust is confined to the axial trough so that most of the Red Sea and the Afar Depression is underlain by thinned continental curst. Which of the above views is correct is still a subject of debate. In the following discussions we shall make use of the arguments of LE PICHON and FRANCHETEAU (1978).

To establish the kinematic plate pattern in the Red Sea — Gulf of Aden — Levant area, seafloor spreading is assumed to be stable. This is reasonable because in the early stage of opening adjustments in the kinematic pattern are mechanically difficult and are unlikely as long as the two continent-bearing lithospheres are in contact. The fact that transform faults especially in the Gulf of Aden do not indicate a change in trend implies that the poles of rotation have remained much the same throughout time, i.e., the total poles of opening are approximately the same as the present instantaneous poles.

To determine the instantaneous pole between Arabia and Nubia one can use the detailed bathymetric survey of the Valdivia and the resulting transform fault pattern between 19° and 23°N (PLAUMANN, 1975). Since this pattern is consistent with the pole at 36.5°N, 18°E determined by McKENZIE et al. (1970) from a morphological fit of the two Red Sea coasts, one can accept this as the correct instantaneous as well as total pole (Table 1). Note that this does not imply that the Red Sea coasts should be fitted together sometime in the geological past.

South of 19°N, the McKENZIE pole no longer applies. The magnetic survey of ROESER (1975, 1976) suggests that the pole for the southern Red Sea must lie to the south. The best fit to his isochrons gives a pole position of 7°N, 50.5°E, with a rate of opening for the past 2 m.y. of 5×10^{-7} deg yr^{-1}. The corresponding half-spreading rate decreases from 0.75 cm yr^{-1} at 19°N to less than 0.6 cm yr^{-1} south of 16°N.

The pole just stated for the southern Red Sea has the important implication in that it predicts a different motion at 19°N from the Nubia-Arabia pole. We are therefore forced to conclude that a plate exists between Arabia and Nubia in the southern Red Sea, which as been named the North Danakil Plate, separated from Nubia by a zone of extension (Fig. 12). This extensional boundary is assumed to run along the western rift (discussed earlier from seismic data) through the Gulf of Zula to the Danakil Depression. As morphology suggests that the motion between the North Danakil and Nubian plates become negligible near 19°N since the zone of irregular topography terminates to the north near 19°N, the rate of opening between Nubia and Arabia may be determined. This is 3.2×10^{-7} deg yr^{-1}, which corresponds to a spreading rate of 0.7 cm yr^{-1} near 22°N and 0.5 cm yr^{-1} at the northern extremity of the Red Sea.

Since the orientation and motion along the Levant Shear System precludes it from being the transform boundary between Nubia and Arabia, a separate plate, the Sinai Plate, must be assumed. The Sinai-Arabia pole may then be determined by assuming that the Levant Shear between the Lebanon and Anti-Lebanon Mountains and the Araba Depression lies on the same line of pure slip. This pole can thereby be estimated to be at 32.5°N, 4.4°W. The predicted motion along the Levant Shear is then as follows: pure strike-slip in the Gulf of Aqaba and the Araba Depression, strike-slip with extension in the Dead Sea and Jordan River area, compression with minor slip in Lebanon, pure strike-slip in the Homs region, and strike-slip with minor extension in Er-Rharb. The Arabia-Sinai rate of motion is then determined by minimizing the motion between these two plates.

The Sinai-Nubia pole obtained by vector composition can be determined to be at 35.9°N, 42.6°E, with a rate of 1.65×10^{-7} deg yr^{-1}. A total rate of opening of 0.3 cm yr^{-1} in the Gulf of Suez in a WNW-ESE direction then results.

The Arabia-Somalia pole (26.5°N, 21.5°E) can be determined from the magnetic anomalies in the Gulf of Aden. The angular opening rate is 4×10^{-7} deg yr^{-1}. The remaining Somalia-Nubia and North Danakil-Nubia poles can then be computed by vector addition.

The total motion since early Miocene based on this kinematic pattern (Fig. 12) agrees reasonably well with the available data. Of particular interest is that the extension rates across continental grabens (e.g. in the Ethiopian Rift), the western Red Sea Rift and in the Gulf of Suez is of the order of 0.3 cm yr^{-1}. At this slow rate, features commonly associated with seafloor spreading are not observed.

REMARKS ON THE STRUCTURAL EVOLUTION OF THE RED SEA, GULF OF ADEN, AND EAST AFRICAN RIFT SYSTEM

Figure 13 summarizes the structural evolution in the Red Sea, Gulf of Aden and the East African Rift. Both the Red Sea and the Gulf of Aden have been undergoing more-or-less synchronous seafloor spreading in one or more phases. In contrast, the East African Rift does not appear to be characterized by active plate accretion, or else the accretion rate is so small that features common to active spreading centers are not observable at present. Instead, intensive normal faulting, uplift and alkaline volcanism are the basic structural styles.

Based on the idea of CLOOS (1939), we have proposed a sequence of events leading to seafloor spreading (DEGENS et al., 1971; WONG and VON HERZEN, 1974). These events are:

1) uplift or arching caused by thermally activated volume increase resulting from convective upwelling in the asthenosphere (Fig. 14 I);
2) block faulting and the propagation of normal faults from the top of the lithosphere downwards (Fig. 14 II);
3) igneous and hydrothermal activity (regulated by the availability of meteoric water) and subsequent erosion and non-marine deposition (Fig. 14 III);
4) lithospheric thinning, possibly by divergent convective flow in the asthenosphere, perhaps accompanied by "diffuse extension" (COCHRAN, 1981, 1982b);

5) continued thinning and rifting causing subsidence (McKENZIE, 1978), horst and graben formation, and the tilting of blocks. This leads to an extensive marine incursion and a thick sedimentary sequence; and

6) seafloor spreading with basaltic intrusions and hydrothermal convection (Fig. 14 IV).

It has been suggested that in the early stages of continental rupture a three-arm-system often develops in a more or less symmetric pattern (BURKE and DEWEY, 1973; BURKE, 1977, 1978). In the normal course of events, two of these arms develop while the third arm "fails". Tectonic activity in the failed arm comes to a stop after a period of active development. It could have reached the seafloor spreading stage and then the spreading axis died out (e.g. in the Benui Trough in Nigeria) or it may not even have reached the spreading stage (as has been postulated for the East African Rift System). In East Africa, a number of triple junctions have been recognized. The Red Sea, Gulf of Aden and the East African Rift constitute one such junction. Sedimentation in the failed arm is particularly rapid so that conditions are inducive to mineralization and fluid hydrocarbon accumulation (ROBBINS, 1983). In the Benui Trough mineralization is in the form of lead, zinc and copper sulphides in the cross-joints of anticlines formed in relation to the Benui ocean closure. The oil fields there are well-known.

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Table 1. Instantaneous poles and rates of rotation between plates

Plate pair	Pole of rotation Latitude	Longitude	Angular rotation rate (x 10^{-7} deg/yr)	References
Arabia/Nubia	36.5°N	18°E	3.2	McKENZIE et al., 1970
N. Danakil/Arabia	7°N	50.5°E	5.0	LE PICHON and FRANCHETEAU, 1978
Sinai/Arabia	32.5°N	4.4°W	-1.74	LE PICHON and FRANCHETEAU, 1978
Sinai/Nubia	35.9°N	42.6°E	1.65	
Arabia/Somalia	26.5°N	21.5°E	4.0	
Somalia/Nubia	6.6°N	149.7°W	-1.03	
N. Danakil/Nubia	19.1°N	39.5°E	7.7	

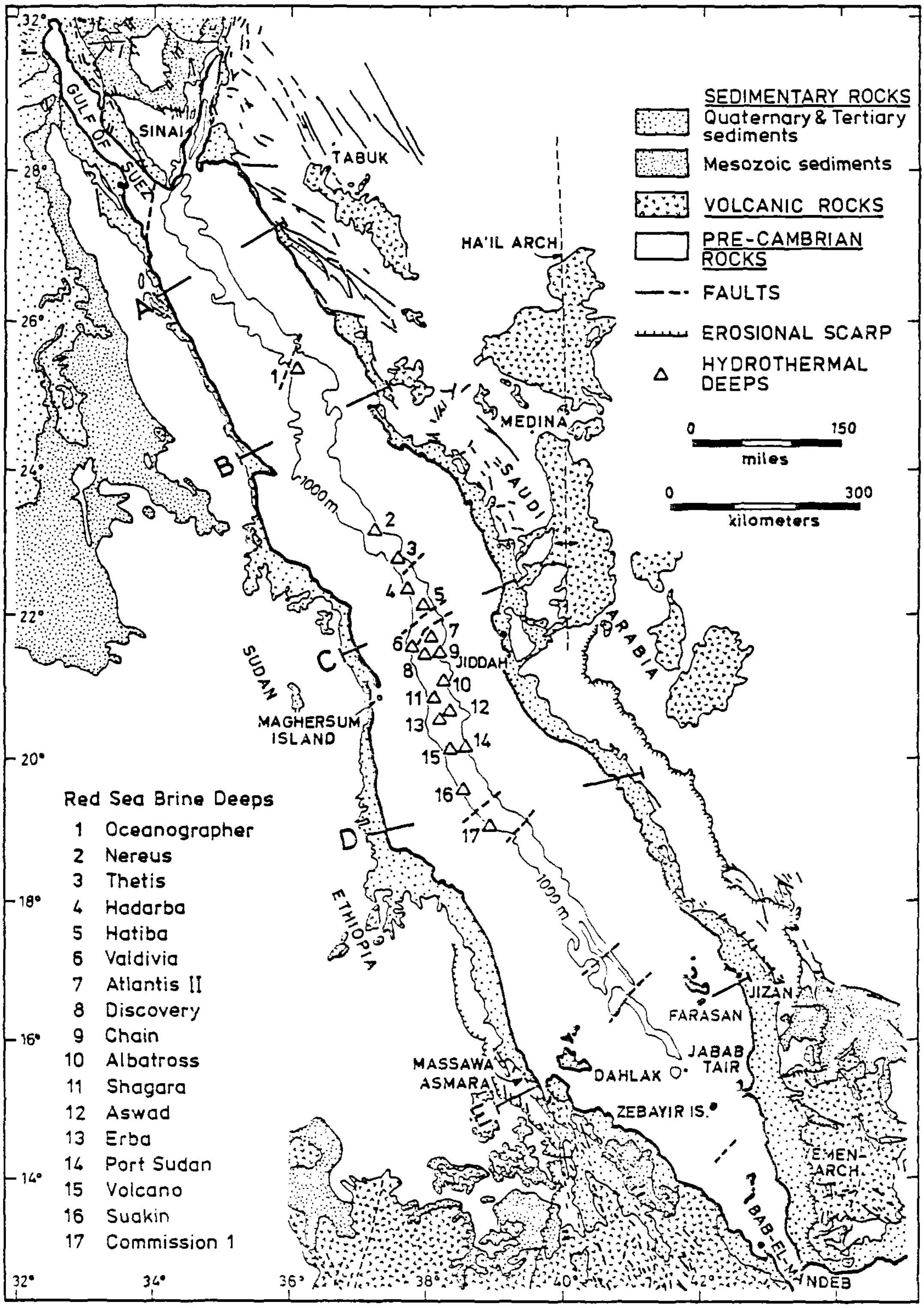

Figure 1. Simplified morphology and geology of the Red Sea area, showing locations of the hot brine deeps, and the 5 schematic transverse profiles of Fig. 2. Data from BÄCKER *et al.* (1975).

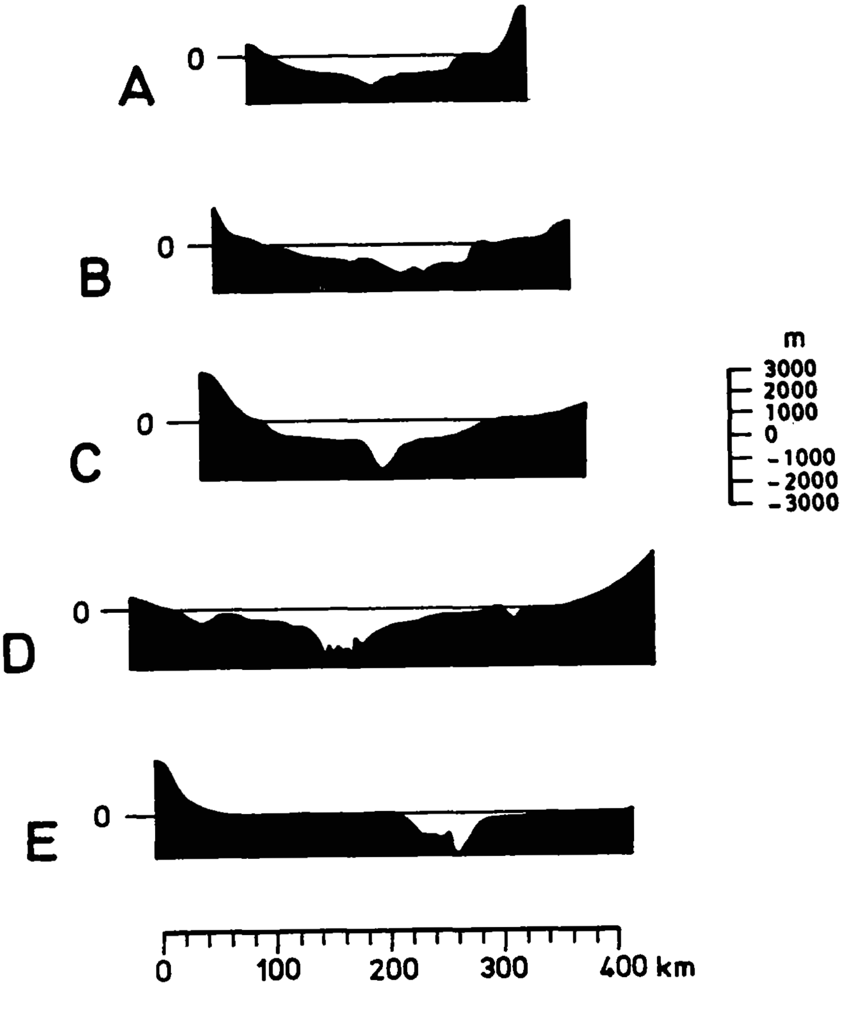

Figure 2. Schematic transverse topographic profiles across the Red Sea, showing the main trough, the axial trough, and the transition from shelf to coastal plain. After ALLEN (1970). Location of profiles shown in Fig. 1.

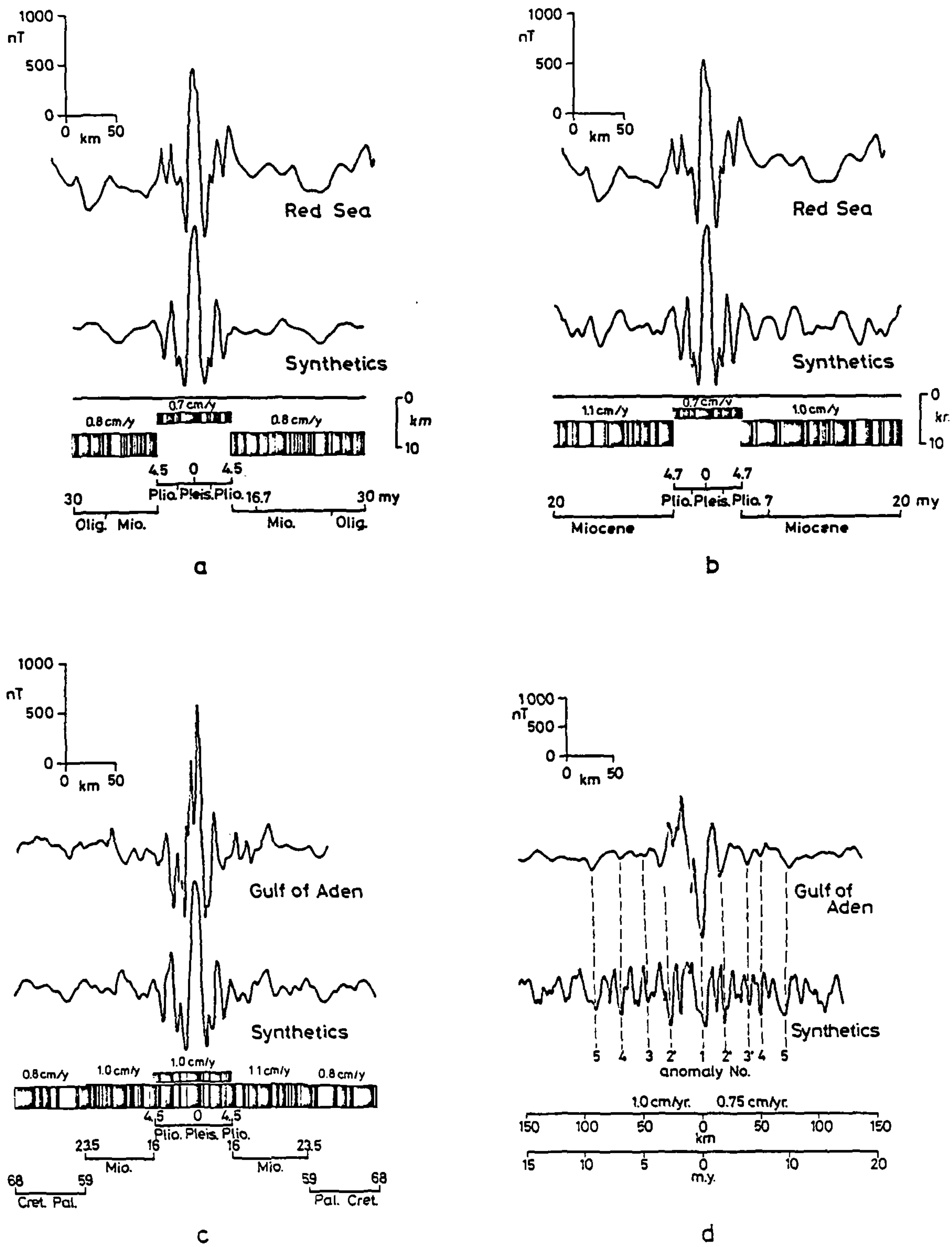

Figure 3. a) Composite magnetic profile of the central Red Sea reduced to the pole and its interpretation in terms of a two-stage spreading model,

b) Alternative model for the interpretation of the profile in a),

c) Magnetic profile in the western Gulf of Aden reduced to the pole and its interpretation in terms of two-stage spreading model similar to a),

d) Alternative model for the interpretation of the profile in c) (see text). Magnetic profiles not reduced to the pole.

After GIRDLER (1978), NOY (1978), GIRDLER *et al.* (1980) and COCHRAN (1981).

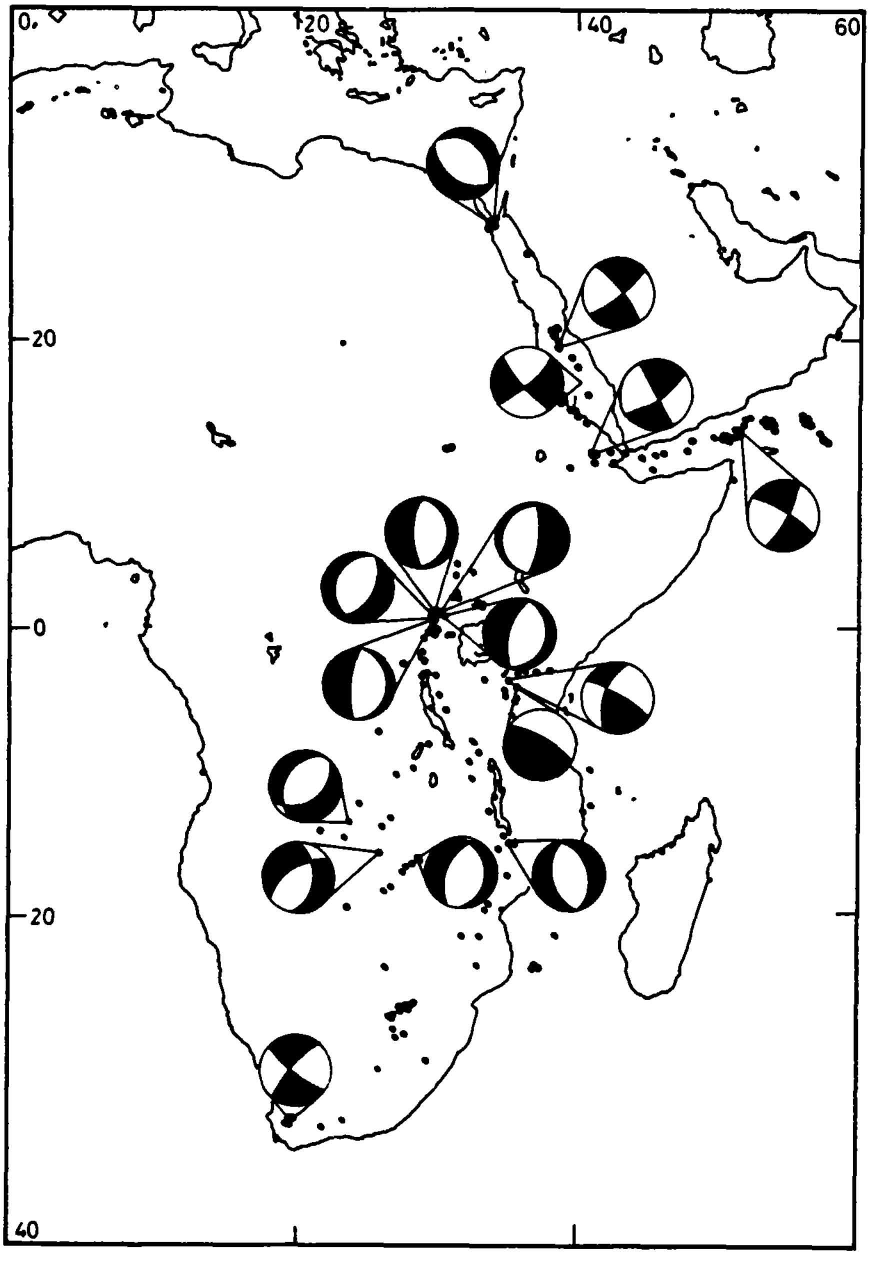

Figure 4. Seismicity and fault plane solutions of the Red Sea, Gulf of Aden and East Africa. Plots are equal-area projections of the lower hemisphere of the focal sphere. Darkened areas represent compressions, and blank represents dilatation.

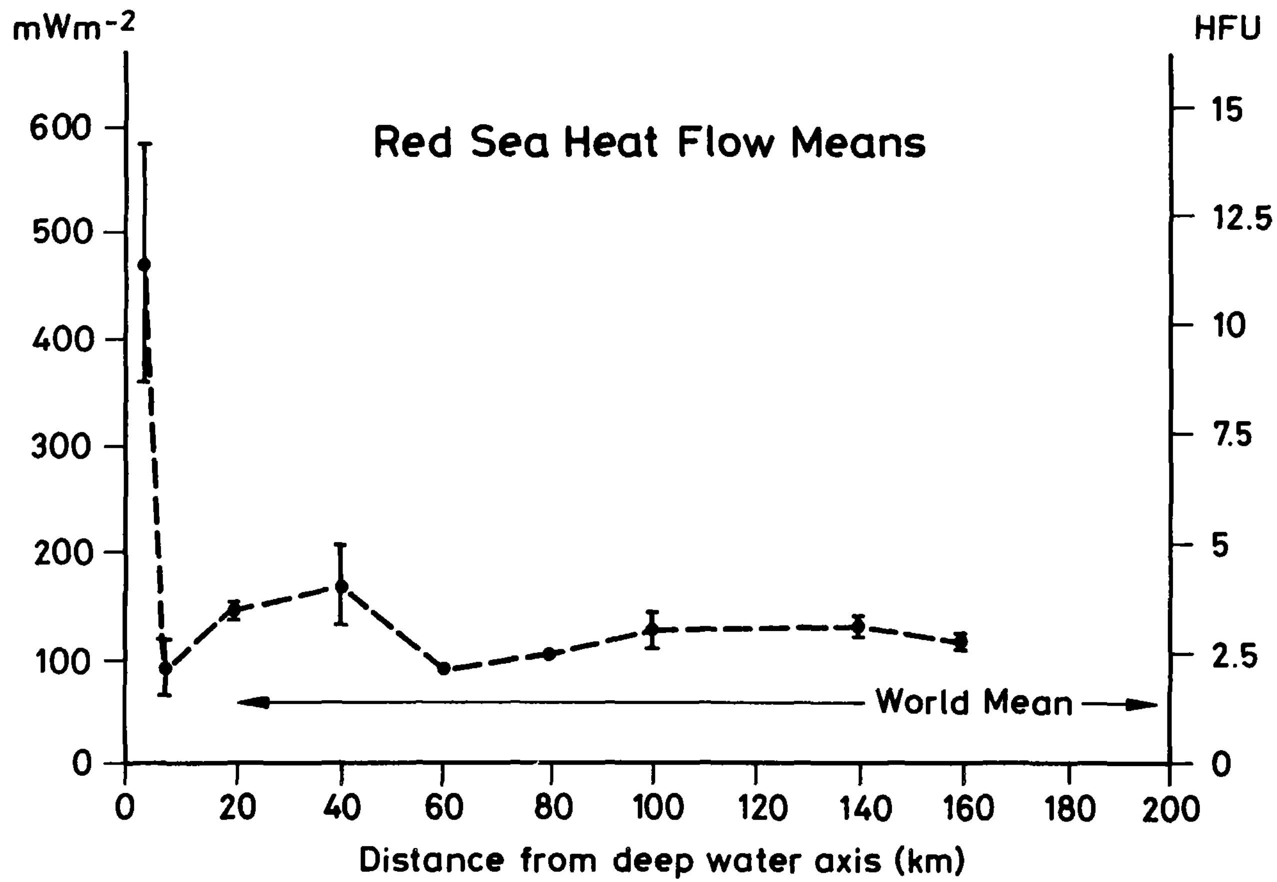

Figure 5. Mean values of heat flow measurements in the Red Sea as a function of distance from the deep water axis. Error bars indicate standard deviations. After GIRDLER and EVANS (1977).

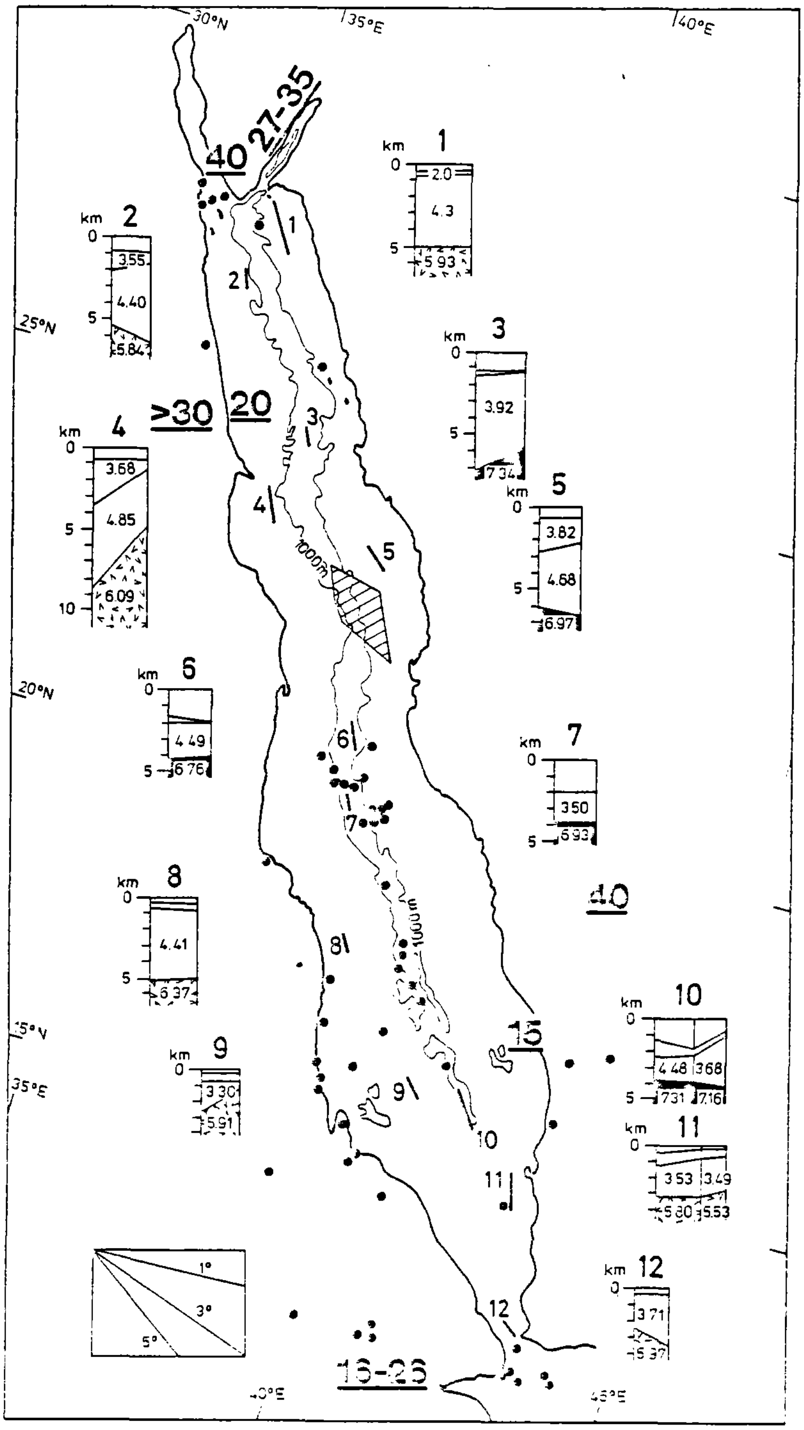

Figure 6. Summary of seismic refraction profiles in the Red Sea. Velocities are in km s^{-1}. Shaded area marks region of detailed seismic refraction survey. Earthquake epicenters are shown by black dots, and 1000 m isobath as a continuous line. Crustal thickness in km is given as large, underlined numbers.

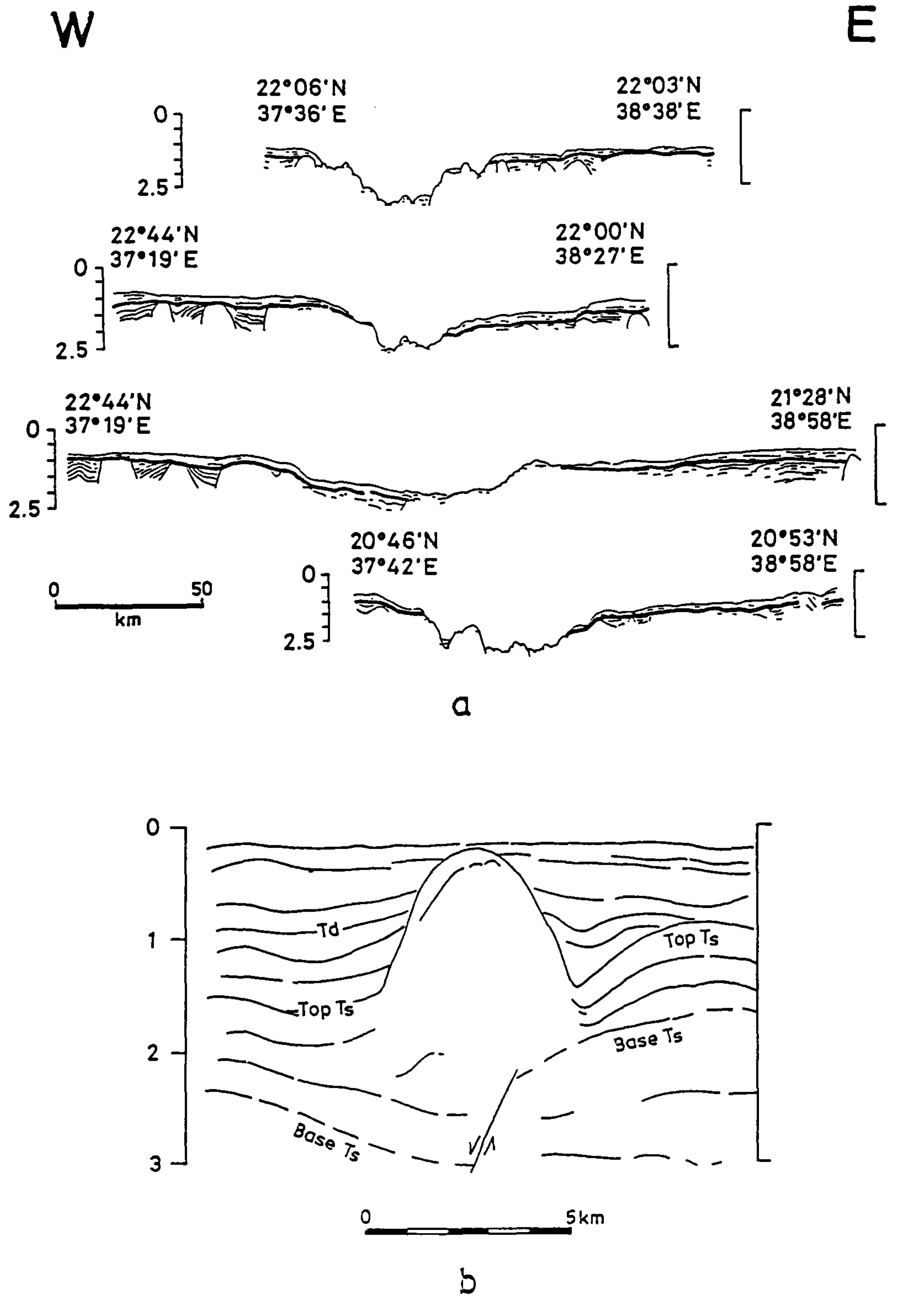

Figure 7. a) Seismic reflection profiles over the central Red Sea transverse to the axial trough. Reflector S is indicated by a heavy line. After SEARLE and ROSS (1975). Vertical scale is in seconds of two-way travel time. b) Seismic reflection profile with a 2:1 vertical exaggeration showing a piercement dome of Miocene salt possibly triggered by normal faulting from below. T_s represents Miocene salt, and T_d the Mio-Pliocene series of clastics and coral reefs, bounded by an unconformity at the base. After LOWEL et al. (1975).

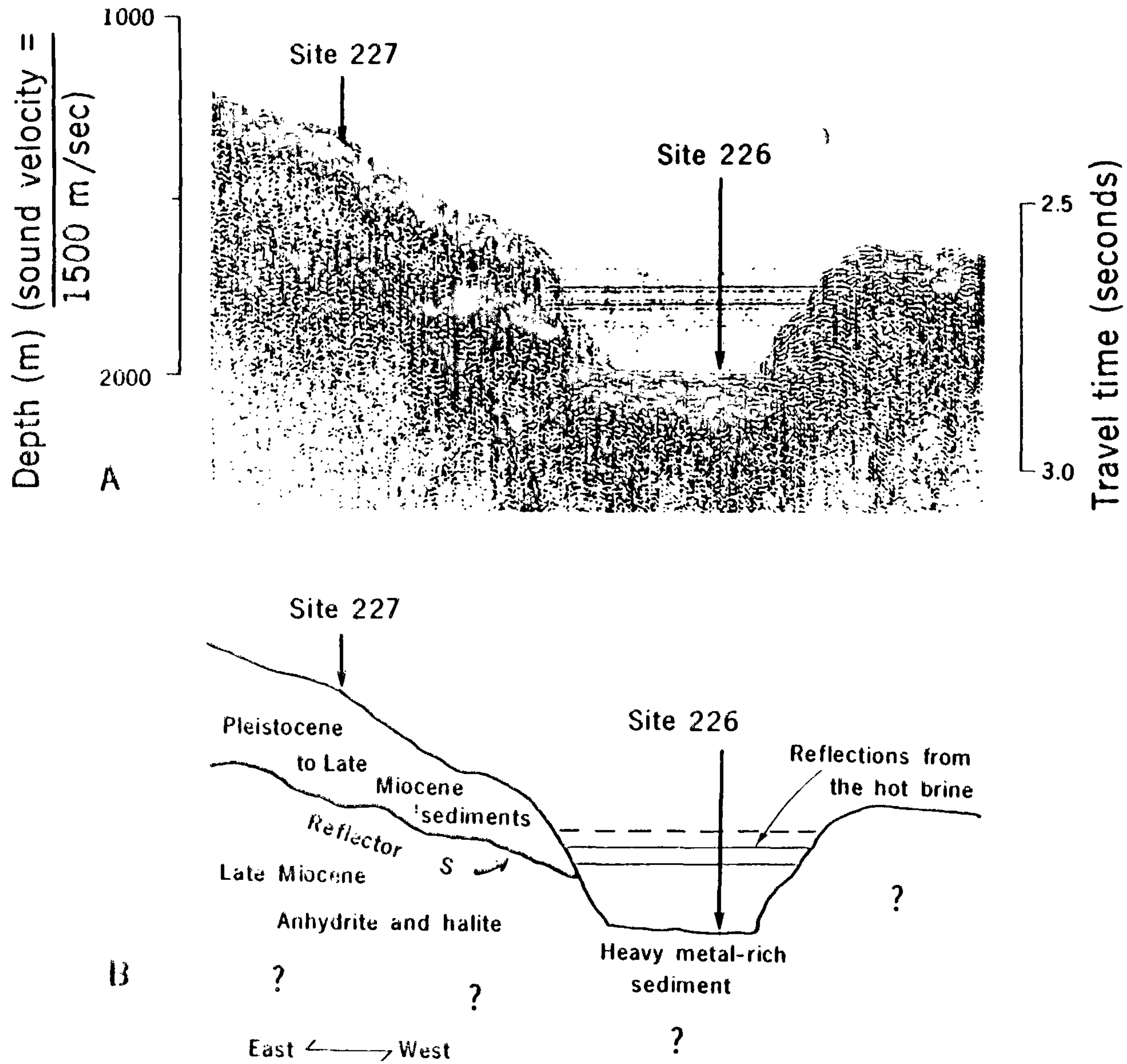

Figure 8. Echo-sounding record showing acoustic echo between hot brine water and Red Sea bottom water. After ROSS *et al*. (1973).

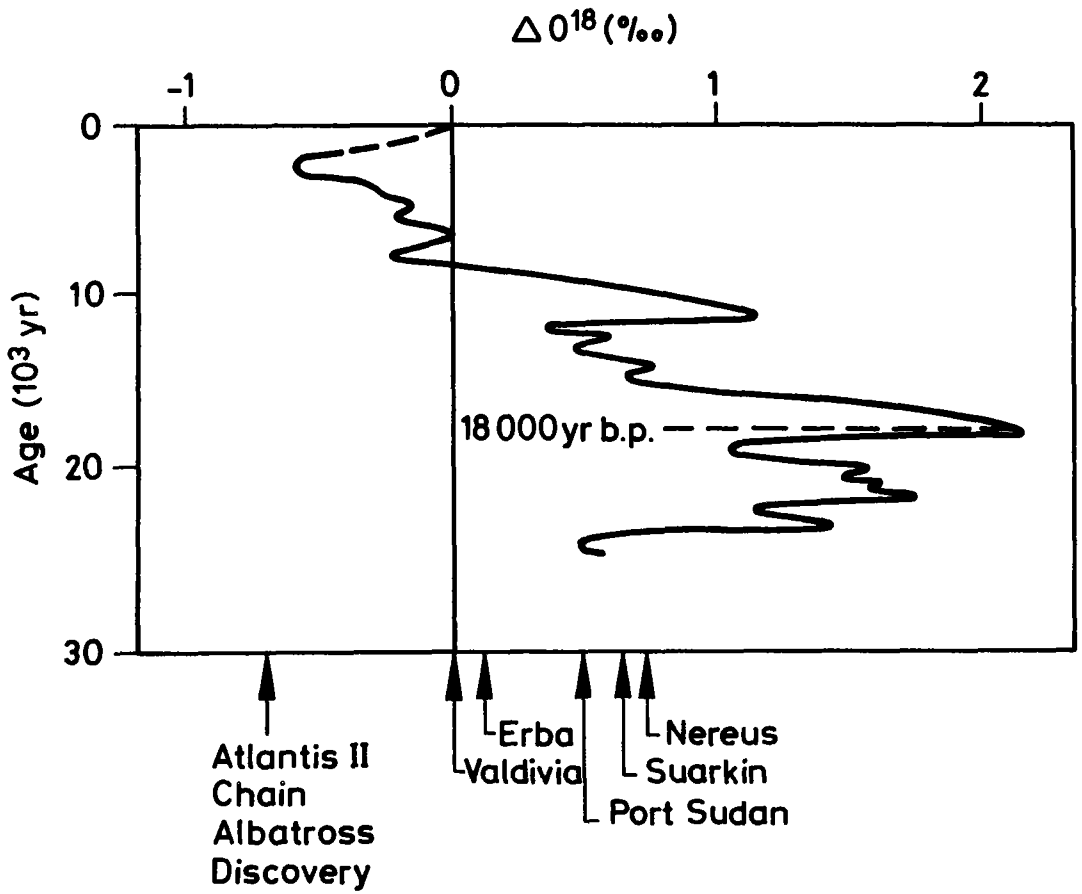

Figure 9. Change of 0^{18} concentrations in foraminifera in the Red Sea relative to present day values as a function of age (0^{18} = 0^{18} (past) - 0^{18} (present). 0^{18} values for the various brine deeps are also shown.

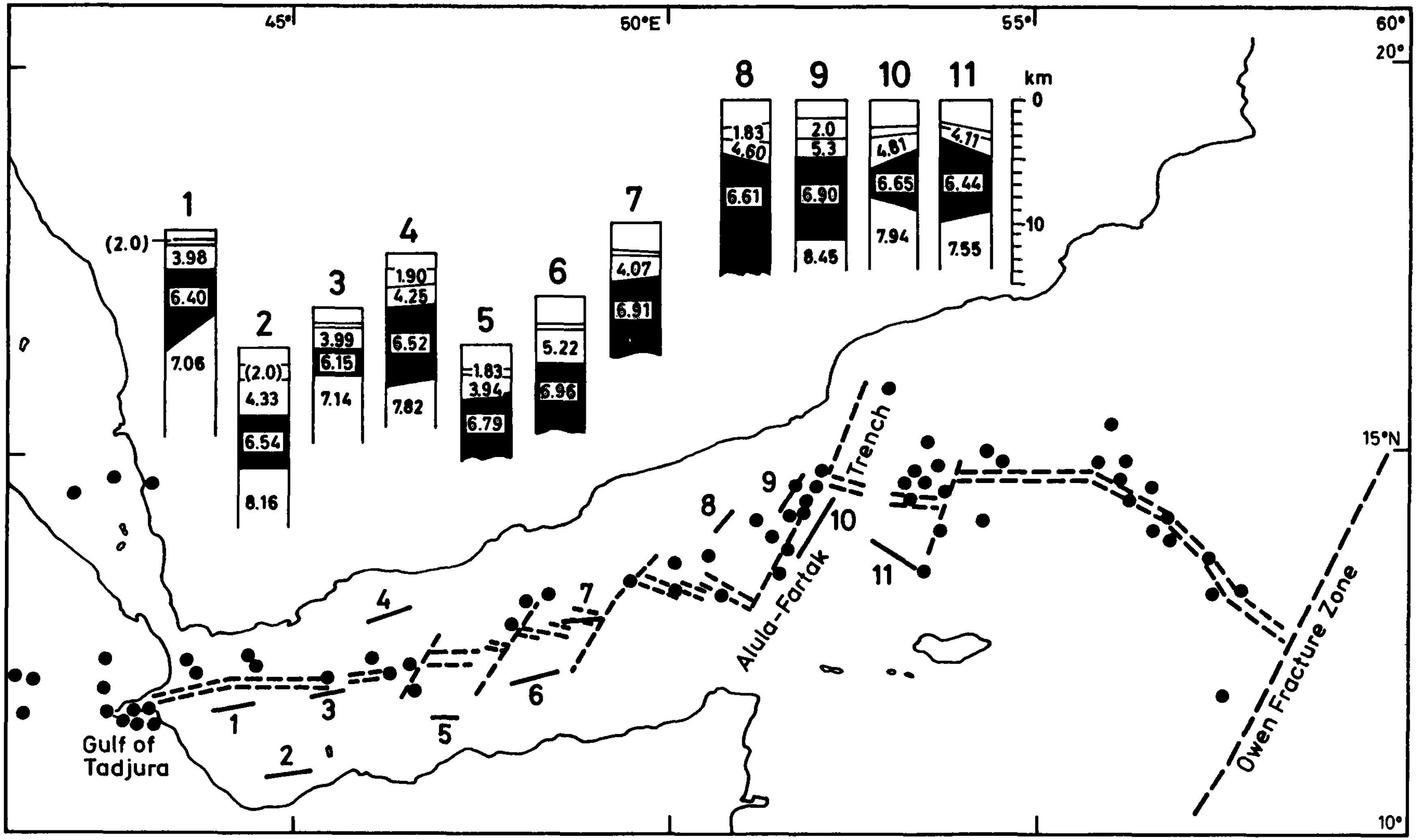

Figure 10. Summary of seismic refraction profiles in the Gulf of Aden; velocities are in km s^{-1}. Black dots represent earthquake epicenters, single-dashed lines indicate fracture zones, and double-dashed lines represent the Sheba Ridge.

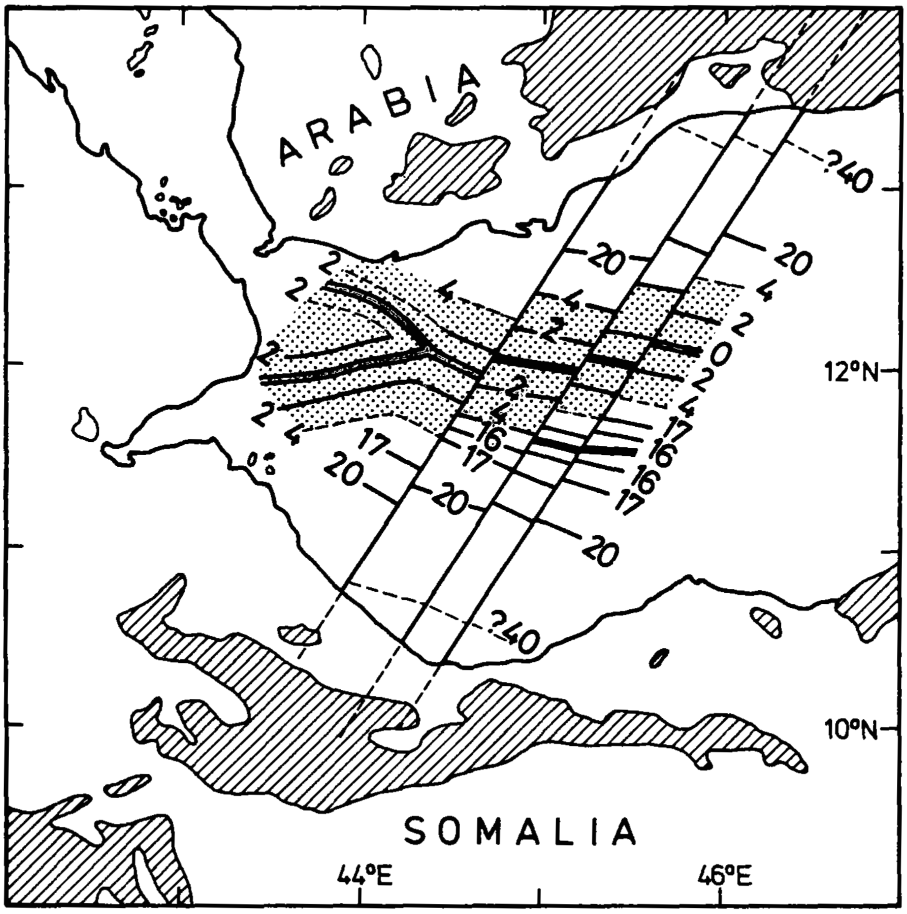

Figure 11. Isochrons in m.y. deduced from magnetic anomalies in the western part of the Gulf of Aden. The present spreading axis is shown by double lines, and the former spreading axis 75 km to the south is represented by a thick line. Three postulated transform faults are shown schematically. Shaded areas give outcrops of Precambrian age. The dotted part of the Gulf is younger than 4 m.y. After GIRDLER et al. (1980).

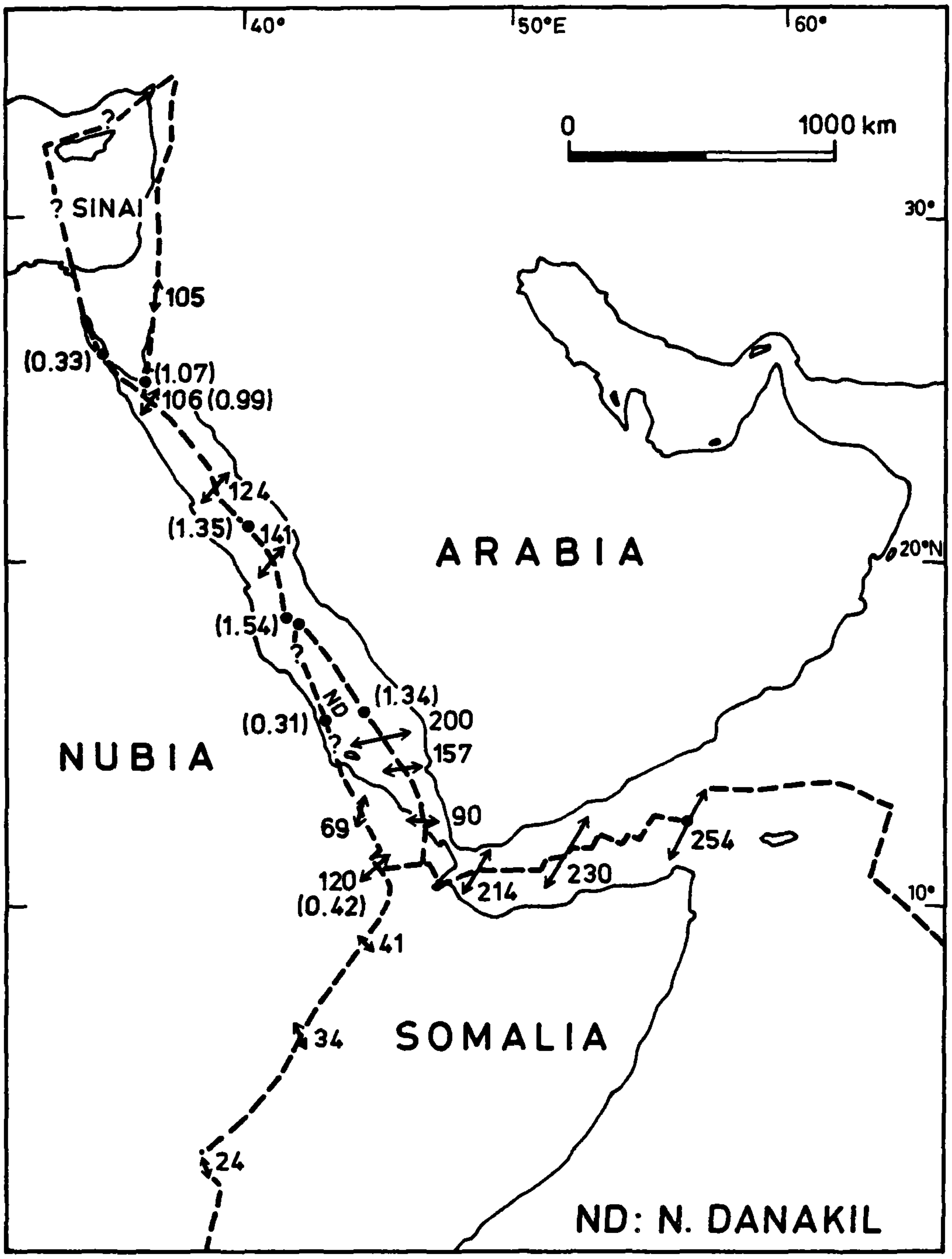

Figure 12. Kinematic pattern in the Red Sea, Gulf of Aden and East African rift area. Dashed lines represent plate boundaries. Arrows give the total motion (in km) since the early Miocene. Numbers in brackets are total opening rates in cm/yr. After LEPICHON and FRANCHETEAU (1978).

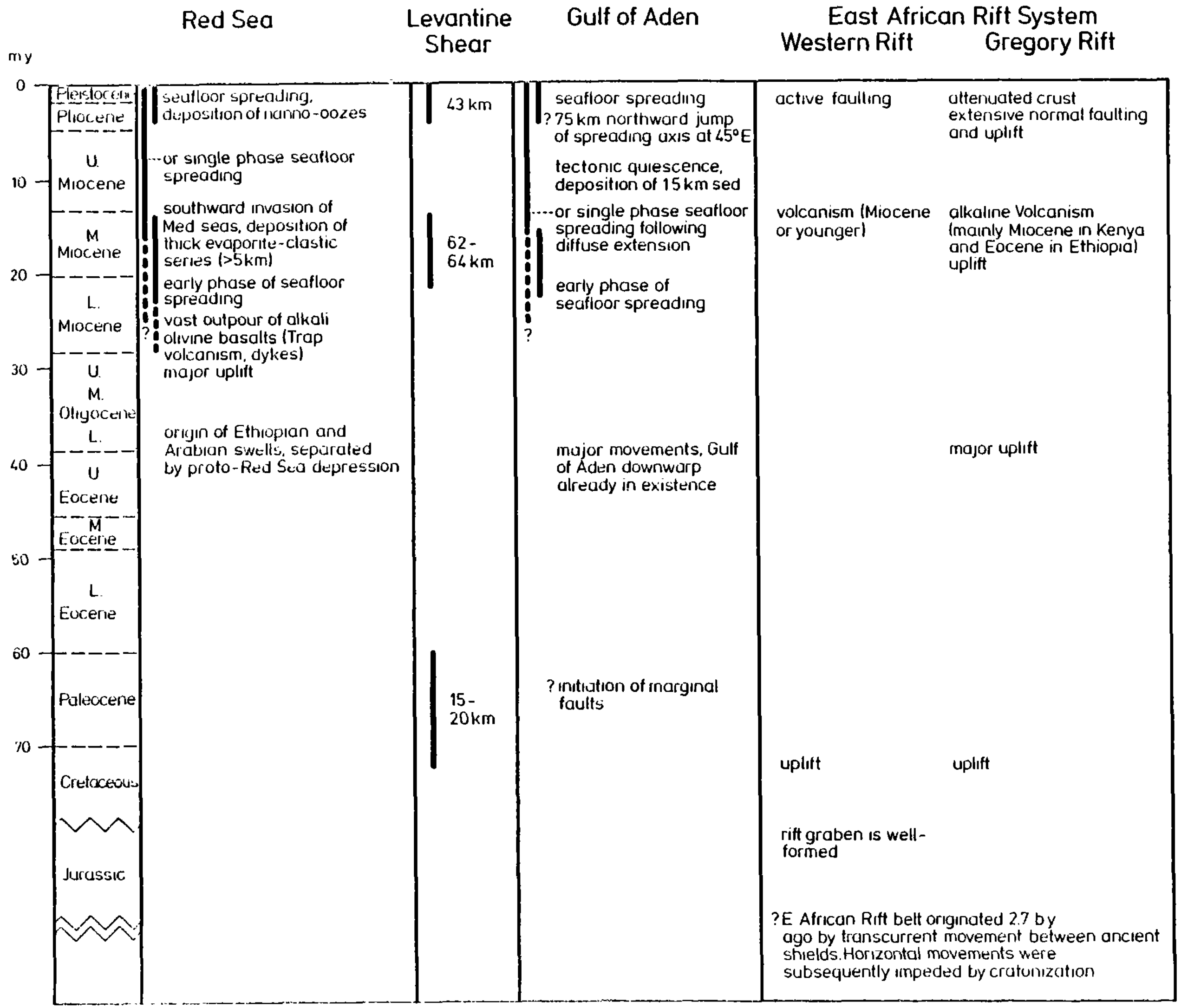

Figure 13. Summary of the structural evolution in the Red Sea, Gulf of Aden and the East African Rift.

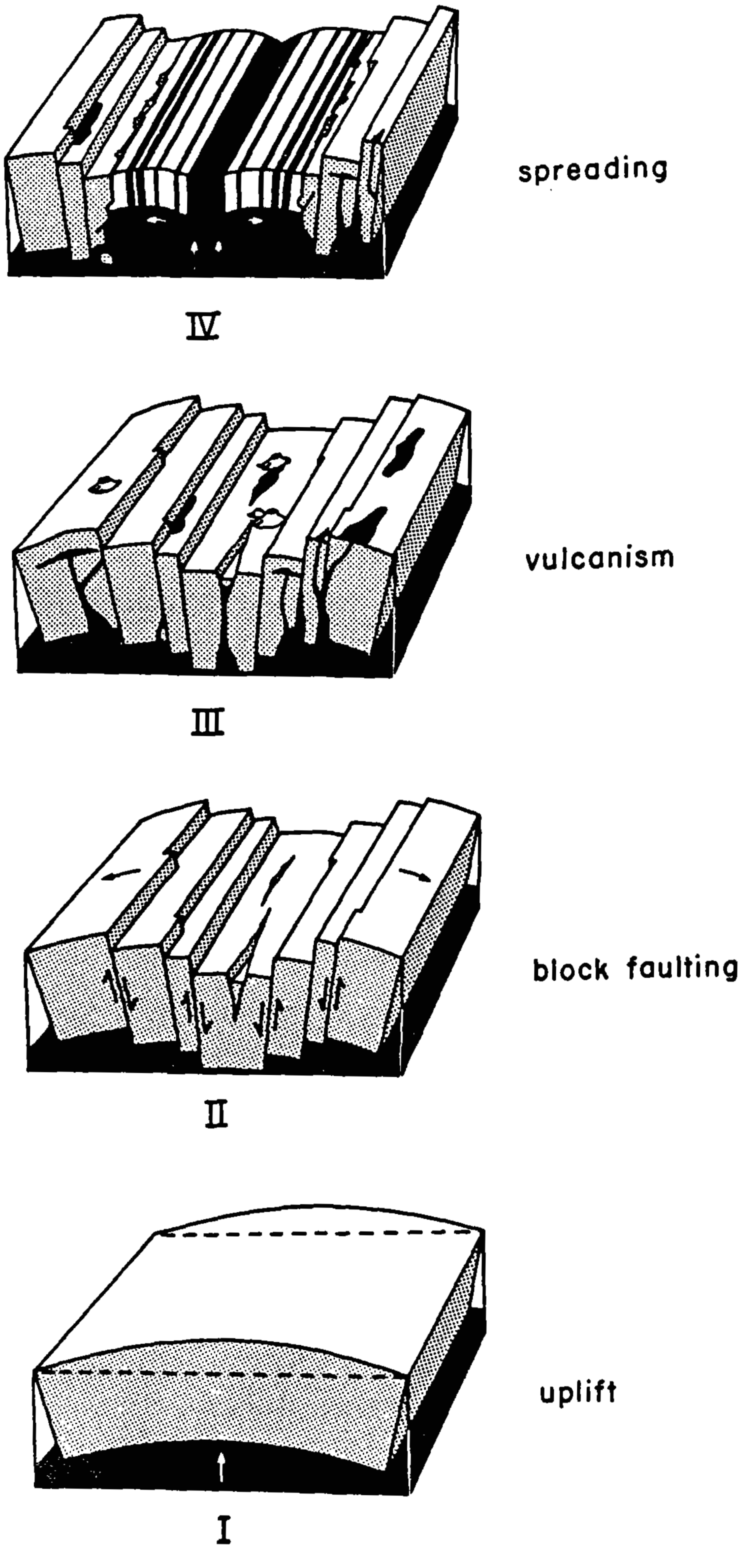

Figure 14. Hypothesized sequence of events starting with uplift and block faulting, ranging over igneous and hydrothermal activity, lithospheric thinning, leading finally to seafloor spreading.

SEDIMENTARY STRUCTURE OF THE NORTHERN RED SEA PREPARATION OF A1: 1,000,000 BATHYMETRIC CHART OF THE NORTHERN RED SEA AND STUDY OF THE MORPHOSTRUCTURAL FRAMEWORK BY CARTOGRAPHIC ANALYSIS

A. FERRAGNE, R. GRIBOULARD, R. PRUD'HOMME,
J-M. RATEL AND M. VIGNEAUX

Institut de Geologie du Bassin d'Aquitaine
Universite de Bordeaux I
351, Cours de la Liberation
33405 Talence Cedex, France

ABSTRACT

Continuous analysis of the northern Red Sea domain has been made for the marine and the continental areas. A1: 1,000,000-scale bathymetric chart has been drawn on the basis of relevant GEBCO plotting sheets, correlated with marine charts from the Service Hydrographique de la Marine. The morphology is described in the structural framework of this domain, with a three-dimensional visualization of the seafloor relief. The orientation and deep variations of the axial valley are shown, with the presence of submarine deeps. The main characteristics of the marginal shelf are indicated both in the morphology and the lithological and sedimentary structure. The evident dissimilarities between the Gulfs of Suez and Aqaba, because of their genetic differences, are perfectly illustrated. Cartographic analysis, by a filter effect, allows the elaboration of a synthetic morphostructural sketch. The essential part of the ante-opening regional geological structure (structure at 30°N, Zabargad transitional zone, at 20°N Al Nadj fault zone) is revealed; this part controls the evolution of the marine domain. The role of rifting structures, such as transform faults, is confirmed. The Aqaba Gulf Fault appears clearly as a sinistral zone. The main deeps bordering the median axis of the rift are observed to cross transversal structures. Through the investigation of depth variability by morphostructural analysis, a domain with different morphostructural characteristics can be defined, particularly with regard to regional geological evolution.

SUPERFICIAL SEDIMENTS OF NORTHERN RED SEA

MAHMOUD KH. EL-SAYED

Department of Oceanography, Faculty of Science
Alexandria University, Alexandria, A.R. Egypt

INTRODUCTION

As one of the youngest oceanic zones on Earth - an ocean in the making - the Red Sea is of enormous interest in understanding the structural and tectonic phenomena related to ocean floor spreading and in reconstructing the history of global tectonics. Although investigation of the Red Sea goes back into the 19th century, the mid-1960's mark the true beginning of its scientific exploration, with oceanic studies increasing both in number and diversity (THIESSE et al., 1983). However, most of these investigations, particularly those in marine geology, have been associated with brine area in the central axial valley (near 21°N), and zones characterized by metalliferous sediments. Numerous studies have been made concerning the paleontology, mineralogy and geochemistry of recent Red Sea sediments (see STOFFERS and ROSS, 1974).

Although some geological information of the northern Red Sea has been published, the surficial sediments in this area received little attention. Much of the available information on the texture, mineralogy, chemistry and paleontology of the surface sediments of the northern Red Sea comes from the fifty sediment samples collected during the Egyptian Preliminary Expedition to the northern Red Sea in the R.R.S. MABAHITH in 1934-1935. Studies of these samples were carried out by SHUKRI and HIGAZY (1944a,b); MOHAMED (1949); SAID (1949, 1950a,b, 1951); SHUKRI (1953); EL-DEEB (1978) and CHERIF et al. (1983). Some sediment samples from the Gulf of Suez and the Gulf of Aqaba also were studied within these investigations. Subsequently, both gulfs were investigated intensively, particularly the Gulf of Suez, after the reopening of the Suez Canal for navigation in the late 1970's (MOHAMED, 1979, 1980; STANLEY et al., 1982; NAWAR, 1983). Supplementary information on the sediments of the northern Red Sea was recently obtained from deep drilling in this area.

The major sedimentary constituents of the northern Red Sea sediments revealed from various investigations in the area will be summarized and critically reviewed. Sediments from the two gulfs are excluded.

GEOLOGIC SETTING OF THE NORTHERN RED SEA

The Red Sea extends in a northwesterly direction from the narrow strait of Bab-El-Mandab at its southern end to the tip of Sinai Peninsula, a distance of about 1930 km. It cuts across a huge dome of Precambrian basement rocks (Arab-Nubian Massif) flanked by epicontinental and marine sediments. At its southern end, the Red Sea is connected to the Gulf of Aden which trends ENE-WSW, opening into the Indian Ocean and separating the southern part of the Arabian Peninsula from the Somali Plateau. At the northern end, the Red Sea bifurcates into the Gulfs of Suez and Aqaba, creating the triangular Sinai Peninsula (ABDEL-GAWAD, 1969) (Fig. 1). The northern Red Sea proper (excluding the two Gulfs) lies north of about 24°N.

The geology of the northern Red Sea was previously reported by ABDEL-GAWAD (1969), SAID (1969), COLEMAN (1977) and many others. From the southern tip of Sinai to Ras Benas and Abu-Madd, the Red Sea is characterized by relatively straight and parallel coasts (Fig. 1). According to HALL (1976) these straight coasts suggest the absence of typical sea floor spreading anomalies in the northern Red Sea, as the magnetic lineations are poor and often non-existent in this region. According to STOFFERS and ROSS (1974) wells drilled around the edge of the Red Sea basin generally reach evaporites of Miocene and younger ages. Three wells in the northern Red Sea (Fig. 2) did not hit any evaporites, but bottomed in granite.

On the African side, Precambrian shield occupies about two-thirds the width of the Egyptian Desert. The thin coastal strip between the shield and the Red Sea is occupied by Miocene and younger sediments (Fig. 1). On the Arabian side, the shield is affected by an enormous shear zone. The shield on the east is covered by extensive Tertiary and Quaternary basaltic lava flows. Towards the Red Sea an irregular fault-controlled strip is occupied by Miocene sediments and Quaternary alluvial sediments and gravels. According to SAID (1969), the Pleistocene deposits of the Red Sea coastal plain include emergent thick sections of alternating massive coral reefs and gravel beds. AKKAD and DARDIER (1966) described four gravel beds along the Egyptian coastal plain. A similar succession was recorded along the coastal plain between Jeddah and Yanbu to the north (BEHAIRY, 1983). In addition to the old interbedded coral reefs, there are a series of Late Pleistocene reefs on wave-cut terraces of different heights above the modern sea.

The Red Sea is unique amongst the seas of the world in that no permanent streams flow into it; only winds, mostly northwesterly, and torrential rains transport terrigenous sediments (SHUKRI and HIGAZY, 1944a). A considerable amount of sediment in the Red Sea may be aeolian, but the actual amount is not known. During summer, northerly winds prevail through the whole Red Sea, while during the winter the prevailing wind is only northerly in the northern half of the sea (MORCOS, 1970).

RED SEA SEDIMENTS

STOFFERS and ROSS (1974) listed the major sedimentary constituents of the Red Sea sediments (LEG 23B Drilling) as follows:

Biogenic constituents: Nannofossils, foraminifera, pteropods, siliceous fossils, other fossils.

Volcanogenic constituents: Tuffites, volcanic ash, montmorillonite, cristobalite, zeolites.

Terrigenous constituents: Quartz, feldspar, rock fragments, mica, heavy minerals, clay minerals.

Authigenic minerals: Sulfide minerals, aragonite, Mg-calcite, protodolomite, dolomite, quartz, chalcedony.

Evaporitic minerals: Magnesite, gypsum, anhydrite, halite, polyhalite.

Brine precipitates: Fe-montmorillonite, goethite, hematite, siderite, rhodochrosite, pyrite, sphalerite, anhydrite.

The great range in composition of Red Sea sediments reflects depositional environment as well as the source of deriviation. Therefore the summarized list presented by STOFFERS and ROSS (1974) does not necessarily apply to the entire Red Sea. This list also does not include reefal sediments, which are of prime interest as their production is enormously high in the Red Sea.

The surficial deposits of the northern Red Sea can be placed into the following catagories:
1. Normal sediments of the northern Red Sea; 2. Reefal sediments; and 3. Hydrothermal deposits.

NORMAL SEDIMENTS OF THE NORTHERN RED SEA

The recent sediments of the northern Red Sea were described by SAID (1951) as whitish "coral muds" of high calcium carbonate content; these sediments do not possess significant quantities of chemically precipitated $CaCO_3$. The coarse fractions of these sediments (>20μ) consist largely of fragmentary organic remains. On the other hand, MILLIMAN et al. (1969) concluded that inorganically precipitated carbonates may account for more than half the deep-sea carbonates in the central and southern Red Sea. However, the paucity of lithic fragments, and the relatively low amount of magnesian calcite in cores north of 24°N (e.g. core C61-69G in MILLIMAN et al., 1969) suggest a lower rate of carbonate precipitation in the northern Red Sea. This supports SAID's conclusions, but one cannot ignore the minor occurrence of inorganic carbonate sediments in this region. Organic carbonate sediments also are found in the Gulf of Suez (MOHAMED, 1980).

The texture of northern Red Sea surface sediments (based on fifty samples, Fig. 3), varies widely SHUKRI and HIGAZY, 1944a). Sediments of the most northern section of the Red Sea are normally sorted, with median diameters ranging between 0.015 and 0.78 mm. However, samples near the entrance of the Gulf of Aqaba are generally fine-grained. There is also a remarkable difference in texture between the sediments of the Gulf of Suez and the Gulf of Aqaba.

SHUKRI and HIGAZY showed that the texture controls the distribution of some chemical constituents; organic matter (0.27%-1.7%; MOHAMED, 1949) is concentrated in the finer fractions, whereas carbonates (50%-95%; MOHAMED, 1949) are concentrated in the coarse fractions. Water depth also controls the sediment distribution in the northern Red Sea, with increasing of depth associated with a decrease in median diameter; sediments tend to be better sorted in the shallow areas (SHUKRI and HIGAZY, 1944a).

The mineralogy of the northern Red Sea sediments was studied by SHUKRI and HIGAZY (1944b), SAID (1951) and MILLIMAN et al. (1969). SHUKRI and HIGAZY (1944b) found 38 types of heavy and light minerals in the sand fractions. The light minerals are composed essentially of angular quartz indicating a crystalline rock source. Next in abundance are muscovite and the feldspars. Heavy fraction minerals are dominated by iron minerals (ilmenite, magnetite, hematite and limonite). Pyrite averages about 43% of the heavy residues of the Gulf of Suez but is less dominant in the northern Red Sea. The grains are rounded and range between 13 and 61% of the total heavy fractions. Biotite ranges between 2 and 58% and occurs in dark brown and green varieties. Amphiboles are represented by hornblende and actinolite, and range between 5 and 48%. Both monoclinic (augite, diopside and diallage) and rhombic pyroxenes (hyperthene, enstatite) are found; their total relative frequencies vary from 2 to 30%. Epidote is represented by pistachite, zoisite and clinozoisite; its total frequency varies from 1 to 10%. Minor occurrences of chlorite, tourmaline, sphene, apatite, zircon, rutile, garnet, staurolite, kyanite, fluorite, glauconite, spinel, olivine, anatase and rock fragments are recorded. The distribution of minerals in the northern Red Sea sediments along the different sections is shown in Figure 4.

The minerals of the Red Sea surface sediments are mainly derived from crystalline rocks. According to SHUKRI and HIGAZY (1944b), the mineral assemblage of a sediment depends on three main factors: a) Nature of distributive rocks; b) Type of weathering and mode of transportation; and c) Post-depositional processes, leading to simplification of detrital grains and increase of authigenic minerals.

The detrital minerals are transported to the Red Sea mainly by wind, though heavy rains and waves undoubtedly contribute smaller amounts. The mineral constituents of the sediments are uniform in all parts of the sea and they all display the same general character, whether collected from the southern, middle or northern portions of the Red Sea. Local conditions, however, give rise to minor variations in the frequencies or types of minerals.

The carbonate mineralogy of the sediments in the northern Red Sea shows that aragonite is the most common mineral and is definitely of organic origin; calcite is also present as well as minor amounts of dolomite-ankerite mineral group (SHUKRI and HIGAZY, 1944b). On the other hand, the finer calcareous fraction is mainly calcitic, with very little aragonite. SAID (1951) mentioned that the coarser fractions, from which some of the fine fractions is derived, consist mainly of aragonite. In neither of these studies was magnesian calcite distinguished from calcite.

The carbonate mineralogy of the northern Red Sea sediments might indicate two different sources of organic fractions, one is the nannofossils and benthic foraminifera (mostly calcite) and the other is the coarse coral remains (mostly aragonite). According to MILLIMAN et al. (1969) the deep-sea sediments of the Red Sea have the unique presence of lithified carbonate layers, which are mostly aragonitic. GEVIRTZ and FRIEDMAN (1966) estimated that 11% of the Red Sea sediments are lithified and aragonite is the dominant mineral. Aragonite is dominant in the coarse fraction, while magnesian calcite becomes dominant in the fine fractions (MILLIMAN et al., 1969). This variation of carbonate minerals relative to the size agrees with earlier observations of SAID (1951).

REEFAL SEDIMENTS IN THE NORTHERN RED SEA

One of the major contributors to nearshore and shelf sediments in the Red Sea are coral reefs. SAID (1951) reported that recent Red Sea sediments are generally coral muds and that the coarser fractions of these sediments are composed largely of coral fragments. MERGNER (1984) defined the Red Sea as a "coral sea" regarding its topography, climate and hydrography, and throughout its length the Red Sea is characterized by large areas of vigorously growing corals on fringing and barrier reefs (CROSSLAND, 1939).

In coral reefs, carbonate is deposited as both the reef framework and unconsolidated sediments. In terms of total carbonate production, the unconsolidated sediments produced by reefs appear to be far more important than the carbonate incorporated in the reef itself (STODDART, 1969; MILLIMAN, 1974). Because of the generally high wave energies, unconsolidated reef sediments are mostly composed of sand and gravel; silt and clay usually constitute less than 1-2% of the total carbonate (MILLIMAN, 1974).

Geological studies of coral reefs in the northern Red Sea started about one hundred years ago, but reefal sediments have been studied in this area for only about forty years. Modern and Pleistocene reefs of the south Sinai Peninsula were investigated by WALTHER (1888), HUME (1906), CROSSLAND (1939) and GUILCHER (1979). In Aqaba, RUPPEL (1829), WALTHER (1888), HUME (1906), FRIEDMAN (1966, 1968), MERGNER (1971), MERGNER and SCHUMACHER (1974), GVIRTZMAN et al. (1977), SNEH and FRIEDMAN (1980) and BOUCHON et al. (1981) likewise described coral reefs.

Texture and Distribution:

Red Sea reefal sediments are generally composed from debris of corals and coralline algae as well as mollusks, echinoderms and foraminifera.

In Aqaba, the composition of particles differs spatially: The sands surrounding patch reefs are composed almost entirely of skeletal debris, including fragments of corals, crustose coralline algae, gastropods, and pelecypods, together with foraminifera tests, small pelecypod shells, echinoid plates and spines (FRIEDMAN, 1968). GABRIE and MONTAGGIONI (1982) recognized eight sediment facies in Gulf of Aqaba reef sediments based on the total component composition and foraminiferal assemblages; four sediment facies were recognized using grain-size data only. Well-sorted, fine to medium, quartzofeldspathic sands (terrigenous facies) occur on beaches and outer sandy slopes near wadi mouths. Back-reef areas exhibit relatively well-sorted fine sands of terrigenous origin, coral and Milliolidae-Soritidae facies. Poorly sorted coral sands, and coral-coralline algal and Homotremid facies characterize the reef flat and the upper parts of coral-built fore reef areas, which respectively display an *Amphistegina-Spirolina* subfacies and an *Acerrulina* one. Poorly sorted medium sands of coral-molluscan-foraminiferal facies are restricted to the lower parts of the fore reef zone.

In Al-Ghardaqa area (Fig. 5), the coastal sediments are generally coarse, well-sorted, angular to sub-angular arkosic sands mixed with common rock-forming detritus from surrounding formations. These sands are mixed with significant amounts of biogenous fragments. Sediments in the intertidal zone are much finer and rich in carbonates. The lagoon is also covered by fine carbonate deposits mixed with reef fragments (EL-SAYED and HOSNY, 1980; EL-SAYED, 1984).

Considerable quantities of insoluble residues characterize both Aqaba and Ghardaqa sediments (FRIEDMAN, 1968 and EL-SAYED, 1984). The insoluble residue includes quartz and feldspar grains as well as chlorite, kaolinite and illite. In Aqaba the terrigenous debris is derived from the local Precambrian rocks, while in Al-Ghardaqa it originates from the surrounding Neogene formations.

Geochemistry:

Little information is available on the elemental composition of reefal sediments from the northern Red Sea. Owing to the admixture terrigenous and bioclastic constituents, the chemical constituents vary considerably with locality (EL-SAYED, 1984).

The average chemical composition of the reefal sediments of Aqaba and Al-Ghardaqa is listed in Table 1. High positive correlation values are obtained between Fe_2O_3 vs MnO; CaO vs MgO; SrO vs CaO and MgO for most of the reef sediments in the northern Red Sea (EL-SAYED, 1984). The average concentrations of some heavy metals (Fe, Mn, Cu, Zn and Cd) in the reefal sediments are shown in Table 2.

EL-SAYED (1983) used the trace metals as a possible tool to descriminate amongst the different reefal depositional environments in the area north of Jeddah. Strontium and magnesium in aragonite also were used by FRIEDMAN (1968) as a tool in the identification of the constituents in skeletal sands from Aqaba.

Mineralogy of Reefs and Reefal Sediments:

The framework builders of the reefs in the northern Red Sea utilize skeletons of aragonite and high Mg-calcite (FRIEDMAN, 1968 ; EL-WAKEEL et al., 1984). Aragonite is basically produced from corals, fragments of *Halimeda*, and inorganic carbonate precipitates in local areas (e.g., Ras Matarma lagoon). On the other hand, Mg-calcite is mostly produced from coralline algae and foraminifera. The relative abundance of the major carbonate minerals depends on the proportional mixture of various organisms in the reefal sediments. The relation between the major carbonate minerals in the reefal sediments from different areas of the Red Sea is shown in Figure 6.

HYDROTHERMAL DEPOSITS IN THE NORTHERN RED SEA

Metalliferous sediments associated with hot brines in the axial trough deeps were reported two decades ago. Ever since their discovery, the geothermal system of the Red Sea rift zone has been investigated nearly continuously. These investigations have revealed the presence of several deeps characterized by the metalliferous sediments. BACKER and SCHOELL (1972) located 17 brine pools in the Red Sea floored by sediments of hydrothermal origin. Further investigations of the Red Sea rift zone resulted in the discovery of new deeps in the northern part, increasing the number of deeps along the axial trough of the Red Sea to 20 (Fig. 7). Most of the deeps are filled with brines of different salt concentrations. However, the sediments underlying the brines in some deeps appear to be normal deep-Red Sea sediments with a significant content of nannofossils and pelagic microfossils. In other deeps, where the hydrothermal influence is strong, the bottom is covered with multicoloured, metal-rich sediments. The process of formation of these metalliferous sediments is clearly represented in the Atlantis II Deep, where metal-rich hydrothermal solutions are being discharged at the present time.

The first major scientific results indicating the potential economic value of sediments and of the overlying brines in Atlantis II Deep were published by DEGENS and ROSS (1969). Subsequent studies by BACKER and SCHOELL (1972) and BACKER and RICHTER (1973) focussed on various aspects of the brine sediments and helped to draw the attention of several ore-prospecting companies and scientific organizations to the Red Sea deeps. Intensive investigations were carried out on the Atlantis II Deep, which covers an area of 60 km^2. However, less attention was paid to the other deeps, particularly in the northern Red Sea, and therefore little is known about the composition and the commercial values of the metalliferous sediments in these areas.

Recently, however, BEHAIRY et al. (1985) studied some five sediment cores from the Red Sea deeps, three of which came from the northern Red Sea (Fig. 7). The study of these cores reveal that a biostratigraphic division of sediments can be made by using the distribution of some planktonic forams which are sensitive to temperature and salinity. Most of the sediments in the cores represent Holocene normal marine conditions and Upper Würm pleneglacial hypersaline conditions. Mineralogical and chemical data show that the sediments of Kebrit, New and Shaban deeps (Fig. 7) are similar to the normal deep-Red Sea sediments, which contain a mixture of organo-detrital components. However, at various levels, there are intercalations of volcanic and/or hydrothermal materials.

The Holocene sedimentation rates in the deeps are highly variable (9-31 cm 1000 y^{-1}), with a general increase northward. Slumping associated with tectonic activity in the north seems to have contributed large quantities of sediments to the northern deeps.

SUGGESTIONS FOR FUTURE RESEARCH IN THE NORTHERN RED SEA

Although a number of investigations have been carried out on the surface sediments of the northern Red Sea, it is hard to draw concrete conclusions due to the local nature of these individual studies. The paleontological character of the sediments of the northern Red Sea recently was discussed by CHERIF et al. (1983) using modern statistical methods.

Although it has been fifty years since the "Mabahith" Expedition, more scientific information is needed. In particular, there is need for:

a) Baseline studies on the various geological aspects of the northern Red Sea sediments;

b) Economic potentiality of the brine sediments in some deeps.

Geological studies should concentrate on the following:

a) More detailed knowledge is needed on the surface sediments of the northern Red Sea, particularly their geochemistry, biogeochemical cycle and rate of sedimentation.

b) Various sources of Recent sediments should be defined more precisely, especially terrigenous sediments and eolianite deposits.

c) Both the eastern and western flanks of the northern Red Sea are required, particularly studies on the geology of coral reefs and reef sediments. This requires detailed transects on both sides along the Egyptian and Saudi coasts. Such information does not exist yet.

d) Degradation of coral reefs (biological and mechanical) and rate of production of reef sediments, calcification, diagenesis and reef conservation should be studied.

e) The sediments in cores from some deeps in the northern Red Sea show some indications of brine precipitates. Long cores or deep drillings in the northern Red Sea are urgently needed to assess the possible occurrence of hydrothermal metalliferous sediments. From scientific point of view, this will also help us to understand the evolution of the northern Red Sea.

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Table 1. Major chemical constituents of reefal sediments from the northern Red Sea.

Concentration %	Aqaba*	Al-Ghardaqa**
SiO	12.41-45.47	12-33
Al O	0.96-5.34	6-12
Fe O	0.27-0.73	0.17-0.6
MnO	0.02	0.01-0.03
CaO	21.7-43.20	16.34.1
MgO	0.97-1.40	3.06-5.26
K O	0.42-1.40	0.1-0.3
Na O	0.42-1.27	0.1-0.4
Sr O	0.42-0.72	0.33-1.3

* Data From FRIEDMAN (1968)

** Data From EL-SAYED (1984)

Table 2 Heavy metal concentrations in reefal sediments from the Red Sea and other regions.

Element (ppm)	A	B	C	D	E
			Area		
Fe	197	785	304	2657	250
Mn	20	51	16	120	15
Zn	11	10	--	30	--
Cu	5	13	<10	20	--
Cd	1.5	0.9	--	--	--

A North of Jeddah, Red Sea (EL-SAYED, 1983)

B Jeddah-Yanbu nearshore area (BEHAIRY et al., 1983)

C Aqaba (FRIEDMAN, 1968)

D Gardaqa (EL-SAYED, 1984)

E Bahama coral mud (MILLIMAN, 1974)

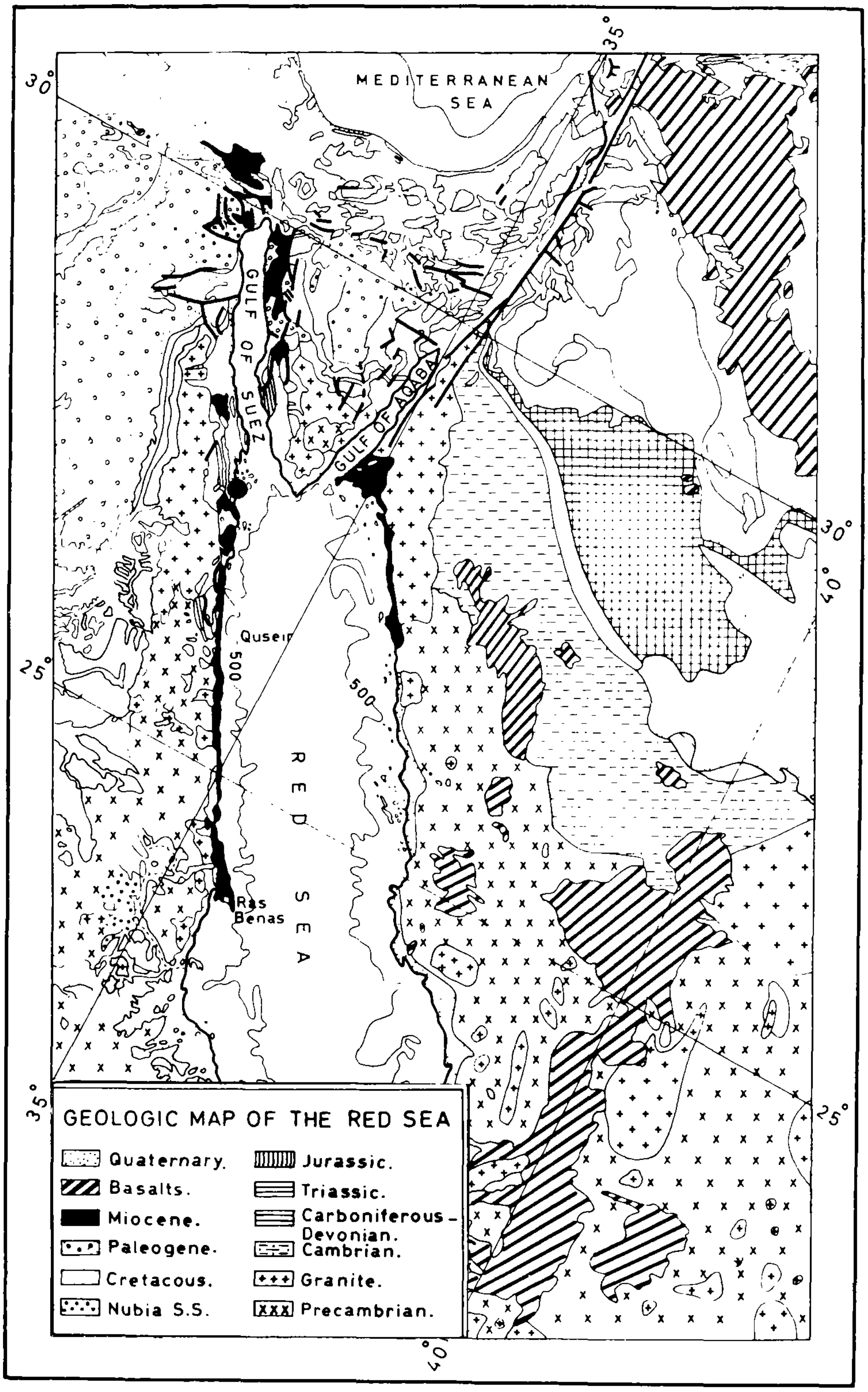
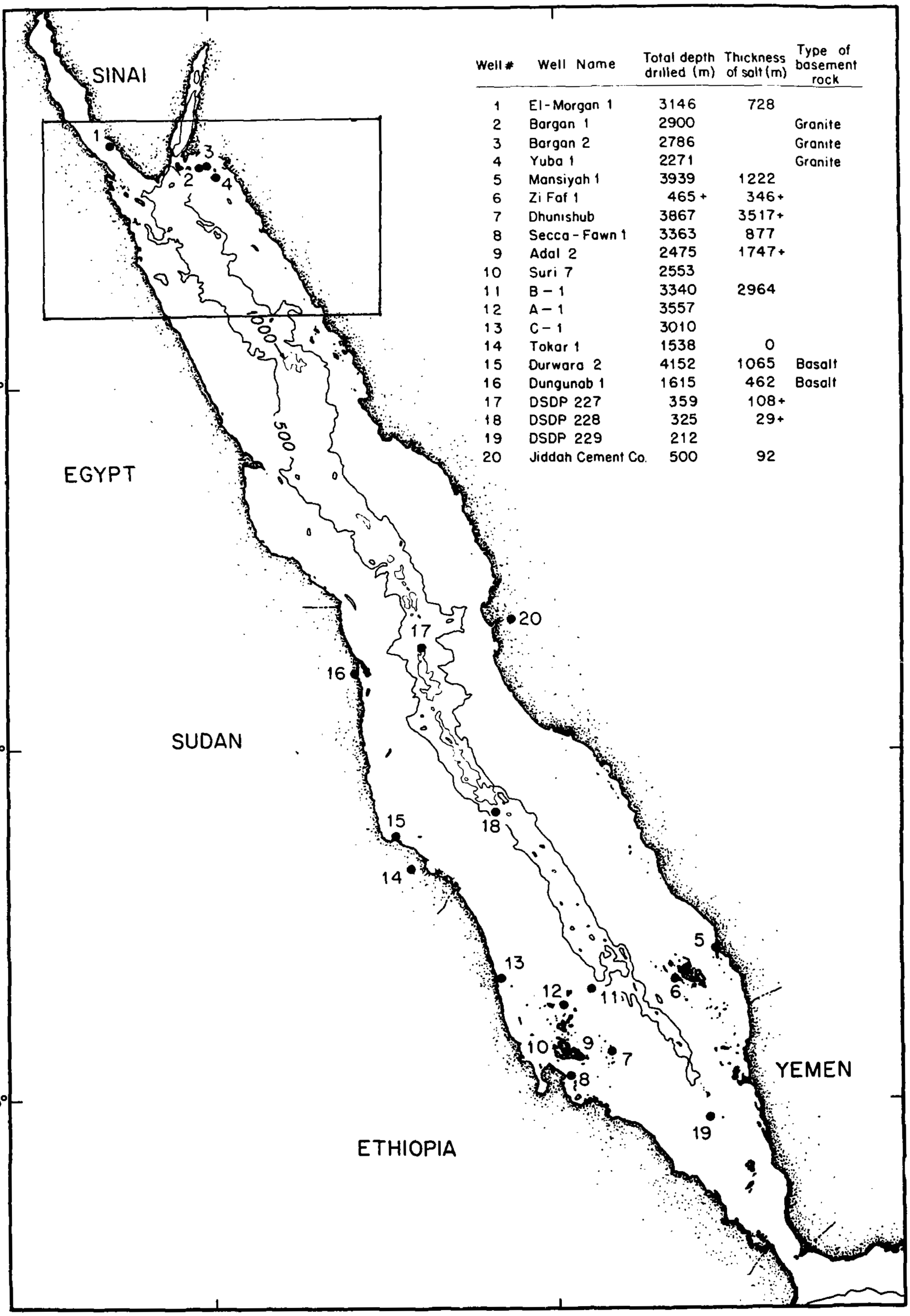


Figure 1. Geologic map of the northern Red Sea (after SAID, 1969).



Well #	Well Name	Total depth drilled (m)	Thickness of salt (m)	Type of basement rock
1	El-Morgan 1	3146	728	
2	Bargan 1	2900		Granite
3	Bargan 2	2786		Granite
4	Yuba 1	2271		Granite
5	Mansiyah 1	3939	1222	
6	Zi Faf 1	465+	346+	
7	Dhunishub	3867	3517+	
8	Secca-Fawn 1	3363	877	
9	Adal 2	2475	1747+	
10	Suri 7	2553		
11	B-1	3340	2964	
12	A-1	3557		
13	C-1	3010		
14	Tokar 1	1538	0	
15	Durwara 2	4152	1065	Basalt
16	Dungunab 1	1615	462	Basalt
17	DSDP 227	359	108+	
18	DSDP 228	325	29+	
19	DSDP 229	212		
20	Jiddah Cement Co.	500	92	

Figure 2. Locations of the drilling sites in the Red Sea (after STOFFERS and ROSS, 1974).

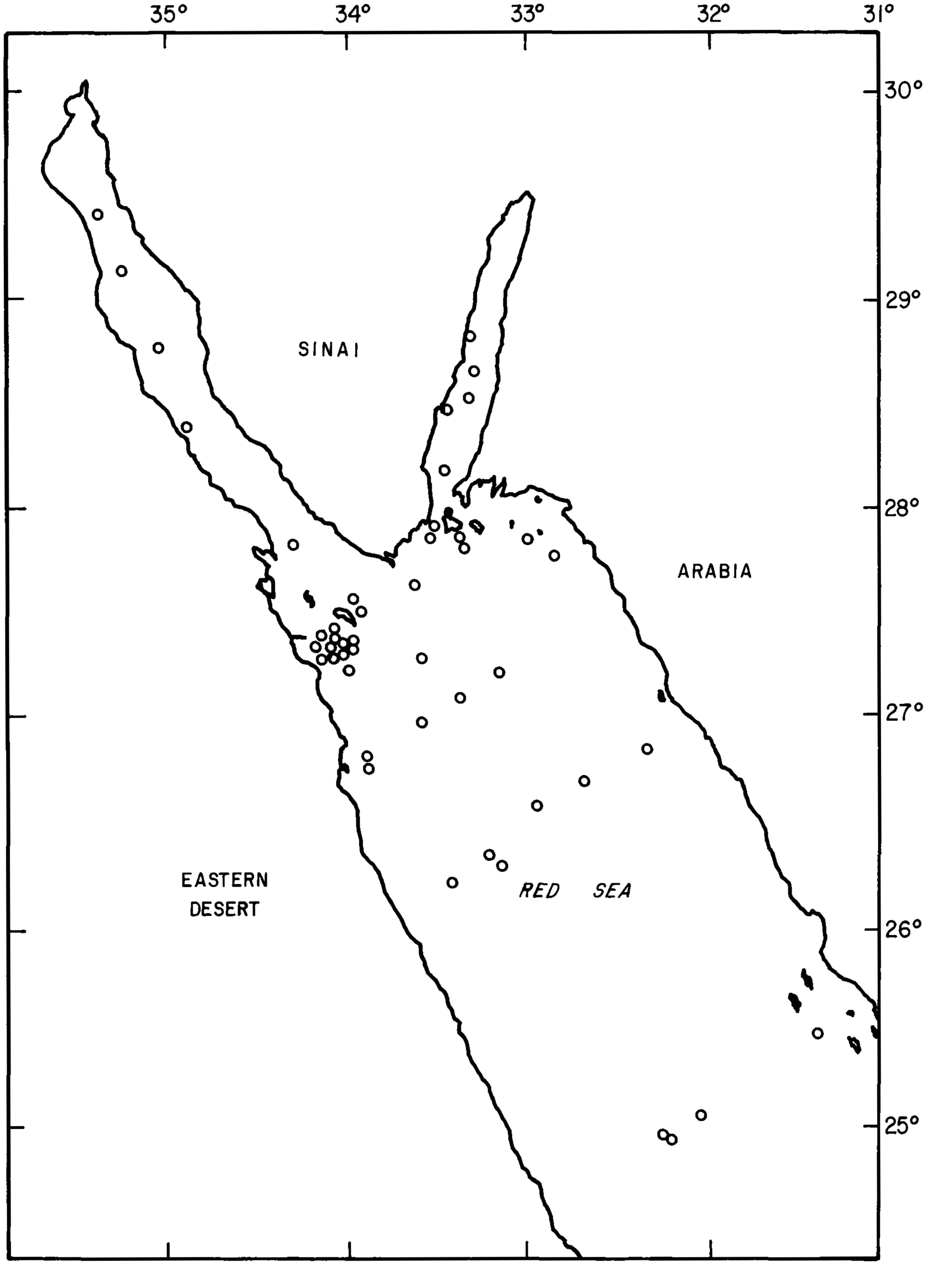


Figure 3. Location of surficial sediments in the northern Red Sea (after SHUKRI and HIGAZY, 1944a).

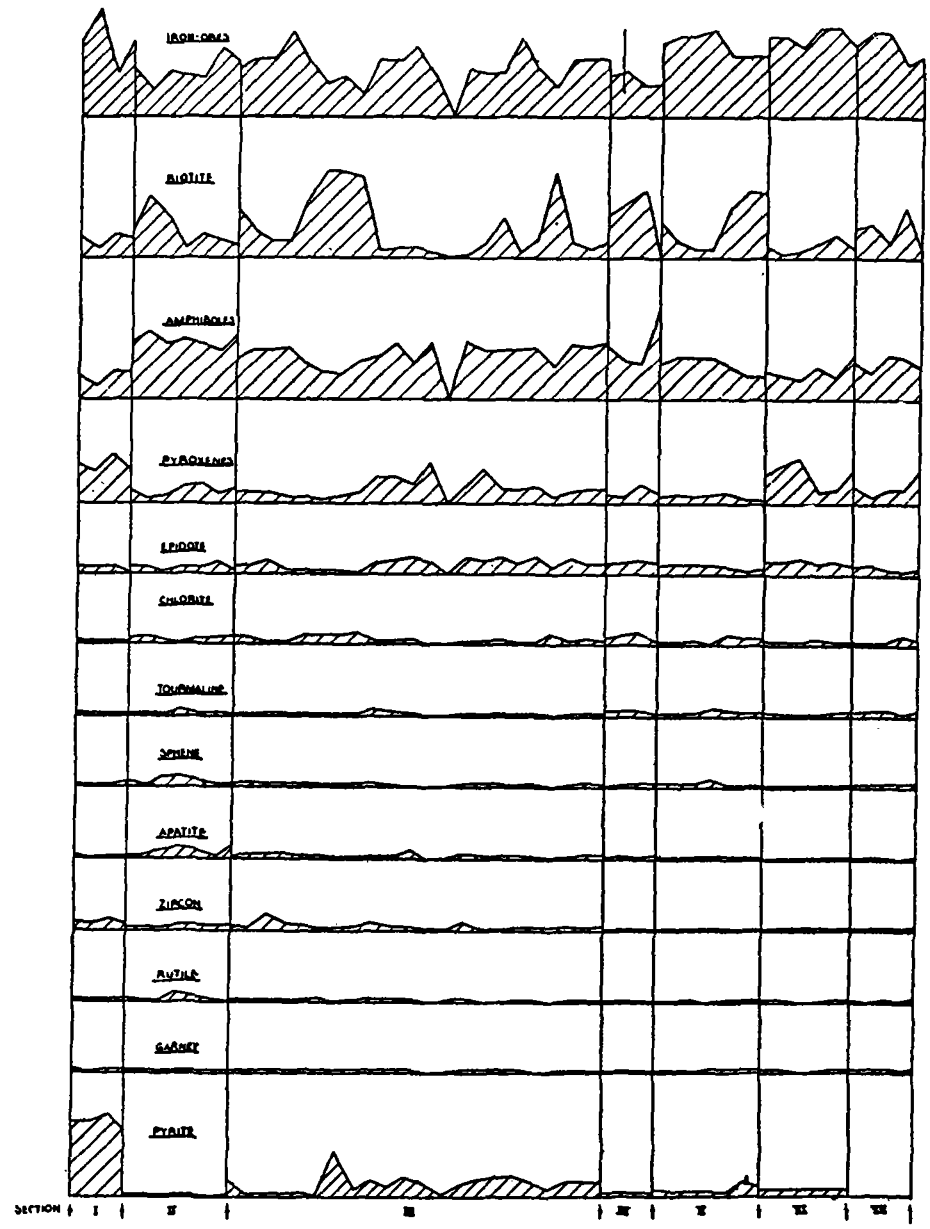


Figure 4. Distribution of minerals in the different sections in the northern Red Sea (after SHUKRI and HIGAZY, 1944b). Sectional: Gulf of Suez; 2: Gulf of Aqaba; 3: Ghardaqa-shadwan-Sanafir; 4: Yuba-Marsa Daba'a; 5: Safaga-Mowila; 6: Qoseir-Brothers No'man; 7: Daedalus-Hanak.

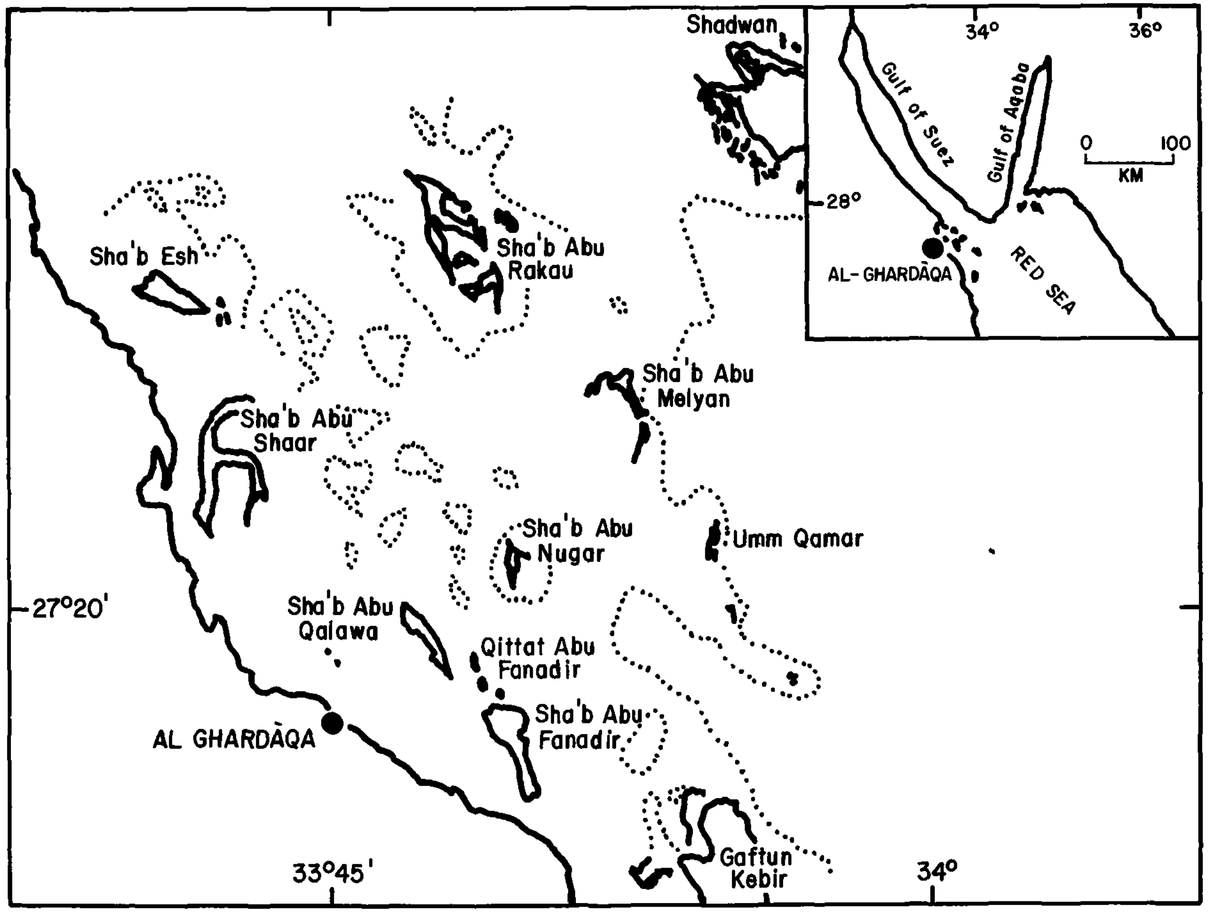


Figure 5. Ghardaqa Reef formation (after EL-WAKEEL et al., 1984).

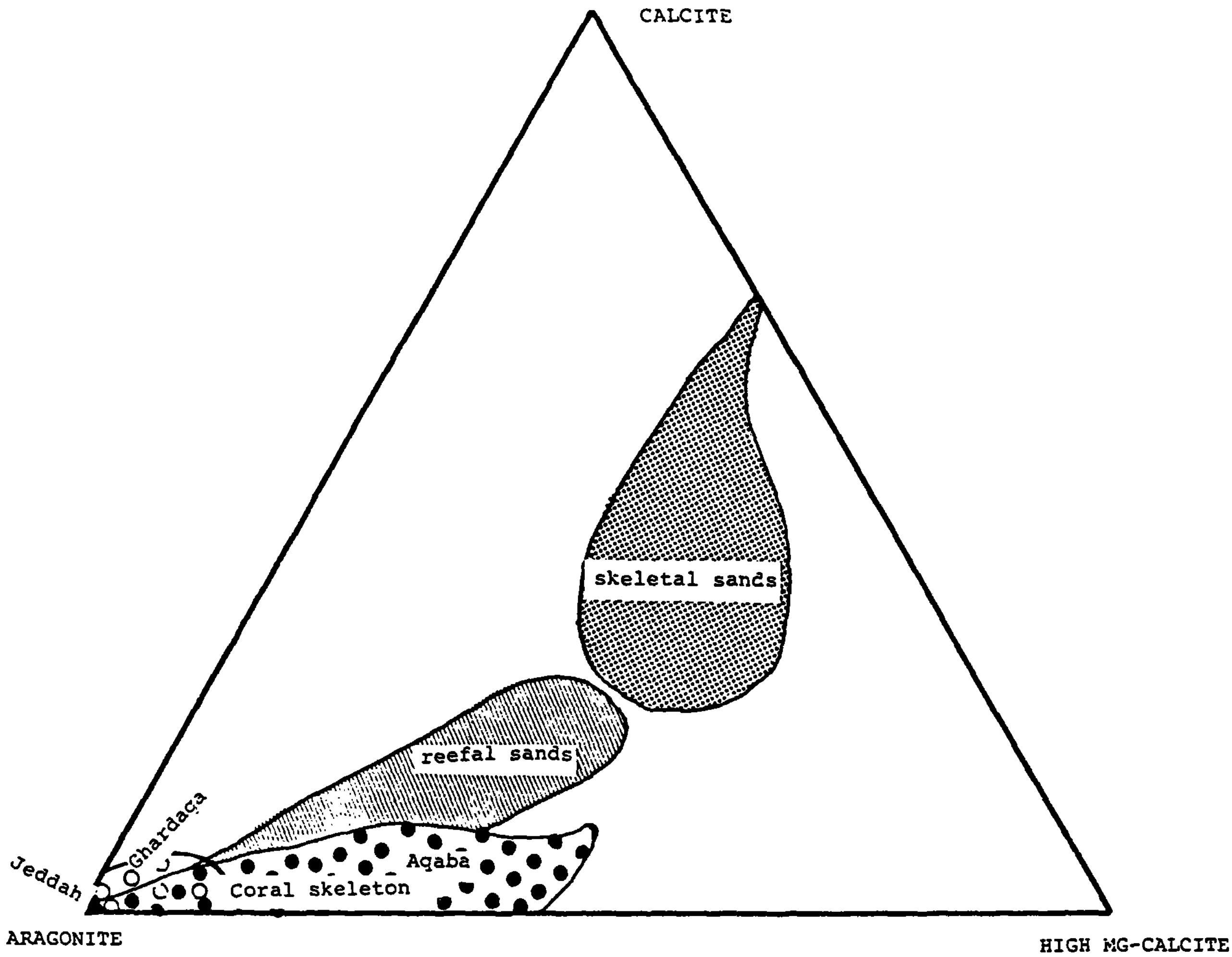


Figure 6. Ternary diagram showing the relation between the major carbonate minerals in the reefal sediments from the Red Sea.

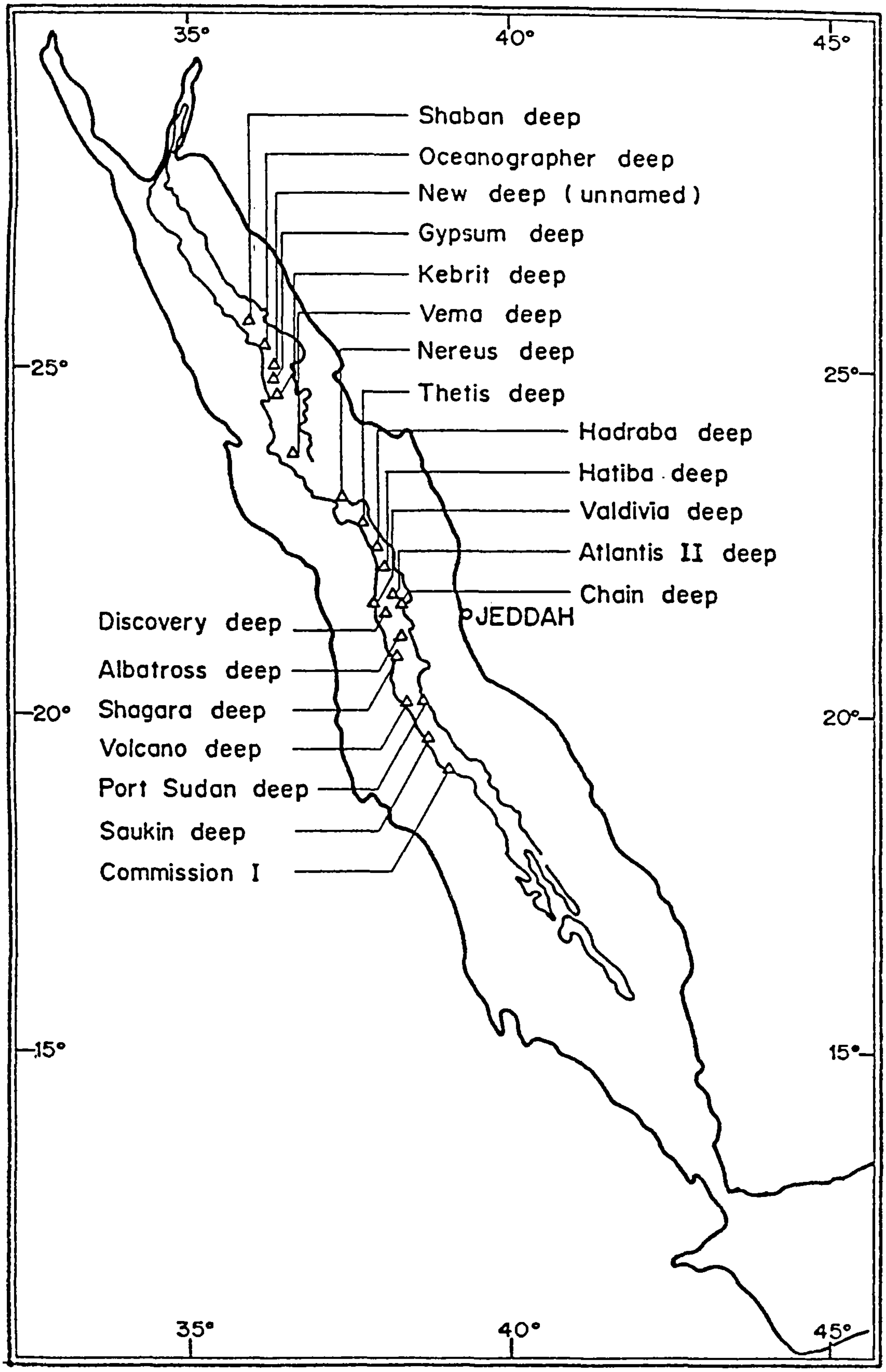


Figure 7. Location of the discovered deeps in the Red Sea (after BEHAIRY et al., 1985).

MARINE LIVING RESOURCES OF THE RED SEA

A. REFAI BAYOUMI

Director
Institute of Oceanography and Fisheries
Academy of Scientific Research and Technology
101 Kasr El-Ainy Street
Cairo, A.R. Egypt

ABSTRACT

The fauna and flora of the Gulf of Suez, the Gulf of Aqaba and the Red Sea show great variability, but consist almost entirely of members of the tropical Indo-Pacific assemblage; many of these species range across the Indian Ocean into the tropical central Pacific. Some algal species, however, come from the Mediterranean Sea. Red Sea flora include about 400 species of seaweeds, 10 sea grasses, 50 cyanophytes and two species of mangrove trees. As such the Red Sea fauna and flora offer good examples for the study of migration, endemism and distribution of marine organisms.

TOPOGRAPHY OF THE RED SEA

The Red Sea has a rifted origin, so that its parallel coasts contain systems of parallel faults running southeast - northwest. In the north, the Gulf of Aqaba marks the beginning of south-southwest — north-northeast rift which ends near Aleppo in Syria. Ras Mohamed, the southernmost tip of the Sinai Peninsula, thus lies at a triple junction between the continental plate of Asia, Sinai and Africa. Lateral displacements also have occurred, especially in the Gulf of Aqaba rift (VON WISSMANN, 1932; LAMARE, 1936; VON WISSMANN et al., 1942; RATHJENS, 1947; LAUGHTON, 1966).

Because of its origin, the Red Sea displays a fairly single topography (e.g. GOHAR, 1954), with a narrow straight trough extending more than 2000 kms in length. The sides run almost parallel to one another, with a maximum distance of some 340 kms at Massawa. The entrance at Bab el-Mandab is only 26 kms wide.

Deep areas in the Red Sea, exceeding 2000 metres, tend to be narrow, elongated, steep, and irregular. They generally occur near the long axis, but slightly to the east. Such rift-related deeps often descend by a series of steps about 400 metres in relief. Bottom topography is highly irregular, and locally seamounts are shallow enough to be capped by corals [e.g. Abul Kizan (Daedalus) and Panorama Reefs]. Similarly the Gulf of Aqaba has a trench over 1800 metres deep, discovered by the E.R.S. "Mabahiss". This also lies nearer to the east side of the Gulf.

The Gulf of Suez offers an extreme contrast. It is flat-bottomed and shallow, having an average depth of only 50 metres. It shoals to the north; greatest depths in the south are about 80 metres.

An important feature of the topography of the Red Sea is the shallow sill (about 100 metres deep) in the south, slightly north of the Straits of Bab el-Mandab, and on which lie the Hanish and Jebel Zukur Islands. As will be seen later, this "sill" has a profound effect not only on the hydrography of the Red Sea but also on its biology, and may be considered as the oceanographic boundary of the Red Sea.

THE RED SEA ENVIRONMENT

The pelagic and benthic communities of the Red Sea are intimately linked with its hydrography which differs from normal oceanic conditions (see GOHAR, 1954, for a detailed review). This abnormality is largely due to the partial isolation of the Red Sea basin from the open ocean, the prevailing wind system and excessive evaporation. The shallow sill north of Bab el-Mandab allows only a partial inflow of Indian Ocean surface water from the Gulf of Aden into the Red Sea. In return, hypersaline Red Sea water exits over the sill into the Gulf of Aden.

The inflow of Gulf surface water is significant from November to March, when southeasterly winds prevail in the southern part of the Red Sea. Further north, the northward flow of surface water is against the wind field, blowing from the NW. These northwesterly winds prevail throughout the year north of about 19°N. In summer, i.e. from May to October, the NW wind field covers the entire Red Sea, reversing the surface flow pattern in the south. The induced outflow of the Red Sea surface water into the Gulf of Aden from June to September, causes a compensatory current from the Gulf into the Red Sea. The core of this countercurrent ranges between 30 and 80 m depth, penetrating the Red Sea proper to approximately 19°N (PATZERT, 1974).

Due to high evaporation, temperature and salinity of the surface water of the Red Sea are considerably higher than in the Gulf of Aden. The mean surface temperatures fluctuate between 25° and 32°C in the south, and between 21.3° and 27.9°C in the north. Salinity increases from 37‰ in the south to 40‰ in the north. In winter, thermal convection in the northern part of the Red Sea makes the cooled (21.3°C) saline surface water sink below 200 m, thereby feeding the homogenous deep water (21.5°C, 40.5‰S) (MORCOS, 1970). Although this deep-water body is not totally isolated, any exchange of deep water with the surface layer is limited and slow.

Nutrient levels of the Red Sea are low. Their decrease from the north reflects the main nutrient influx from the Indian Ocean and (therefore) limited vertical mixing. Dissloved oxygen is present throughout the water column, its concentration is balanced between biological and chemical oxygen consumption as opposed to oxygen transport with water masses. The upper water layer has an oxygen contents of 4.25 ml l^{-1}, while 2 ml l^{-1} (minimal values of 1.5 ml l^{-1}) are found in greater depths. A fairly sharp oxycline exists around 300 m (MORCOS, 1970). It is assumed that the biological oxygen demand in deep water is low. Because no measurements of oxygen consumption are available, replacement times of oxygen and urnover rate of water masses cannot be computed. The oxygen present at all depths demonstrates that the entire water body is included in this exchange, although the exchange above the oxycline clearly is much faster than below it.

PELAGIC COMMUNITY

The pelagic community of the Red Sea (both phyto- and zooplankton) is still poorly investigated, particularly compared to the Mediterranean and the Arabian Seas. Nevertheless, we can observe some basic features concerning the diversity of species and their distribution and productivity in the Red Sea. These features are derived from incidental sampling of biological data, which are extrapolated on the basis of the slightly better-known hydrographic properties of the Red Sea.

In comparison with other deep-sea areas the Red Sea is not particularly deep, nor is the deep sea far from land. Most of the Red Sea lies between 500 and 1000 m, while the greatest inhabitable depth is only around 2000 m. Surface sediments of the hot brines in the deeper holes may be colonized by a few bacterial species only, while deeper brine layers are anoxic and without life. Estimated standing stock weight of organisms, based on knowledge from other oceans, may amount to a few grams per square meter. However, in the Red Sea productivity is low, and nothing is known on the fate of organic matter during sinking. Because of the high temperature throughout the water column, it can be assumed that bacterial degradation is fast and little food reaches the deep-sea bottom. In other oceans, the deep temperature below a few hundred meters causes a refrigerator effect, and more food material is preserved.

PRODUCTIVITY

The knowledge on the productivity of the Red Sea is insufficient for both the phytoplankton and the zooplankton communities. For example no carbon-14 productivity measurements are available to calculate primary production in the Red Sea. Chlorophyll values, although somewhat ambiguous for productivity calculations, suggest that Red Sea productivity is low. Species distribution seems to decrease to the north, and a similar trend is suggested for productivity.

There are indications of seasonal large-scale blooms in the Red Sea. In winter, during the NE monsoon, greater plant pigment concentrations are reported from the northern part of the Red Sea.

This elevated productivity may result from thermal convection. During this season diatoms seem to occur in increased numbers of cells. In summer, during the SW monsoon, higher chlorophyll values are measured south of 20°C. Dinoflagellates, which clearly are more abundant than diatoms throughout the Red Sea, reach their climax in cell number during this season (HALIM, 1969).

BENTHIC COMMUNITIES

Our knowledge on the deep-sea fauna and communities of the Red Sea is limited. The first expedition dredging in deep water was the Austrian "Pola" in 1897, and later some material was collected by the "Mabahiss" in 1933. However, during these early years, abundance of organisms was not of interest. Further bottom samples in deep water have been taken by several expeditions, but no quantitative studies of epi- or endo-fauna have been conducted.

CORAL REEF ECOSYSTEM IN THE RED SEA

The steep slopes of the Red Sea allow only relatively small fringing reefs along the coastline. However, in the central parts of the Red Sea (around Port Sudan) the reefs can extend some kilometers seawards, forming features very similar to barrier reefs.

From the zoogeographical point of view, the Red Sea belongs to the Indo-Pacific region, although its unique hydrographic character means that some Red Sea species (fish and invertebrates) are endemic. Moreover, the fauna (particularly the hermatypic coral fauna) is more diverse than the fauna of the neighbouring Indian Ocean. Therefore some caution should be used when transferring results of investigations on coral reefs from the Indo-Pacific to the Red Sea.

Little is known about the reefs of the central parts of the Red Sea. Some investigations on faunistic and geologic zonations have been done in the Gulf of Aqaba and the Farasan Archipelago (Sarsi Bebir) (e.g. CROSSLAND, 1907, 1939; FISHELSON, 1970, 1971; LOYA, 1972, 1976; LOYA and SLOBODKIN, 1971; MERGNER, 1967, 1971; SCHEER, 1967). The results of these investigations demonstrate the diverse zonations in the coral reefs. Zonations are primarily dependent on the shape of the coastline, its exposure to wave action, on fluctuations of water level, and on regional geology and geologic history. In general, however, three typical ecological zones can be distinguished: the reef flat, the reef slope, and the fore reef.

As in other areas, corals in the Red Sea constitute the basic reef framework and substrate for many other organisms which attach to or penetrate the skeletal mass, (sponges, polychaetes, sipunculides, bivalves and gastropods). Corals also provide shelter for many fishes as well as various invertebrate species. Reef communities are resource-limited systems in which diversity is maintained mainly by biotic specialization. Space is probably the major limiting factor in coral reef fish communities; many species have exploited biotic rather than physical habitats. Coral reef fishes exhibit a particularly diverse range of feeding, some species consuming benthic algae growing and others consuming zooplankton or bottom fauna, including the corals themselves.

Many reef building organisms harbor symbiotic algae which have long been thought to be both an important carbon source and a catalyst for calcification of these organisms. Symbiotic algae hosts include corals, bivalves, ascidia and foraminifera. Corals are major consumers of primary production on the coral reefs, and they (or their symbiotic algae) in turn provide a source of food for many other species on the reef. Organic material excreted by the corals provides nutrients for heterotrophic bacteria, which also constitute a prominent food source for other organisms.

Algae provide food, substrate and shelter for numerous organisms and may be the main food source for various animals with economic importance, such as fish, turtles and, under certain conditions, dugongs. Although algae are not presently utilized by man in the region, some areas may be suitable for the farming of economically important sea weeds. Distribution, abundance, seasonal variation and relative importance of the various algae in different types of reefs and reef communities in the region, however, are poorly understood, as are the factors influencing their distribution. In the Red Sea we are particularly ignorant of algae in deeper waters.

Comparative floristic and ecological studies of the algal communities of different reef types, including aspects such as their relative importance in bioerosion, and their role in the "algal rim" on wind reef margins, are required. The possible use of algae as environmental indicators, at both small-scale and broad-scale levels also needs to be investigated.

THE FAUNA OF THE RED SEA

During the last decades investigations conducted on the different fauna in the Red Sea have resulted in numerous publications. Particular mention should be made of reports on fish. SMITH (1950), CLARK and GOHAR (1953), BUDKER and FOURMANOIR (1954), MARSHALL (1964), ROUX et al. (1955), ROUX-ESTEVE (1956), RANDALL et al. (1978) and GOHAR (1954) discussed the distribution of the different biota and environmental controls for these distributions. TIRMIZI and KAZMI (1983) presented an overview of the marine crustacea of Pakistan with remarks on species in the Red Sea and the Mediterranean.

In his review of the shark fauna of the Red Sea, COMPAGNO (1983) noted the absence of endemic shark species, the lack of members of the orders Squaliformes, Hexanchiformes, Pristiophoriformes, Heterodontiformes and possibly squatiniformes, and the virtual absence of deepwater benthic or bathypelagic species except *Iago omanensis* (a triakid specialized for environments with low oxygen levels). No members of the family Scyliorhinidae occur, a shark that is particularly abundant in the western Indian Ocean. All of the Red Sea shark species also are found in the western Indian Ocean. The limited deepwater shark fauna in the Red Sea is attributed to the high salinity, high temperature and low oxygen levels, plus the shallow sill at Bab el-Mandab. Campagno concluded that the Red Sea shark fauna probably originated by direct dispersal from the western Indian Ocean.

FRAZIER (1983) mentioned that marine turtles are poorly documented in the Red Sea. There are five recorded species of marine turtle, all pantropical in distribution, but only *Chelonia mydas* and *Eretmochelys imbricata* have been documented to nest there.

The Red Sea has an exceptionally rich in molluscs, but few data are present on the differences in fauna from various parts of the Red Sea. In this connection several questions arise: What ecological processes allow more species to co-exist in some parts of the Red Sea than in other parts? Are important resources (such as food and habitat) more finely partitioned? Do physical and/or biological disturbances (i.e. predation) keep species populations reduced in size?

FISH RESOURCES IN THE RED SEA

Eight hundred species of fish are present in the Red Sea as compared to 2000 species in the Indian Ocean. The fish can be characterized more by their diversity in forms and colours rather than in wealth of fish stock. Nevertheless, the marine life of the Red Sea is remarkably rich, and many of the fishes are able to tolerate wide extremes of temperature and salinity. The richness of marine life is apparent in the relatively large number of species seen in commercial catches.

Nevertheless, the Red Sea is considered a particularly unexplored area in terms of commercial fisheries. One of the reasons for this lack of data is the poor development of the fishing industry in coastal countries, and as a result, incomplete and fragmentary statistics of fish landings. Analysis of literature and data shows that the western Indian Ocean and the Red Sea are rather difficult for commercial utilization, especially for development of trawl fishing. The difficulties lie in the nature of atmospheric circulation and complicated weather conditions as well the narrow and rocky shelves along most part of the coast. The difficulties in commercial utilization in this region are aggravated by the great variety of fish species, which makes sorting and treatment of catches more difficult.

Therefore, increased oceanographic and biological research could help us understand variations in yields and their relationship to changing oceanic conditions. Knowledge of oceanic events during the early life stages (ichthyoplankton) of many commercial fishes can be used to predict population size two or three years hence and thus to determine changes in management strategies.

REVELLE (1981) mentioned that the problem in fisheries management is really an ecological one. All species of important finfish and shellfish in a particular region compete for food or serve as food for each other at different stages of their life histories.

When we review the work done on the biology of fish of the Red Sea it appears that there is a profound gap in knowledge of ecology of these fish as follows: How do the various planktonic fish larvae find their way back into their specific eco-niches? How are juveniles recruited? How and when do the larvae or postlarvae of the edible fish congregate to form the huge aggregations observed along the shores? The need to manage natural resources (living and non-living) in the Red Sea at the regional dimension is a pressing one. At the same time efforts should focus on the role of variable environmental conditions, both natural and man-made, in affecting *in situ* behaviour of individual species. Joint research between biologists and physicists is necessary to understand the inter-relationships between the time and space scales of physical forcing functions and the resulting phytoplankton populations and higher organisms which comprise the food web within different Red Sea environments.

CONCLUSION

PREVIOUS RECOMMENDATIONS

The workshop on marine sciences programme for the Red Sea convened in 1974 reached a number of recommendations regarding marine scientific research on the Red Sea. The report of this workshop was used by ALECSO as a working paper in the "First Experts' Meeting on a Regional Programme for Environmental Studies on the Red Sea and the Gulf of Aden" in 1974. Preparatory studies for a Plan of Action were discussed in the "Second Experts Meeting", held in Jeddah, Saudi Arabia in 1976.

In 1981 the International Conference on Marine Science in the Red Sea made several recommendations, one of which encouraged research in the marine sciences with emphasis on shore ecology, coral reefs and associated littoral invertebrates and fish communities. This conference also recommended taxonomic studies based on the examination of adequate samples from different localities and examination of soft bottom benthic fauna within the Red Sea coastal area, with special attention to mangrove and coral reef communities. All of these research areas require additional hydrographical, physical and chemical baseline information. At the same time investigations on fish populations of benthic and pelagic subregions, dynamics of local and territorial populations of invertebrates, and the distribution and quality of mineral deposits are needed to provide a firm basis for future regional and informational use and protection.

The Workshop on Coastal Zone Management held in December 1981 made a number of recommendations, one of which dealt with marine resources: scientific and technical agencies (governmental and non-governmental) should collaborate in the exploration, assessment and exploitation of marine resources, so that they can make an adequate contribution to the development of the individual countries as well as the region as a whole.

In 1983 a round table discussion took place during the Mabahiss/John Murray International Symposium on Marine Science of the Northwest Indian Ocean and Adjacent Waters. The outcome of this discussion was published in a report including several suggestions and recommendations on marine scientific research projects which could benefit from international co-operation.

PRESENT RECOMMENDATIONS

As a result of the present review several questions can be raised and proposals made concerning living marine resources of the Red Sea. They can be summarized as follows:

1. What mechanisms control primary productivity in the different areas of the Red Sea, particularly in highly productive areas?
2. What are the characteristics of normal coastal processes? And by what mechanism(s) is this production transferred through the food chain to the fish?
3. What key features of the dynamic food web are important to the transfer of pollutants?

4. What is the relationship between production in the water column and sea bottom?
5. Is the present knowledge of fish migration and/or movement in the region sufficient to explain the spatial separation between the areas of highest primary production and the areas of greatest importance to the fisheries? What aare the distribution and abundance of different stocks?
6. What are the water chemistry parameters in the reef crevices in which many of the small fish live?
7. How do unusual weather conditions affect reef organisms?
8. What ecological processes allow more species to co-exist in some parts of the Red Sea than in other parts?
9. Are important resources (such as food and habitat) more finely partitioned?
10. Do physical and/or biological disturbances (i.e. predation) keep species populations reduced in size?
11. To what distance from their hatching places into the open sea do the larvae of near-shore fish travel?
12. The annual changes in biomass and composition of the ichthyoplankton.
13. How is the diversity of the reef fish community is established and maintained?
14. To what extent are there adequate qualified manpower resources to design and implement the wide variety and number of science projects in the region?

To answer the above mentioned questions there is great need to identify the high priority research tools which can be expected to be needed for projects in biological and physical oceanography in the Red Sea. These include:

1. Well-equipped laboratories close to coral reefs; they should be designed for long-term continuous operation over a time frame of decades rather than years.
2. To build a cohesive data base to make information readily available to all workers and managers.
3. Reef conservation programme and a strategy for reef management.

As our ability to intervene in physical and chemical processes that renew certain resources can affect natural systems at very large scales and at relatively rapid rates,

* We should try to understand the processes linking the physics with biology.
* A determined effort should be made to produce a clear regional inventory of existing information on the Red Sea environment, including both fauna and flora with emphasis on fisheries as an important item of the living resources.

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